CZECH UNIVERSITY OF LIFE SCIENCES PRAGUE
Faculty of Agrobiology, Food and Natural Resources
Department of Water Resources

Temperature influence on determination of the saturated soil hydraulic conductivity

Diploma Thesis

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Czech University of Life Sciences Prague

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Declaration

I declare that the Diploma Thesis on the theme "Temperature influence on determination of the saturated soil hydraulic conductivity" is my own work and all the sources I cited in it are listed in References.

In Prague, ...

Signature: ....

Acknowledgement

I would like to thank my supervisor prof. Ing. Svatopluk Matula CSc. for his support and guidance during the course of this thesis and Ing. Markéta Miháliková, Ph.D. for the great discussions, guidance, useful advices and suggestions.

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ABSTRACT

Measurement of saturated hydraulic conductivity (K_s) of soils is affected by temperature due to changing viscosity of water, as the temperature increases, the dynamic and kinematic viscosity decreases and thus the hydraulic conductivity also increases. This relationship can be easily calculated; however under natural soil conditions many other factors affected by changing surrounding temperature have to be considered as well, i.e. other soil hydraulic, mechanical and biologic properties such as contact angle in pores, solubility of salts, microbiological activity, evapotranspiration etc. The aim of this study is to investigate, if the particular daily temperature of soil and also water used for infiltration can affect the value of the saturated hydraulic conductivity in range of common operating temperatures.

Measurements consisted of manual readings from field and laboratory experiments. Field experiment was conducted at experimental site of Czech University of Life Sciences in Prague with double ring infiltrometer. Every measurement was carried out with different water temperatures used for infiltration. Soil physical properties were measured before and after the infiltration and the initial water content was tried to be kept as similar as possible for each measurement. Three sets of eight undisturbed samples were taken 10 m from the experimental site for the laboratory experiment. Samples were saturated, tempered and then processed on constant head apparatus under different temperature conditions. Three different temperatures of water were used, 10°C, 23°C and 35°C.

Results showed that there is no significant dependence of temperature either of water, either of soil on the values of saturated hydraulic conductivity while there is strong dependence of K_s on initial soil moisture content. There is a correlation between initial electrical conductivity and K_s, but weak. Generally, the values of saturated hydraulic conductivity obtained during lab measurement were in agreement with those obtained in a field, nevertheless, the lab measurements showed high variance and inconsistency due to insufficient representative elementary volume. From this preliminary study can be concluded, that for double ring infiltrometer and constant head apparatus, the values of saturated hydraulic conductivity are not significantly affected by changing the temperature in a normal operational range.

Keywords: *saturated hydraulic conductivity, soil and water temperature, double ring infiltrometer, constant head apparatus*

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INTRODUCTION

Soil properties are greatly influenced by intrinsic factors of soil formation as well as extrinsic factors associated with land use and management and vary both in time and space. Intrinsic variability is caused by the pedogenesis and usually takes place at large time scales. The variability caused by extrinsic factors could take effect relatively quickly and could not be treated as regionalized (Deb and Shukla, 2012).

The hydraulic conductivity of saturated soil is an important soil physical property for various agronomic, engineering, and environmental problems. It is also a key parameter in the design and performance assessment of irrigation and drainage systems, earthen waste impoundments, waste water leach fields, and many other agricultural, geotechnical, and environmental structures (Kohne et al., 2011). Saturated hydraulic conductivity K_s controls the infiltration of water into soil. Low values of K_s are associated with ponding of water on the soil surface, anaerobic (reducing) soil conditions, run-off, flooding and erosion (Dexter et al., 2004).

Different values of saturated hydraulic conductivity are result of some of the factors which have influence such as: soil layer texture, soil structure, seasonal and climatic change, soil salinity and acidity, microbiological activity, geomorphology, testing liquid properties, entrapment of air (Reynolds and Elrick, 2002).

This thesis reviews the calculations and laws used for measurement of saturated hydraulic conductivity and description of measurement methods of saturated hydraulic conductivity. Focus of this thesis is centered on the influence of water and soil temperature on the measured value of saturated hydraulic conductivity taken under changing weather conditions. Practical part consisted of measurements in field with double ring infiltrometer and in laboratory with constant head apparatus. Both measurements were carried out with different temperature conditions in a range of normal operational temperatures, while other conditions were kept as homogeneous as possible.

AIMS AND OBJECTIVES

The aim of the thesis is to evaluate the influence of soil and water temperature on practical determination of saturated hydraulic conductivity of the soil under the field and laboratory conditions.

Hypothesis

Temperature of the soil and water has significant influence on determination of saturated hydraulic conductivity of the soil. Some of the seasonal changes of the measured saturated hydraulic conductivity in long time measurements may be caused by the daily fluctuations of soil and water temperature.

1. REVIEW OF LITERATURE

1.1 Flow of water in soils

Water is nearly always moving in the soil like a liquid or vapor. It moves downward after rain or irrigation. It moves upward to evaporate from the soil surface. It moves towards and into plant roots, and eventually into the atmosphere through transpiration (Gardner, 1988). The water movement through a soil system influences aeration, nutrient availability to the plants, and soil temperature.

Movement of water in soils occurs under both saturated and unsaturated conditions. Saturated conditions occur below the groundwater table where water movement is predominantly horizontal with lesser components of flow in the vertical direction. We can assume that the water content by volume (θ) is equal to the total porosity and the air filled porosity is zero, in the case when the soil pores are entirely filled with water. Soil pores can be assumed to be fully saturated below the water table, while saturated soils above the water table may retain some residual entrapped air, especially near the soil surface (Warrick, 2002). For example most drainage designs are based on steady flow under saturated conditions. The tile spacing can be calculated from the known values of saturated hydraulic conductivity, soil texture, and drainage design parameters (Lal and Shukla, 2004).

Above the groundwater table (vadose zone or the zone of aeration) usually unsaturated conditions predominate. Unsaturated zone is usually considered to extend from the soil surface to the level at which pressure is zero, i.e. the groundwater table, observed as the level of water in a pipe or well when introduced to an aquifer (see Figure 1). As general rule, movement of water in the unsaturated zone is vertical, but can also have large lateral components (Freeze and Cherry, 1979; Price, 1996).

Water flow can be either steady or transient. A steady state flow conditions exists when the soil is normally fully saturated. In these flow conditions, a state of equilibrium exists for recharge and discharge. This kind of condition actually rarely exists and it is used whenever there is insufficient information to apply the more accurate procedures for transient flow conditions, which are more common. Transient flow conditions occur i.e. when long-term recharge rates exceed discharge rates and groundwater tables become elevated. These conditions are more frequent than steady-state conditions (ASCE, 1998).

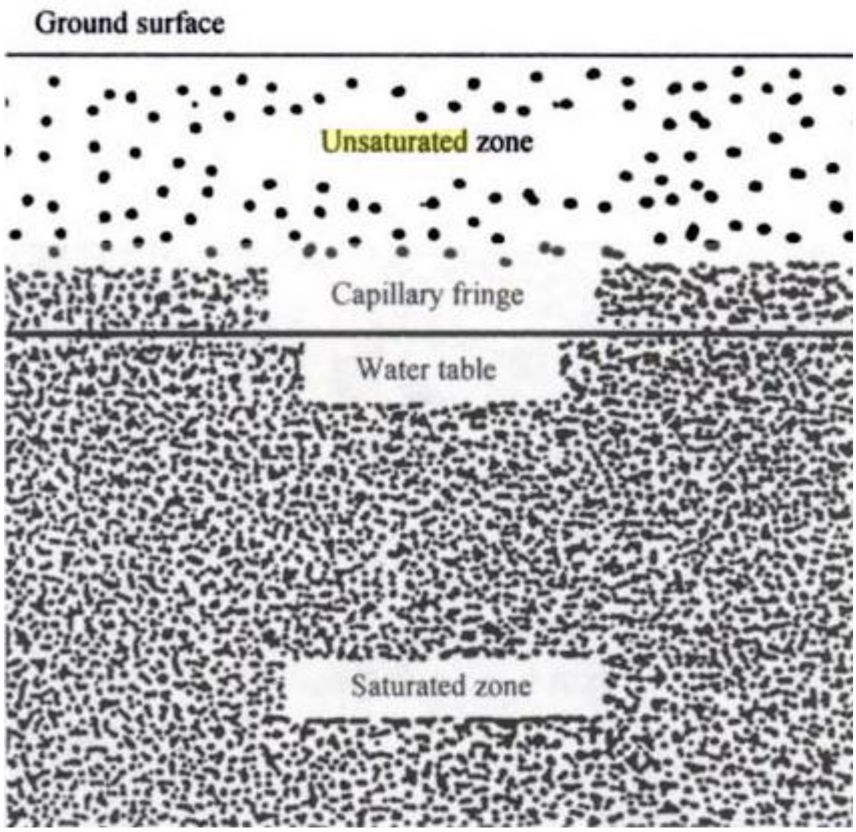


Figure 1 Saturated and unsaturated zones (Warrick, 2002)

Some of the analytic solutions ensure that the soil water flows are abundant for saturated conditions; less available for steady flow in unsaturated zone and almost completely lacking for transient, unsaturated flow conditions. Simplifications of the unsaturated flow equations have been developed to describe infiltration because of the importance of this transient flow process. It has been found that numerical methods are commonly used to solve most transient flow problems in the unsaturated zone and in saturated media where large variations in material properties make analytic solutions impossible. The hydraulic gradient, which is the driving force behind water flow, is a vector that describes the slope of the energy distribution within the soil. The parameter which is required to predict saturated flow is the saturated hydraulic conductivity (K_s). Predicting unsaturated flow requires the unsaturated hydraulic conductivity $(K(h))$ and water retention $(\theta(h))$ functions. $K_s, \theta(h)$ and $K(h)$ are all affected by soil texture and structure (among other soil properties). While soil texture is easily measured and not highly variable in space (in many cases), soil structure is difficult to quantify and highly variable in space and time. Several methods have been developed to measureK_s, and $K(h)$, each of them has some disadvantage (Warrick, 2002). Methods for determination of saturated hydraulic conductivity will be explained more in details later on.

1.2 Darcy's law

1.2.1 History

Henry G. Darcy was born in the city of Dijon in France which had a water supply among the worst in Europe in that time. Darcy was a civil engineer whose task was to improve the city's water supply. He decided to investigate the filtration of water by sands and gravels, reported in 1856 in his report Les Fontaines Publiques de la Ville de Dijon. Using simple but ingenious equipment, he arrived at his universal law for the mass flow of liquids in permeable materials. Thus, Darcy joined the „famous four“who revealed the complexity of Nature's most basic laws of flow: Fourier's law for heat flow; Ohm's law for electric current; Fick's law for gas diffusion; and Darcy's law for liquid flow in materials (Stewart and Howell, 2003).

In Darcy's time the physics of the water flow through the porous media was completely unknown. Because a detailed description of this process at the pore scale level would have been inapplicable, Darcy designed a vertical experimental tank to investigate the water flow shown in Figure 2. It was made of a tank of 3.50 m height, with a circular cross-section, is filled in the lower part on a height of 1 m with porous material. The water introduced under pressure in the upper part is evacuated at the bottom of the tank. At a certain time after starting the experiment, all the pores are filled with water and the inflow rate is equal to the outflow rate. The water pressure is measured at each end by mercury filled manometers, but Darcy expressed the head data in terms of equivalent water height. When flowing through the porous material, a loss of energy occurs and as a result a head drop across the sand filter can be put on view (Fetter, 2001).

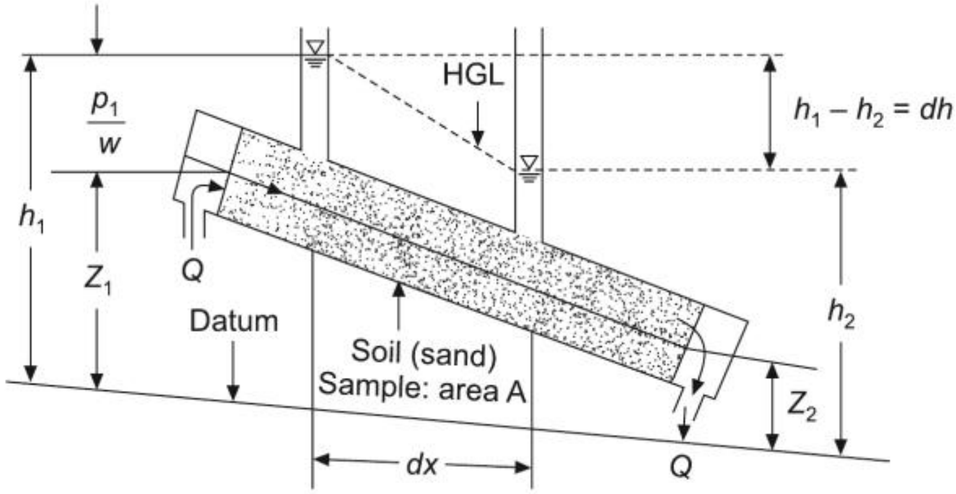


Figure 2 Darcy's experimental setup (Das and Saikia, 2013)

1.2.2 Application of Darcy's law

Darcy's law:

The discharge of ground-water through porous medium is proportional to the product of hydraulic conductivity, cross-sectional area of flow and the change in water level (head) over distance, and is inversely proportional to that distance (Kasenow, 2001).

Darcy (1856) established the law from the results of experiments with water flowing down columns of sands in an experimental arrangement shown schematically in Figure 2. Darcy found that the volume of water Q flowing per unit time was directly proportional to the crosssectional area A of the column and to the difference Δh in hydraulic head causing the flow as measured by the level of water in manometers, and inversely proportional to the length L of the column. Thus

$$Q = \frac{KA\Delta h}{L} = KA\frac{\Delta h}{L} = KAI \tag{1}$$

where the Q denotes the volumetric flow rate per unit area and has units of L/t, K is the Darcy's permeability or hydraulic conductivity, A is the cross sectional area $[cm^2]$, I is the hydraulic gradient $[m/m]$(Smith and Mullins, 2001).

Hydraulic gradient I is defined as the rate of drop of total head along the flow path.

$$I = \frac{\Delta h}{\Delta L} \tag{2}$$

Total head difference Δh $[cm]$ is the sum of differences in velocity head, pressure head and elevation head.

$$\Delta h = \left[\frac{\Delta P}{\rho g}\right] + \left[\frac{\Delta(v^2)}{2g}\right] + \Delta z \tag{3}$$

where ρ is density of fluid $[g/cm^3]$, ΔP is gauge pressure$[N/m^2]$, v is velocity $[m/s]$, Δz is loss head $[m]$.

Hydraulic conductivity expresses a combination of fluid and solid properties revealed by experiments which were conducted using variety of fluids. The flow rate is actually proportional to specific weight of the fluid, δ, g inversely proportional to the dynamic viscosity of the fluid, μ, and proportional to a property of the solid medium, k, which is called intrinsic permeability. Thus

$$K_s = \frac{k\delta g}{\mu} \tag{4}$$

where k has units of L^2 and fluid viscosity μ $[kg/m/s]$. Experiments with sand or glass beads of uniform diameter d_m by Hubbert (1956) and theoretical considerations by Hubbert (1940) further revealed that, for granular porous media, q, K and k are proportional to d_m^2.

The permeability k may vary with the time under certain conditions. It may be caused by external loads which change the structure and texture of the porous matrix by subsidence and consolidation, by the solution of the solid matrix (which over prolonged times may produce large channels and cavities), and by the swelling clay, if present within the void space. When a soil contains argillaceous material, drying of the soil may shrink the clay, especially bentonite, causing the permeability to air of the dried soil to be higher than for water. Fresh water in a soil sample may cause the clay to swell as compared with salt water, thereby reducing k. Biological activity in the medium may produce a growth which tends to clog the matrix, thus reducing k with time. Clogging may also be caused by fines carried by the water (e.g., in the artificial recharge) (Bear, 1979).

Not only Darcy's law provides a means for determining permeability, but it has a great many other practical and theoretical uses. Darcy's law can have a number of forms such as:

$$Q = KIAt \tag{5}$$

$$q = KIA = v_d A \tag{6}$$

$$v = KI = \frac{q}{A} = v_d \tag{7}$$

$$v_s = \frac{KI}{n_e} \tag{8}$$

In these expressions Q or q is the seepage quantity $[m^3/s]$, t is the time $[s]$, v_d(or v) is the discharge velocity $[m/s]$, and v_s is the seepage velocity $[m/s]$. The effective porosity n_e is the ratio of the actual volume of pore spaces through which water is seeping to the total volume (Cedergren, 1989).

In the case that there is a negative sign equation indicated that flow is in the direction of decreasing water level or hydraulic head.

When examining Equation 1. the following proportionality constants can be identified as reported in the study of Kasenow (2001):

In order to double discharge (Q), head loss (Δh) will also double; therefore, discharge is directly proportional to Δh:

$$Q \propto \Delta h$$

When distance (L) between Δh increases, Q decreases; therefore,

$$Q \propto \frac{1}{L}$$

Doubling the cross-section area of flow will also double the discharge, therefore,

$$Q \propto A$$

1.2.3 Validity of Darcy's law

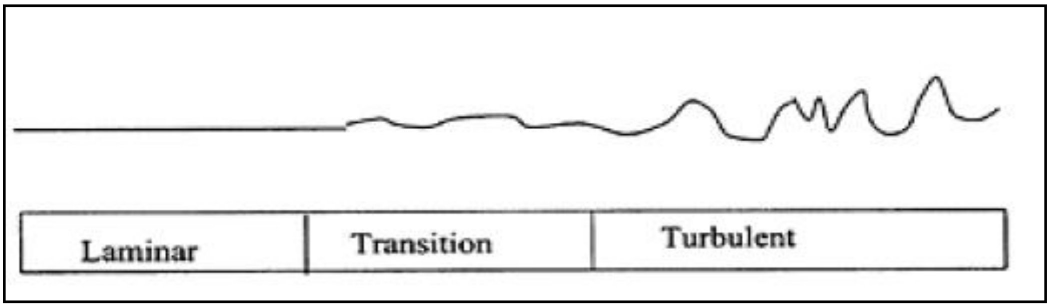
Darcy's law was established in certain circumstances: laminar flow in saturated granular media, under steady-state flow conditions, considering the fluid homogenous, isotherm and incompressible, and neglecting the kinetic energy. Still, due to its averaging character based on the representative continuum and the small influence of other factors, the macroscopic law of Darcy can be used for many situations that do not correspond to these basic assumptions (Freeze and Cherry, 1979; Lal and Shukla, 2004):

- saturated flow and unsaturated flow;
- steady-state flow and transient flow;
- flow in granular media and in fractured rocks;
- flow in aquifers and flow in aquitards;
- flow in homogeneous systems and flow in heterogeneous systems;
- flow in isotropic media and flow in anisotropic media.

Darcy's law states that the discharge velocity is proportional to the first power of the hydraulic gradient (Eq. 1). Since the velocity in laminar flow is also proportional to the first power of the hydraulic gradient, it can be inferred that the flow in porous medium must be laminar for the Darcy's law to be valid. In hydrodynamics, the usual criterion to determine whether the flow is laminar or turbulent is the Reynolds number (see Figure 3). For homogeneous aquifer materials, Darcy's law is valid only under the conditions of the inequality for the Reynolds number:

$$R_e = N_R = \frac{\delta q d}{\mu} = \frac{qd}{v} \leq N_L \tag{9}$$

where ρ is the density of water $[g/cm^3]$, μ the dynamic viscosity $[kg/m/s]$, q the discharge velocity $[m^3/s]$, d the average diameter of soil particles $[m]$, v the kinematic viscosity of water $[m^2/s]$, and N_L a number varying between 3 and 10 (Batu, 2006).



$N_R = 0$ $\qquad$ $N_R = 2000$ $\qquad$ $N_R = 4000$

Figure 3 Flow regimes in pipe flow (Lal and Shukla, 2004)

According to Bear (1972), Darcy' law which supposes a laminar flow is valid for Reynolds number less than 1, but the upper limit can be extended up to 10.

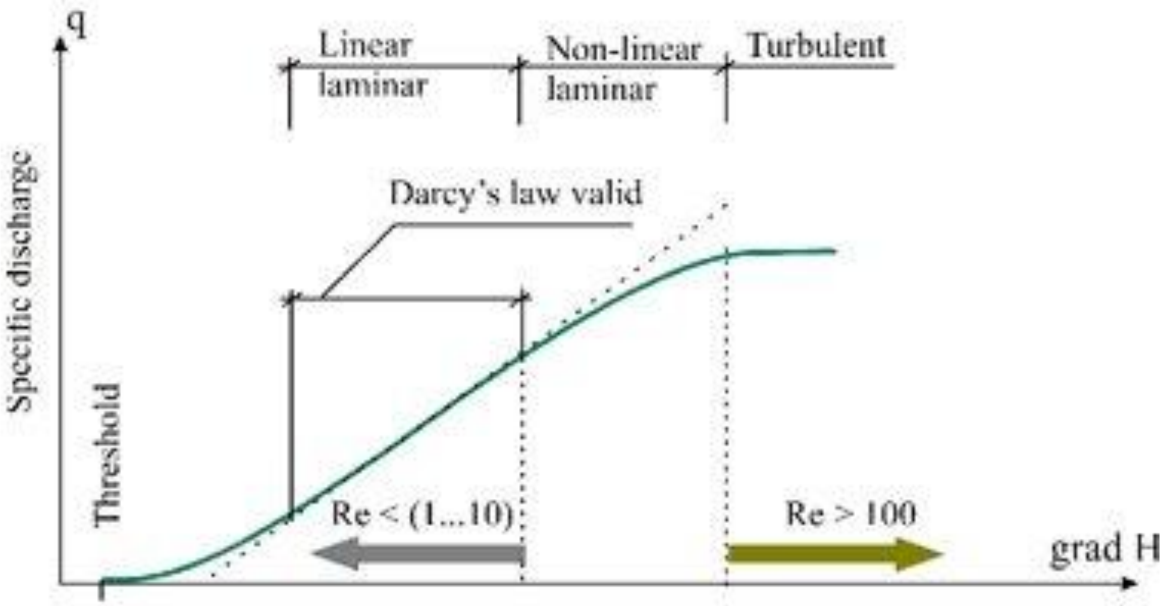


Figure 4 Range of validity of Darcy's law (VICAIRE, online, accessed 2014)

The specific discharge is always small enough in practice for Darcy's law to be applicable. Only cases of flow through coarse materials, such as gravel, deviate from Darcy's law. Darcy's law is not valid for flow through extremely fine-grained soil, such as colloidal clays (Liu and Liptak, 2000).

Experimental data from different sources have helped to involve a general consensus that there is an upper and lower limit beyond which Darcy's linear law does not hold. Basak (1978) has combined the work of all the researchers and arrived at five zones as shown in the Figure 4.

1. No flow zone- The groundwater flow is possible only after a certain hydraulic gradient that is greater than a treshold gradient. In other words, for the groundwater motion to start, the hydraulic head difference between two points must be great enough to counteract the surface forces. The finer the medium the greater the value of this threshold value.
2. Prelinear non- Darcian laminar zone- The surface forces arising from the solid-fluid interaction due to strong negative changes in the clay particle surfaces and dipolar nature of water molecules cause a nonlinear and thus non-Darcian flow in the turbulent flow domain.
3. Darcian laminar flow- Almost all of the natural porous and finely fractured media exhibit this zone to a certain extent. The inertial forces are comparatively negligible against the viscous forces and the Darcy's law is applicable confidently in this case.
4. Postlinear non-Darcian laminar zone- This zone is the transitional range from the laminar to turbulent flow during which due to gradual increase in the inertial force makes the flow deviate from linear flow.
5. Turbulent zone- Herein the turbulent flow starts and the substantial part of energy becomes dissipated in overcoming the inertial forces and comparatively to the other zones, the slope of curve is smaller (Sen, 1995).

1.3 Hydraulic conductivity

Hydraulic conductivity (K) is defined as the rate at which a geologic material can transmit a liquid under a hydraulic gradient. The liquid of concern is water. Even though it has the same units as velocity (L/T) hydraulic conductivity cannot be expressed as velocity. Permeability and coefficient of permeability are terms also used to express hydraulic conductivity. In practice, it is often used in conjunction with a hydraulic gradient; whereas, permeability is often used in the absence of a gradient (Kasenow, 2002).

The hydraulic conductivity is defined in two separate parts considering the saturated or unsaturated status of the media (Karamouz et al., 2011).

In saturated hydraulic conductivity, only the solid (soil particles) and liquid (water) states of matter exist. All the pore spaces are completely filled with water and the K is constant. However, in unsaturated flow, the K is not constant; it decreases as the water content decreases because the pore spaces are not completely filled, and there is the existence of air in some pore spaces (Edoga, 2010).

1.3.1 Hydraulic conductivity in Saturated Media

Hydraulic conductivity$K$ contains the properties for both medium and fluid. It can be used for evaluating water transmissivity in a porous medium. Hydraulic conductivity of a saturated media has already been defined in Eq. 4.

The physical meaning of hydraulic conductivity is stated as „the volume of liquid flowing perpendicular to a unit area of porous medium per unit time under the influence of a hydraulic gradient of unity.“ (Karamouz et al., 2011)

The saturated hydraulic conductivity is greatly influenced by effects such as macropores, stones, fissures, cracks, and other irregularities formed for various biological and mechanical reasons. Hence, it is the parameter that may be rather difficult to predict (Haverkamp et al., 1999). Because of spatial variability of saturated hydraulic conductivity it is difficult to find accurate representative values to correctly predict soil water flow and design irrigation and drainage systems (Moustafa, 2000).

In the practice different units are used for hydraulic conductivity K (dimensions L/T). Hydrologists prefer the unit m/day (meters per day). Soil scientists often use cm/s. In the USA, as in many countries using the English system of units, two other units are commonly employed by the hydrologists. One is a laboratory, or standard, hydraulic conductivity defined as the total discharge (Q) of water at 60°F, expressed in gallons per day, through a porous medium cross-sectional area (A) expressed in ft^2 under a hydraulic gradient $\{(\phi_1 - \phi_2)/L\}$ or $1ft/ft$. With this definition, the units of K are $gal/dayft^2$. In a similar way, a field, or aquifer, hydraulic conductivity is defined as the discharge of water at field temperature, through across-sectional area of an aquifer one foot thick and one mile wide under a hydraulic gradient of $1ft/mile$. The unit is the same as for the laboratory K following some conversions among these units.

$$1\,US\frac{gal}{dayft^2} = \frac{4.72x10^{-5}cm}{s} = 4.08x10^{-2}m/d$$

Permeability k (dims.L^2) is measured in the metric system in cm^2 or in m^2. In the English system, the unit is ft^2. According to Bear (1979) for water at 20°C, we have conversion

$$\frac{1cm}{s} = 1.02x10^{-5}cm^2$$

Reservoir engineers use the unit Darcy defined by

$$1\,darcy = \frac{\frac{1cm^3}{s}/cm^2x1centipoise}{1atmosphere/cm}$$

with

$$1\,darcy = 9.8697x10^{-7}cm^2 = 1.062x10^{-11}ft^2$$

$= 9.613x10^{-4}cm/s$ (for water at 20°C)

$= 1.4156x10^{-2}USgal/minft^2$ (for water at 20°C)

In Table 1. Some typical values of the hydraulic conductivity are presented (Spitz and Moreno, 1996):

Table 1 Ranges of Hydraulic conductivities (Spitz and Moreno, 1996)

Unconsolidated deposits	**Hydraulic conductivity (m/s)**	**Rocks**	**Hydraulic conductivity (m/s)**
Dense clay	10^{-13}-10^{-8}	Dense sandstone	10^{-9}-10^{-7}
Weathered clay	10^{-8}-10^{-6}	Karstic sandstone	10^{-7}-10^{-5}
Silt	10^{-7}-10^{-5}	Dense limestone	10^{-9}-10^{-7}
Alluvial deposits	10^{-5}-10^{-3}	Karstic limestone	10^{-5}-10^{-3}
Fine sand	10^{-5}-10^{-4}	Dolomite	10^{-10}-10^{-8}
Medium sand	$5x10^{-4}$-$5x10^{-3}$	Dense crystalline rocks	10^{-13}-10^{-12}
Coarse sand	10^{-4}-10^{-3}	Fractured crystalline rocks	10^{-10}-10^{-6}
Fine gravel	10^{-3}-$5x10^{-1}$	Dense basalt	10^{-13}-10^{-10}
Medium gravel	$5x10^{-2}$-10^{-1}	Fractured basalt	10^{-7}-10^{-4}
Coarse gravel	10^{-2}-$5x10^{-1}$	Claystone	10^{-13}-10^{-9}

1.4 Methods for measuring saturated hydraulic conductivity

Saturated hydraulic conductivity of soils can be measured in the laboratory and in the field. Laboratory measurements are inexpensive, quick and easy to make compared to field measurements. They are often used to obtain an initial characterization of a site before on-site characterization is initiated (Artiola et al., 2004). A schematic overview of different methods used for measurement of saturated hydraulic conductivity is presented on Figure 5.

Saturated hydraulic conductivity can be determined directly by measuring water movement through a soil profile or sample. However, there exist also methods of indirect estimation from associated soil properties such as soil texture, bulk density, organic matter content and others. These methods are in general called pedotransfer functions (Bouma, 1989). It has to be considered that estimation is not interchangeable with direct measurement; however, in large regional studies the estimation is often a useful, economic and sufficiently accurate alternative to the direct measurement (i.e. Wösten et al., 2013).

1.4.1 Laboratory measurements

1.4.1.1 General principles

Laboratory measurements are performed on either disturbed or undisturbed samples that are collected in the field. Obtaining an undisturbed core sample is usually possible for consolidated materials, such as structured soils rocks, but is difficult for unconsolidated sediments, such as sand and gravel. Measurements made for a core sample represent that specific volume of media. That sample is then assumed representative of the field site. However, a single sample will rarely provide an accurate representation of the field because of the heterogeneity inherent to the subsurface. Thus a large number of samples could be required to characterize the saturated hydraulic conductivity distribution present at the site (Artiola et al., 2004).

The core sample method determines saturated hydraulic conductivity on samples by either a constant head or falling head test. Core samples are taken in the field and kept from drying. In the laboratory, the samples are saturated from the bottom to prevent air entrapment.

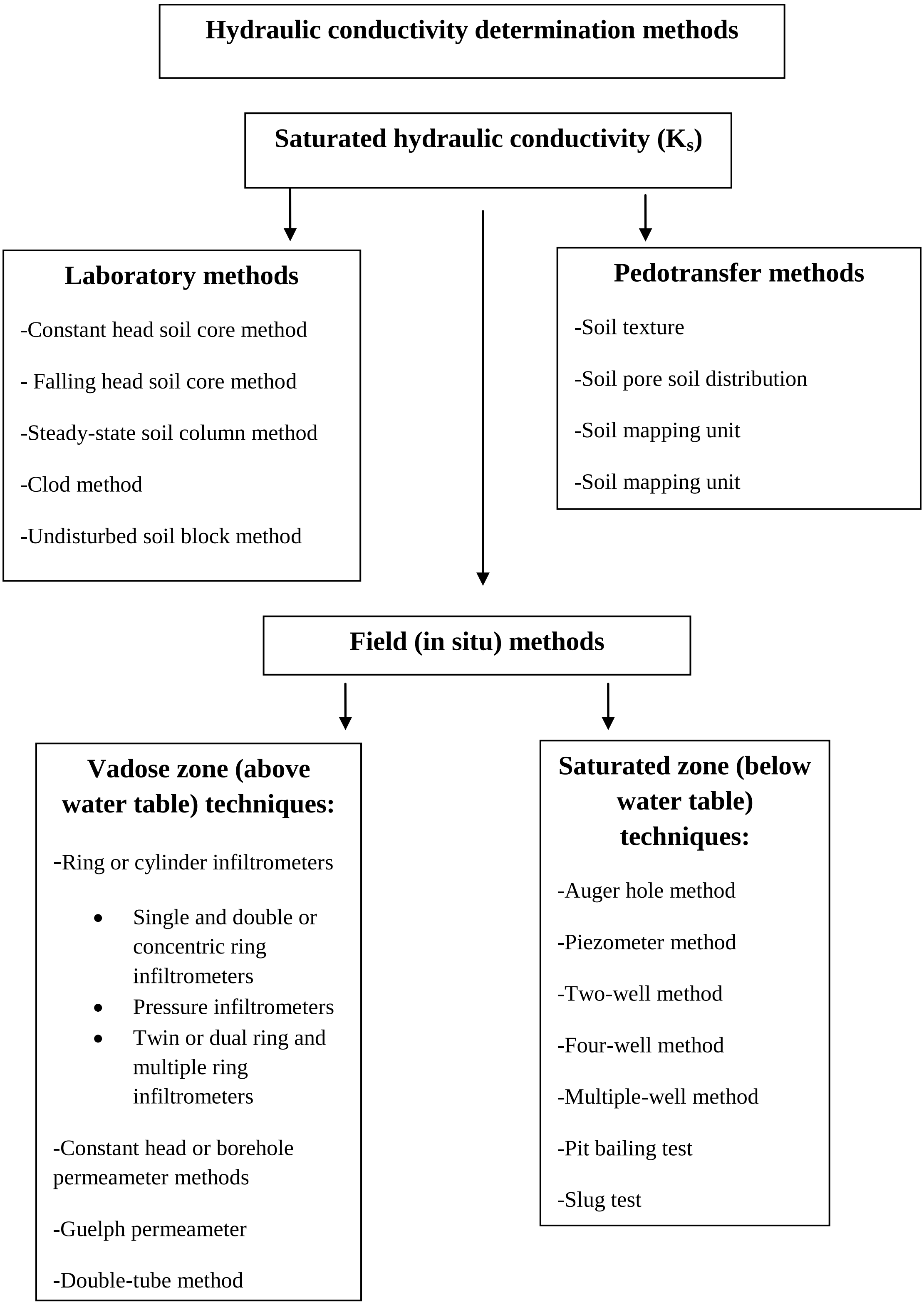


Figure 5 Overview of methods for the determination of hydraulic conductivity of soil (adjusted according to Deb and Shukla, 2012)

In general, values of saturated hydraulic conductivities are influenced by the entrapped air inside the soil column. There are several options to remove entrapped air before the measurement. Some of these measures include flushing the core with carbon dioxide gas, wetting the cores very slowly from the bottom, and conducting experiments under conditions of upward flux (Shukla, 2014; Ward et al., 2004)

In both tests, water moves through a test sample under the influence of gravity alone; in both tests the sample is placed in a tube or cylinder and is usually remolded in the process of placing it into the cylinder. If the undisturbed sample is placed into a permeameter we have to assure no leakage along boundary between the sample and the cylinder. This is difficult to accomplish with certainty in practice. Rubber membranes, silicon or wax have been used to seal the sides of the samples (Nielsen, 1991).

1.4.1.2 Constant head method

The constant head permeability test (see Figure 6) is used for coarse- grained soil only where a reasonable discharge can be collected in a given time (Punmia et al., 2005) and it's more suitable for very permeable soils (Terzaghi et al., 1996).

By the same arrangement as Darcy used in 1856 the soil column is supported on a permeable base such as wire gauze or filter, or sometimes a sand table. Water flows through the column from a constant head of water on the soil surface and is collected for measurement from an outlet chamber attached to the base. When the liquid enters the soil, reduction in pressure appears which gives a small error which is considered to be of no great importance (Smith and Mullins, 2001).

In the constant head method, as the name suggests water flow is measured across a saturated porous media where flow is taking place due to the positive potential head maintained at the inlet. The outflows of water are recorded at specific time intervals.

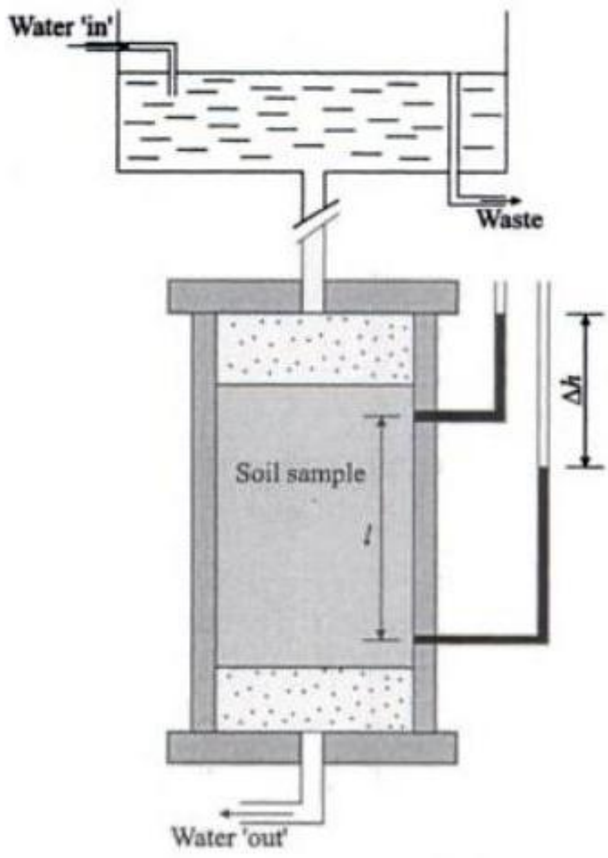


Figure 6 Schematic diagram of constant-head permeability test setup (Potts and Zdravković, 2001)

When several readings of outflow for a given time interval are similar, with small derivations between readings, the saturated hydraulic conductivity of soil for a vertical downward flow is calculated using the rearranged Darcy's law as follows:

$$K_s = \frac{QL}{A(H_1 - H_2)} = +\frac{VL}{At(L+h)} \tag{10}$$

where V is the total volume of water $[mL]$ that came out of the bottom of the saturated soil core in time $t\left(Q = \frac{V}{t}\right)[s]$, h is the constant head of water maintained on the soil surface $[m]$, L is the lenght of the soil column $[m]$.

If the flow is taking place in the horizontal direction, the saturated hydraulic conductivity of the soil is calculated as follows:

$$K_s = \frac{QL}{A(H_1 - H_2)} = \frac{VL}{tAh} \tag{11}$$

where h is the difference of positive potential heads between inlet and outlet $[m]$. The constant head method can be used to determine the hydraulic conductivity of different types of soils. However, for highly permeable soil, maintaining a constant head could be difficult (Shukla, 2014). In the Table 2 and 3 are shown values of saturated hydraulic conductivity.

Table 2 Saturated Hydraulic conductivity ($\boldsymbol{K_s}$) of different soils (Smith and Browning, 1946)

Class	$\boldsymbol{K_s(cm/day)}$	**Comments**
Extremely slow	<0.1	Very nearly impervious
Very slow	0.1-1	Poor drainage, staining, too slow for artificial drainage
Slow	1-10	Poor aeration, poor root development
Moderate rapid	1-1000	Good to excellent drainage, water holding
Very rapid	>1000	Poor to holding, excessive drainage

Table 3 Saturated Hydraulic conductivity (K_s) of different soils (Bear, 1972 and Fetter, 2001)

Soil classification	$\boldsymbol{K_s(m/s)}$**; Bear, 1972**	$\boldsymbol{K_s(m/s)}$**; Fetter, 2001**
Coarse sand	10^{-2}-10^{-5}	10^{-3}-10^{-5}
Sand	10^{-2}-10^{-5}	10^{-3}-10^{-5}
Loamy sand	10^{-5}-10^{-9}	10^{-5}-10^{-7}
Sandy loam	10^{-5}-10^{-9}	10^{-6}-10^{-8}
Loam	10^{-4}-10^{-9}	10^{-6}-10^{-8}
Sandy clay loam	10^{-5}-10^{-9}	10^{-6}-10^{-8}
Clay	10^{-6}-10^{-13}	10^{-8}-10^{-11}

1.4.1.3 Falling head method

Falling head test (see Figure 7) is used for relatively less permeable soils when the discharge is small (Punmia et al, 2005). In the falling head method, there is usually standpipe attached at the top of the soil core. The head at the top of the core is allowed to drop one value to the other during a given time interval (Shukla, 2014).

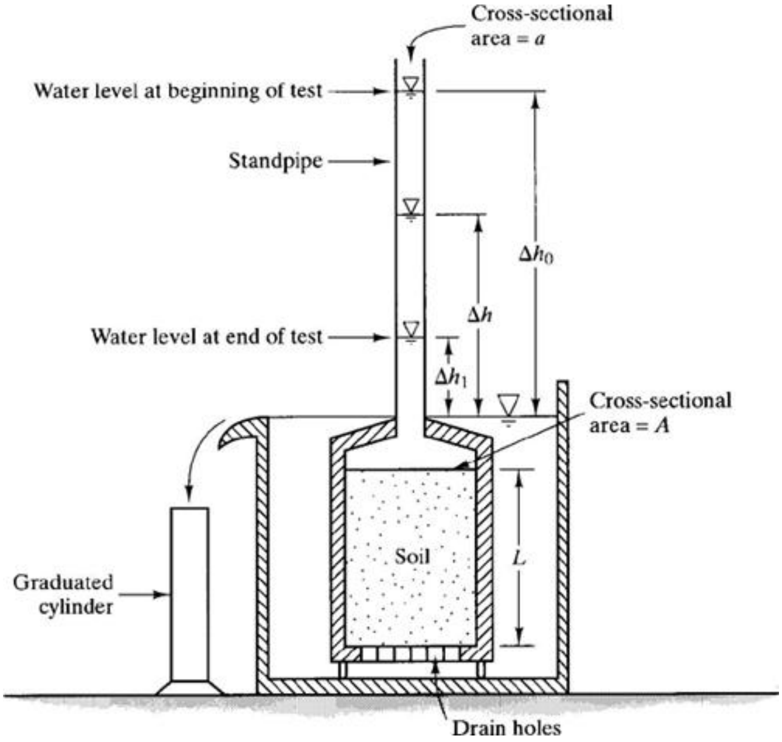


Figure 7 Schematic diagram of falling-head permeability test setup (Coduto, 1999)

The principle of the falling head test can be given by the following mathematical relationship:

$$\frac{\partial V}{\partial t} = \frac{K_S \Delta H}{L} \tag{12}$$

where V is the volume of water $[mL]$ displaced in time $t[s]$; Δ is the change in the magnituded of a quantity, ΔH is the total head difference $[m]$ and L is the length of soil sample $[m]$. Integrating the above equation between limits t_1, H_1 to t_2 $, H_2$ we can calculate saturated hydraulic conductivity by the formula:

$$K_s = \frac{aL}{At} log_e \left(\frac{H_1}{H_2}\right) \tag{13}$$

where A iscross-sectional area of the sample $[m^2]$ and log_e is the natural logarithm (Lal and Shukla, 2004).

The falling head method is useful for soils with very low hydraulic conductivity, and collection of a sufficient amount of outflow takes several hours so that evaporation might influence the accuracy of measurement.

1.4.2 Field (in situ) measurements

In-situ measurement of saturated hydraulic conductivity is preferred, if applicable, because the measurement is taken under natural conditions, when the soil has hydraulic connections to the surrounding and the volume of soil involved in measurement is more representative than a small core sample.

Methods for field measurement are generally divided to two main groups:

a) groundwater table is available
b) groundwater table is not available

in the desired measured soil layer. Thus the measurements are taken in saturated or vadose zone.

1.4.2.1 Saturated zone (below groundwater table)

Several widely used methods are described below. According to the depth of saturated zone and the scale of area which is involved, the methods correspond with hydropedology or hydrogeology.

For example in hydrogeology, field techniques generally measure the horizontal hydraulic conductivity and can involve permanent installations such as predrilled or push-in piezometers for conducting slug tests. Predrilled piezometers are time consuming and costly to install and involve other problems including disposal of drill cuttings, positioning and alignment of the screen and proper construction of sand pack and isolation seal. Use of push-in piezometers eliminates many of these problems and are much quicker to install but have the disadvantages of disturbing the soil during installation. Some of the other techniques like piezocome, the flat dilatometer are relatively quick to perform and allow detailed profiling of hydraulic conductivity in a short period of time (Daniel and Trautwein, 1994).

1.4.2.1.1 Auger hole method

In hydropedology the auger-hole method is rapid, simple and reliable method for measuring hydraulic conductivity of soil below a groundwater table. It is very often used i.e.

in connection with the design of drainage systems in waterlogged land and in canal seepage investigations (Beers, 1983).

The general principle of the method is as follows:

A hole is bored into the soil with an auger at least 30 cm below the groundwater table with taking care to minimize disturbance of the sidewalls. When the water in the hole reaches equilibrium with the ground water, part of it is removed. The groundwater then begins to seep into the hole and the rate at which it rises is measured and then converted by a suitable formula to the hydraulic conductivity for the soil (Mahajan, 2009).

In moderately permeable soils the rise in water level can be measured with a tape and float (see Figure 8); in highly permeable soils a pressure transducers should be used (EPA, 1993).

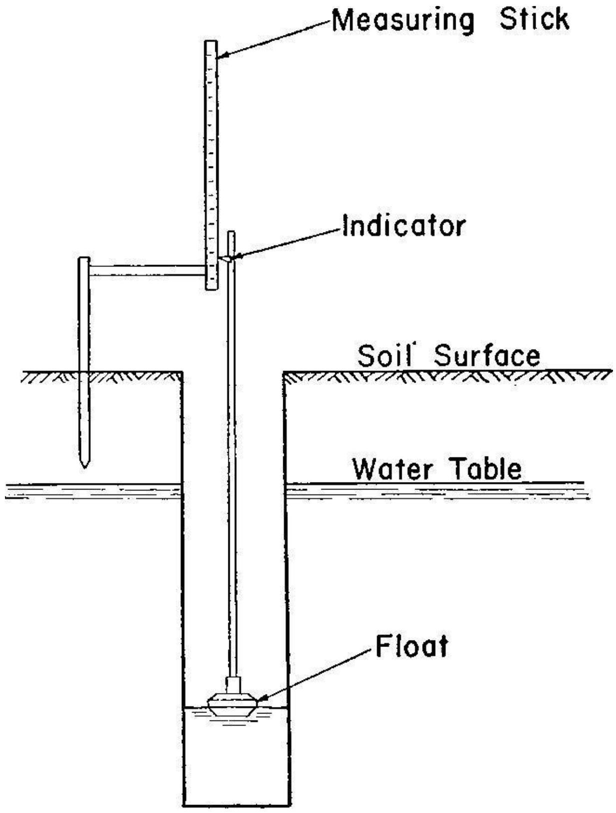


Figure 8 Schematic diagram of auger hole method by (Reeve et al., 1957)

In measuring hydraulic conductivity in the field, four phases can be distinguished, each having its own problems (Beers, 1983):

- The drilling of the holes.
- The removal of the water from the hole.

- The measurement of the rate of rise.
- The computation of the hydraulic conductivity from the measurement data.

There exist several possibilities how to compute the saturated hydraulic conductivity within the auger hole methods, some of them are presented i.e. in Báťková et al. (2013).

According to Ernst and Wasterhof (1950), the saturated hydraulic conductivity is calculated using:

$$K_s = C\frac{dy}{dt} \tag{14}$$

In which:

$$C = \frac{4000\frac{r}{\bar{y}}}{\left(\frac{H}{r}+20\right)\left(2-\frac{\bar{y}}{H}\right)} \tag{15}$$

Where the bottom of the hole is far above the impermeable base $(H > H/2)$, or:

$$C = \frac{3600\frac{r}{\bar{y}}}{\left(\frac{H}{r}+10\right)\left(2-\frac{\bar{y}}{H}\right)} \tag{16}$$

Where the bottom of the hole reaches the impermeable base $(D = 0)$. In these formulae:

$C =$ constant, depending on hole geometry;

$\frac{dy}{dt} =$ rate of rise in water level $[cm/s]$;

$D =$ depth of impermeable layer below bottom $[cm]$;

$h = H - y =$ height of water column $[cm]$;

$h_1, h_2 =$ initial and final water column in hole $[cm]$;

$H =$ depth of borehole below groundwater $[cm]$;

K_s = average saturated hydraulic conductivity$[m/d]$;

r = radius of the borehole$[cm]$;

t = time $[s]$;

y = depth of water below groundwater table $[cm]$

$\bar{y}$ = average value of y in the interval where $y > \frac{3}{4y_0\,(cm)}$;

$\frac{dy}{dt} = \frac{y_1 - y_2}{t_2 - t_1} \quad y_1 > y_2; t_2 > t_1$

Where the impermeable base is close to the bottom of the hole, and interpolation between equations 16 and 17 is used.

The results within the same auger hole are usually quite consistent, but between different holes, even nearby ones, differences may be considerable owing to local soil variations (FAO, 2010).

1.4.2.1.2 Piezometer method

A piezometer (see Figure 9) is an open- ended pipe driven into the soil that measures the ground water pressure below the groundwater table (Smith and Mullins, 2001). The piezometer method is based on the measurement of the flow into an unlined cavity at lower end of a lines hole. Water entering the unlined cavity and rising in the lined hole is removed several times by pumping or bailig to flush the soil pores along the cavity wall. After flushing is completed, the water is allowed to come to equilibrium with the groundwater table.

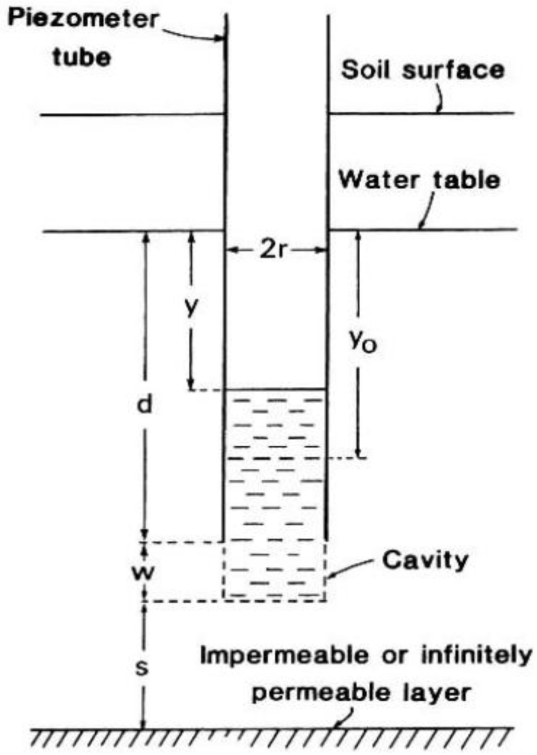


Figure 9 Schematic diagram of piezometer method (Smith and Mullins, 2001).

There are several varieties of piezometer tests used, depending on the geometry and materials located at the point of measurement. Besides the cylindrical cavity outlined in this procedure, there are spherical cavities, sand-filled cavities, piezometer tips placed in sand-filled cavities, or piezometer tips pushed directly into the soil (Shults, 1981).

The hydraulic conductivity can be calculated as reported in Smith and Mullins (2001):

$$K = \frac{\pi r^2 ln(y_0/y)}{A(t - t_0)} \tag{17}$$

where y_0 and y are the depths of the water level $[m]$ in the well below the equilibrium level at time t_0 and at time t $[s]$, A is a shape factor that depends on the depth d $[m]$of water in the well at equilibrium, the length w of the cavity at the bottom of the well$[m]$, and the depth s $[s]$of the soil to a stratum of different hydraulic conductivity, all expressed as a fraction of the radius r of the well; which is, $A = A(d/r, w/r, s/r)$.

This method is mostly accurate in layered soils; as long as measurements are made in different layers with the cavity properly located at least one radius above the change in the soil. With cavities of small length, the flow is mainly vertical, so that values reflect the vertical component of hydraulic conductivity in anisotropic soils.

1.4.2.2 Vadose zone (above groundwater table) methods

When using field techniques for determining saturated hydraulic conductivity above groundwater table, the hydraulic conductivity calculations are based on infiltration into primarily unsaturated soil. Even under the best of circumstances, some air is entrapped during infiltration, and the soil does not become fully saturated. These methods give a value for the field-saturated hydraulic conductivity that may be significantly different between methods, sites, and initial water conditions (Ward et al., 2004).

1.4.2.2.1 Double-Ring infiltrometer (Flooding-type infiltrometer)

Double-ring infiltrometer (Parr and Bertrand, 1960) (see Figure 10) is the most commonly used and is designed to overcome the basic objection of the tube infiltrometer (Subramanya, 2008).

It is used to determine the rate of infiltration of water into the soil. The principles of the double-ring infiltrometer and method of operation are similar to the single-ring infiltrometer.

Two sets of concentrating rings with diameters of about 30 cm and 60 cm and of a minimum length of 25 cm are used. The rings are inserted into the ground and water is applied into both of the rings to maintain constant depth of about 5 cm. The outer ring provides water jacket to the infiltering water from the inner ring and hence prevents the spreading out of the infiltering water of the outer ring. The outer ring helps the horizontal movement of water during infiltration from the inner ring. The water depths in the inner and outer rings are kept the same during the observation period. The measurement of the water volume is done on the inner ring only. The experiment is carried out till a constant infiltration rate is obtained. A perforated disc is provided on the soil surface in the inner ring to prevent formation of turbidity and settling of fines on the soil surface (reported i.e. in Subramanya, 2008).

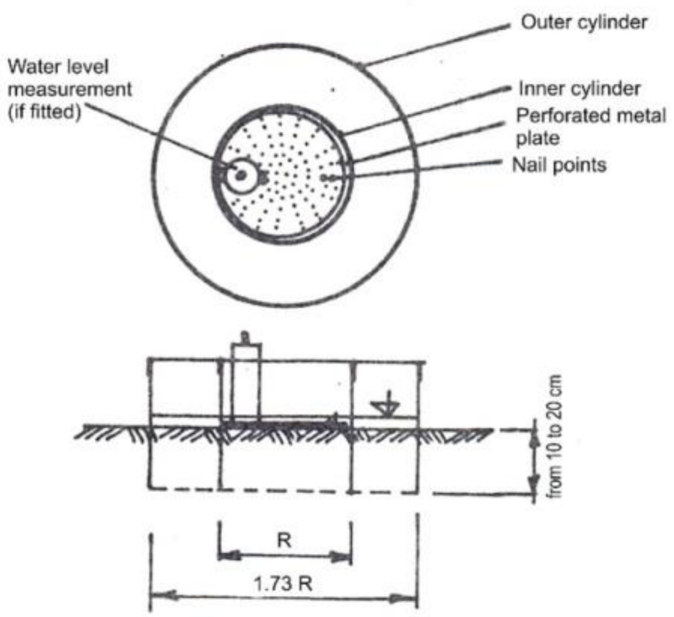


Figure 10 Schematic diagram of double ring infiltrometer (Báťková et al., 2013)

An automation of water level reading can be applied, for example in Matula and Dirksen (1989).

Philip (1957) developed a set of equations which are called Philip's infiltration equations (Eq. 19 and 20). Cumulative infiltration $\boldsymbol{I}$ under water-ponded conditions is approximated at time $\boldsymbol{t}$ by:

$$I = St^{0.5} + At \tag{18}$$

And the rate of infiltration:

$$v = \frac{1}{2}St^{-0.5} + A \qquad \text{(Ошибка! Закладка не определена.)}$$

where I is cumulative infiltration $[L]$, S is sorptivity $[L/T^{0.5}]$, A is a constant $[L/T]$, v is infiltration rate for any given time $[m/s]$.

Saturated hydraulic conductivity can be calculated as:

$$K_s = \frac{A}{m} \tag{19}$$

where m is a constant equal to 2/3 (or 0.66667).

Double-ring infiltrometers may be either open to the atmosphere, or the inner ring is covered to reduce evaporation. The flow rate is measured for open double- ring infiltrometers directly from the rate of decline of water level within the inner ring as a falling head test. For the constant- head method the rate of water input necessary to maintain a stable head within the inner ring is used for calculation of flow rates. The sealed double- ring infiltrometers determines a flow rate by weighing a sealed flexible bag supply reservoir for the inner ring (Sara, 2003).

Double ring infiltrometer or flooding type infiltrometers measures infiltration at a spot only, a lot of other experiments are necessary to obtain representative infiltration characteristics for an entire watershed. Some of the main disadvantages of this type of infiltrometers are observed by Subramanya (2008):

1. The raindrop impact effect is not simulated;
2. The driving of the tube or rings disturbs the soil structure; and
3. The results of the infiltrometers depend to some extent on their size with the larger infiltrometers giving fewer rates than the smaller ones; this is due to the border effect.

1.4.2.2.2 Guelph permeameter method

Guelph permeameter (see Figure 11) estimates besides field saturated hydraulic conductivity (K_s), matric flux potential, and soil sorptivity based on constant head calculations. The Guelph permeameter (for example, Model 2800K1, Soil moisture Equipment Corp., http://www.soilmoisture.com) is a constant head well permeameter consisting of a mariotte reservoir that maintains a constant water level inside an augered hole. This permeameter requires steady discharge from two different water levels (heads) in the augered hole. Steady state discharges are measured at two different water pressure heads (Erickson et al., 2013)

The equipment can be transported, assembled, and operated easily by one person. Measurements can be made in 1/2 to 2 hours, depending on soil type, and require only about 2.5 liters of water. Measurements can be made in the range of 15 to 75 cm below the soil surface.

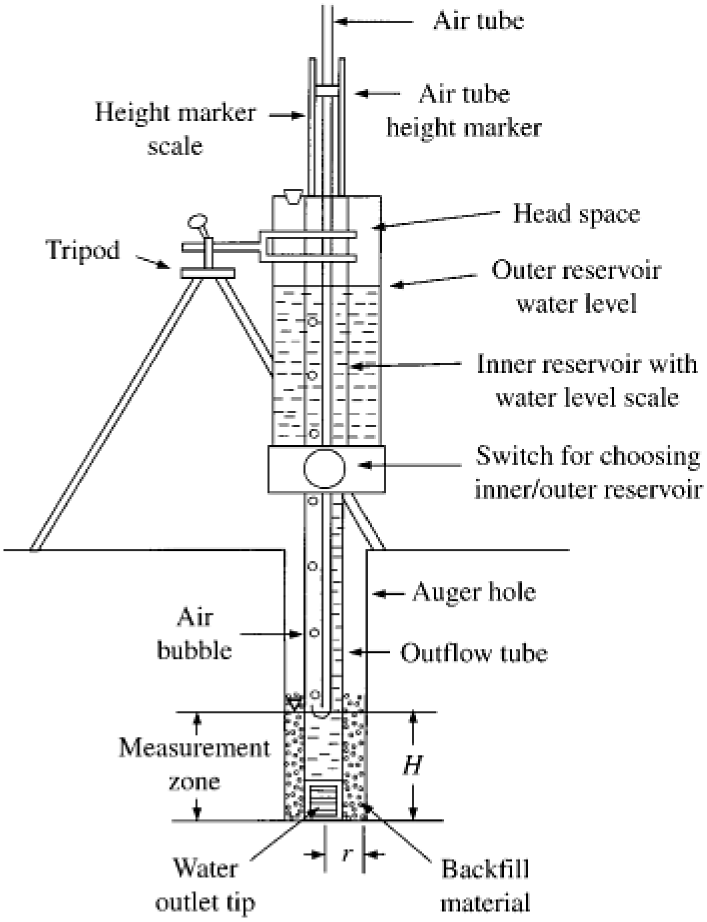


Figure 11 Schematic diagram of Guelph Permeameter (Down and Lehr, 2004)

The Guelph permeameter method measures the steady-state rate $Q[m^3/s]$ necessary to maintain a constant depth of water $H[m]$ in an uncased cylindrical well of radius $a[m]$, above the water table. Then the field saturated hydraulic conductivity K_{fs} and the matrix flux $f_m(m^2/s)$ are calculated from Q and H, and using the following approximate analytical solution:

$$\begin{aligned} Q &= \left(\frac{2\pi H^2}{C} + \pi a^2\right) K_{fs} + \frac{2\pi H}{C}\phi_m \\ &= Ak_{fs} + B_{\phi_m} \end{aligned} \tag{20}$$

where C is dimensionless shape factor primarily dependent on the H/a ratio and soil type.

The values of K_{fs} can be measured with the Guelph permeameter from 10^{-4} to 10^{-8} m/s. Beyond these limits there is a reduction in accuracy and precision. In soils with $K_{fs} < 10^{-8} m/s$ the rate of infiltration is too low to be monitored accurately.

The Guelph Permeameter can be used anywhere a hole can be augered in the soil. Soils typically possess a three dimensional heterogeneity, while the Guelph permeameter method essentially provides a "point" measurement. Therefore, the number of measurements to adequately represent field variability will depend on factors such as soil type, type of application, project objectives, etc... A description of the soil profile (by sampling or from soil survey reports) will greatly complement the value and understanding of data obtained with the Guelph Permeameter (Fortin, 2003).

1.4.2.2.3 Pressure infiltrometer method

This method provides in situ determinations in the unsaturated (vadose) zone or field saturated hydraulic conductivity (K_{fs}) and matric flux potential(ϕ_m). It involves the measurement of the steady-state infiltration rate (recharge) required to maintain a steady depth of water (or constant water pressure) within a single ring inserted a small distance in the porous medium (soil).

The most important difference between the pressure infiltrometer method and the traditional single-ring infiltrometer method is the theoretical treatment of water flow out of the ring and into the unsaturated soil (Carter, 1993). This method can be also used for measuring sorptivity, macroscopic capillary, pressure head, macroscopic capillary length parameter, air entry pressure head and water entry pressure head.

By Matula and Kozakova (1997) pressure infiltrometer (see Figure 12) is a Mariotte type infiltrometer which consists from non corrosive materials (Plexiglass, PVC and Teflon). Device is portable, easy to set up, uses mechanical-hydraulic principle without need of an external energy supply. The device enables the measurement with acceptable accuracy of the cumulative infiltration of ponded water from the infiltration ring.

1- Piston valve to open or close the water outlet
2- Moveable air tube to set the applied water pressure H in infiltrating surface
3- Marriote type water reservoir
4- Plexiglass tube of a small diameter to enable accurate fading of the water level
5- Iron ring
6- Bulb of field saturated soil
7- Wetting front

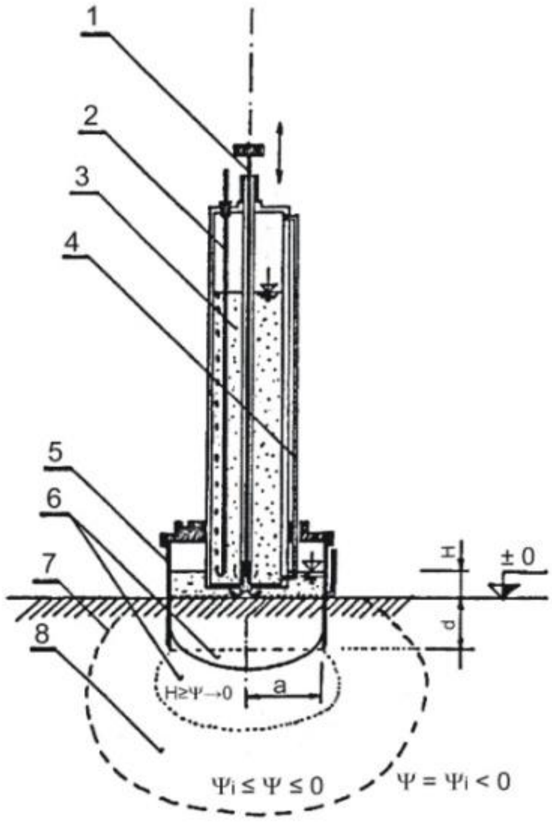


Figure 12 Schematic diagram of pressure infiltrometer (picture taken from Báťková et al., 2013)

Saturated hydraulic conductivity can be calculated by Báťková et al. 2013:

$$K_s = \frac{(Q_{ti}G_{ti})}{\left(aH + a^2 G_{ti}\pi + \frac{a}{\alpha}\right)} \tag{21}$$

where Q_{ti} is the steady infiltration (L^3/T), a is the radius of the infiltration ring (L), G_{ti} is the shape factor (L^3/T), H is the hydraulic head of ponded water in the infiltration ring (L), α is a parameter $(1/L)$. This infiltrometer was used for example in study of Špongrová et al. (2010).

2. MATERIALS AND METHODS

Saturated hydraulic conductivity is going to be measured under the field and laboratory conditions. Field measurement on the natural soil profile is generally better for determination of saturated hydraulic conductivity; however, controlled laboratory conditions with artificially prepared soil samples reduce the natural heterogeneity and allow quantifying the results. Permanently installed double ring infiltrometer (diameter 35.7 cm) will be used for field measurement. Several repeated measurements will be carried out under different water temperatures, approximately 15°C, 25°C, 35°C. Initial water content should be kept as similar as possible to exclude a possible influence of this factor.

24 undisturbed soil samples will be taken from the field and used for the measurement. These core samples will be processed on a lab-built constant head apparatus, which consists of one Mariotte type cylinder bottle (height 49 cm, weight 1154 g), rubber stoppers with air entry and refilling tube, teflon valves, connectors and Plexiglass tube. Cylinder reservoir has volume of 2000 cm^3, inner diameter of 10.5 cm, and height of 44 cm.

8 replications will be always carried out under different temperature conditions, approximately 15°C, 25°C, 35°C. All the results will be statistically evaluated.

2.1 Field Experiment

The double ring infiltrometer was used to carry out the infiltration experiments in order to determine the saturated hydraulic conductivity. Six infiltration experiments under different temperatures were carried out in total. The double ring infiltrometer was installed and left on the same place in order to maintain the same natural conditions and thus reduce the natural space variability of saturated hydraulic conductivity. After each particular infiltration experiment, the soil with installed rings was left for several days in order to let the water redistribute within the soil profile.

2.1.1 Study area and site description

The study was conducted in July and October 2013 at the Experimental Terrain Station of Soil Moisture Dynamics of the Department of Water Resources (see Figure 13, the station is in central part). The measurement area was located in the capital of Czech Republic - Prague in the western part of the city called Suchdol in the campus of Czech University of Life Sciences Prague. Elevation is approximately 279 m, longitude 14°22'25.26"E, latitude 50°04'40.42"N. The average annual air temperature is around 9°C, average annual precipitation of around 500 mm, the appropriate time zone CET (GMT + 1 hour).

Figure 13 General description of the map area: 1. Experimental setup for field measurement, 2. Area used for taking undisturbed soil samples used for laboratory experiment (source: Google Earth)

The soil at the site (see Figure 14) is Udic Haplustoll (according to Soil Taxonomy) or Haplic Chernozem (according to World Reference Base) of loamy texture on an aeolic loessial substrate. A groundwater table is not available. The fine earth contains 22-32.5% sand, 39.5-54% silt and 22-28% clay. There is virtually no textural difference between topsoil and subsoil. The boundary between A and C horizon lies at about 35 cm, the transitional A/C horizon is only about 10 cm thick. The accumulation of carbonates was traceable in the C-horizon but did not lead to a significant reduction of the subsoil hydraulic conductivity.

Information about the site is taken from the HYPRESCZ database (Miháliková et al., 2013). The site used to be a regularly operated field, which lead in compaction of the soil between approximately 15 and 25 cm of the depth. In present time, there is a permanent grass culture since 2009.

Figure 14 Experimental area before installation of double ring infiltrometer

An overview of weather conditions during the field experiments is given in the Table 4 and Figure 15.

Table 4 Weather conditions during experiment

	Temperature extremes			Average temp. [°C]	Average daily air humidity [%]	Daily totals of precipitation [mm/day]
Date	Max	Min	Ground minimum			
13. 7	22.7	13.4	9.4	17.9	59.0	0
19. 7	27.5	15.3	11.8	22.1	55.3	0
22. 7	29.9	13.2	7.8	22.1	47.7	0
05.10	14.7	1	-1.7	7.9	64.1	0
08.10	17.6	4.9	0.1	10.8	75.8	0
24.10	19.6	6.4	1.5	13.2	75.9	0

Source of the data: Meteostation of the Czech University of Life Sciences Prague, Faculty of Agrobiology, Food and Natural Resources, Department of Agroecology and Biometeorology; http://meteostanice.agrobiologie.cz

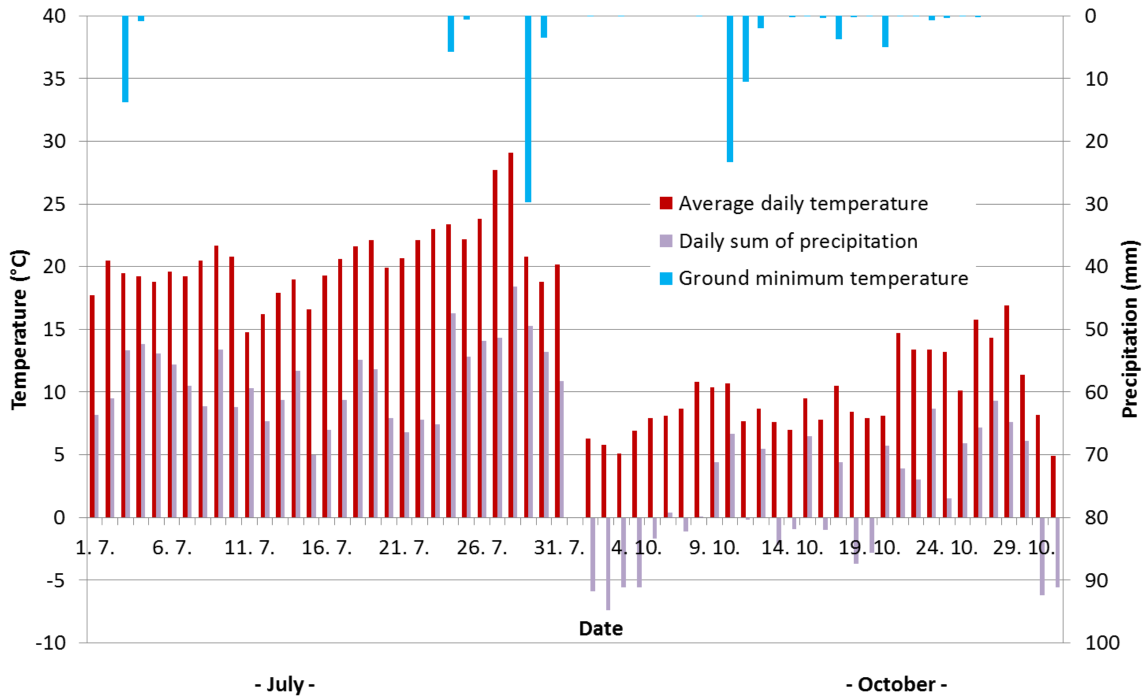


Figure 15 Average values of weather conditions during field experiment

Source of the data: Meteostation of the Czech University of Life Sciences Prague, Faculty of Agrobiology, Food and Natural Resources, Department of Agroecology and Biometeorology; http://meteostanice.agrobiologie.cz

2.1.2 Infiltration tests

As the following experimental equipment was used (see also Appendix 1): the double ring infiltrometer with the diameter of inner ring 35.7 cm and infiltration area 0.1 m^2, wooden piece in order to drive rings into the soil, hammer, bucket, measuring jug, knife, stopwatch, equipment for writing records, measuring tape, thermometer(℃), FDR soil moisture sensor: 5TE Soil Moisture, Temperature, and Electrical Conductivity (Decagon Devices, Inc.; www.decagon.com) plus EC Data logger. Water used for infiltration was tap water of drinking quality.

Figure 16 Photo of FDR soil moisture sensor: 5TE Soil Moisture, Temperature, and Electrical Conductivity (Decagon Devices, Inc.)

Before the measurement soil was prepared by carefully trimming newly grown plants without disturbing soil surface and plant roots (later during the time the grass was trimmed several times inside the rings). Two concentric rings were driven into soil (turf around the rings was cut with the knife) to a depth of about 15 cm by hammering wooden piece on the top of the rings. The soil surface in the inner ring was covered (during the measurements only) by a perforated metal plate which is used in order to dissipate the force of applied water, to distribute water uniformly inside the ring and to prevent disturbance of the soil. Metal plate consisted two nail points with different lengths used for observation of decreasing water level during infiltration (see Figure 17). Measurement was taken in the inner ring and the outer ring was used to ensure that water from the inner ring flows downwards not laterally (Appendix 2).

Figure 17 Metal plate with observation points (Báťková et al. 2013)

During the experiment tap water was kept as much as possible constant with the deviation of +1℃ or−1℃. The temperature of water was adjusted using electric kettle, if necessary. Measured temperature of water used for infiltration is shown in the Table 5. The temperature at the particular days was chosen according to the actual weather (air temperature) conditions. The aim was to carry out the infiltration experiments under several different (but naturally occurring) soil temperatures with adjusted water temperature. Before (see Table 6) and after the infiltration procedure the volumetric water content, temperature, and electrical conductivity of the soil with 5TE sensor were measured. Because of accuracy of calculation of saturated hydraulic conductivity initial water content was tried to keep similar before measurement. Results from Table 5 are actual values taken from thermometer and during measurement were kept similar with variation of 0.5°C, or Table 6 which also shows actual values read from 5TE sensor before measurement.

Table 5 Water temperatures

Date	**Temperature** [℃]
13.07	17.5
19.07	34.5
22.07	25
05.10	13
08.10	31
24.10	25

First, water was infiltrated in both rings, stopwatch was started and the time for the water to drop from the upper nail point to the lower nail point was registered, which indicates the amount 500 ml of infiltrated water to be added. During the measurement the same level of water in both rings was maintained in order to keep same pressure head. The procedure was repeated until steady-state flow conditions were reached.

Table 6 Soil properties before measurement

Date	**Temperature of soil** [°C]	**Soil moisture content** $[\boldsymbol{m^3/m^3}]$	**Bulk density** $[\boldsymbol{mS/cm}]$
13.07	22.5	0.251	0.31
19.07	27.7	0.223	0.29
22.07	23.0	0.191	0.23
5.10	13.2	0.188	0.11
8.10	18.6	0.19	0.21
24.10	22.1	0.269	0.28

2.2 Laboratory experiment

As the first step of this measurement 24 undisturbed soil core samples were taken from the field (see Figure 11, point 2.). Samples were set on saturation mat for few days. For the determination of saturated hydraulic conductivity constant head apparatus was set in the laboratory. 24 infiltration experiments with different water temperatures were carried out. After infiltration each sample was temperature measured and dried at 105°C.

2.2.1 Preparation of undisturbed soil samples

In total 24 undisturbed soil core samples were taken by the end of October 201, approximately 10 meters far from the area where field experiment was conducted. Exact location is shown in the Figure 11 and marked with the number 2. The soil description is given in chapter 2.1.1 and is valid for the site as well. An average actual water content of the samples was 0.3193 cm^3/ cm^3 with standard deviation 0.0071. Few days before soil sampling some precipitation occurred.

Taking of samples in the field is the basic procedure which has a significant effect on the results and conclusions based on the study. That is why a very careful attention was paid to the taking of samples. The following equipment was used: metal rings (diameter 8 cm and volume 251.3 cm^3), with cover lids, sampling head, hammer, knife, rubber band, geotextile cloth, solid box for transport of samples, balance scale, drying oven, watch glass, saturation mat.

Metal rings, geotextile cloths, rubber bands were mass tarred before soil sampling. In order to have good results the place had to be in natural state, not disturbed and during soil sampling make sure not to step on the soil because it would affect bulk density. Grass was removed from surface and soil in depth of 2-3 cm (huge root zone). Rings are placed in soil and pushed deeper with the hammer. After taking them out, the excessive soil was cut with the knife. Some of the samples had big roots and stones so the procedure had to be done again. Samples were covered with cover lids and transported to the laboratory. Samples were turned on their natural position, lower part covered with geotextile and hitched with rubber band and upper was covered with watch glass also previously tarred mass. They were weighed in order to determine the actual soil water content.

2.2.2 Laboratory measurement

The soil samples were saturated on saturation mat by capillary forces prior to the measurement of saturated hydraulic conductivity. Those samples still waiting for operation were kept in fridge. Capillary saturation of the samples took almost three days, until the mass of samples stopped to increase. Laboratory measurement was conducted in the fume hood. Each set of 8 samples were measured under different water temperatures. During measurement infiltrating water was tried to keep constant with variation of 0.5°C. After the measurement was done, each sample was temperature measured and dried at temperature of 105°C.

As the following laboratory equipment we used: lab-built constant head apparatus, stopwatch, scale, ruler, cloth, thermometer, tape, scissors, beakers. Schema of measurement setup is on Figure 17, and photo of the real measurement setup is in Appendix 3.

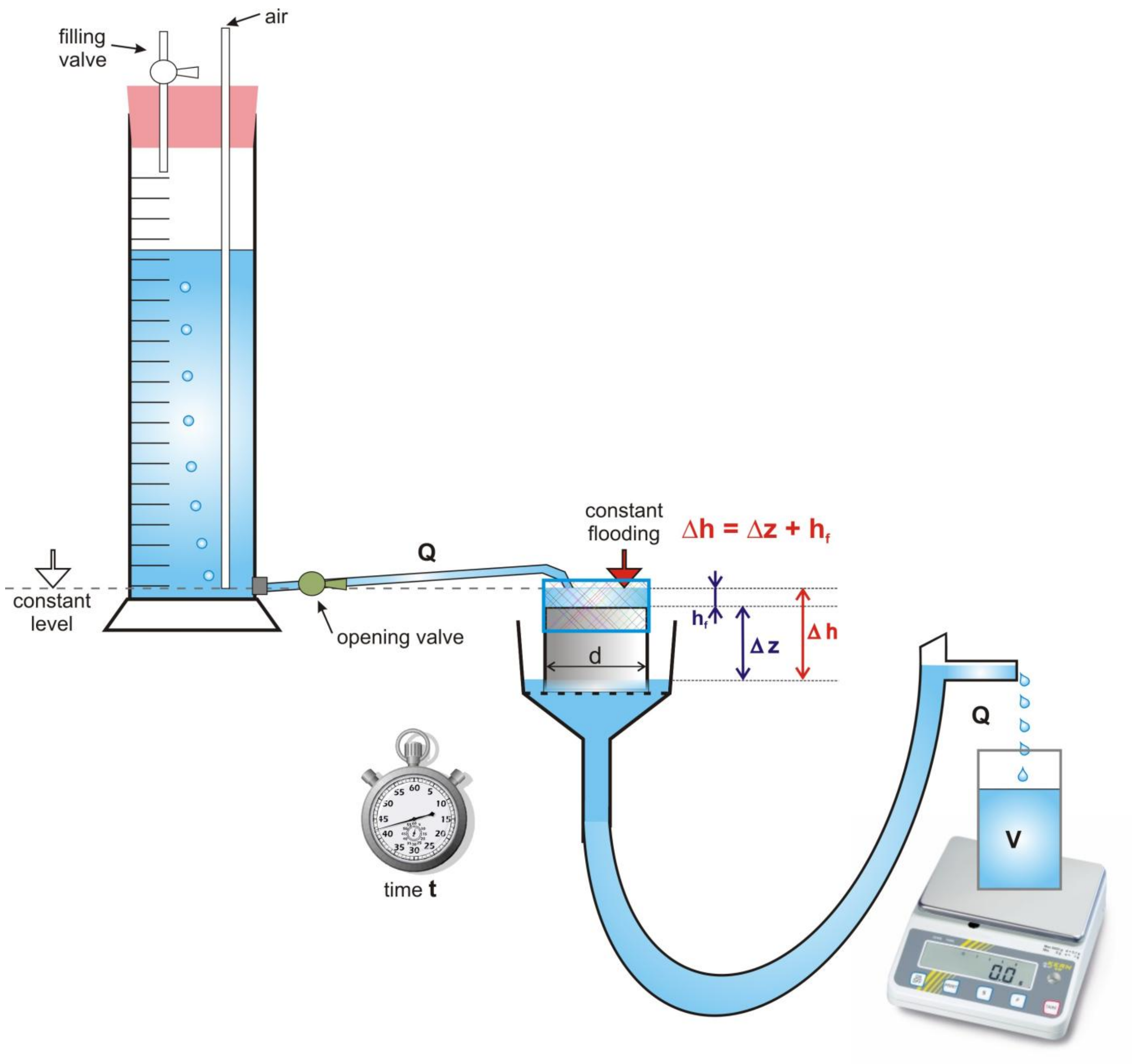


Figure 18 Laboratory measurement setup

Each sample was weighed after saturation in order to obtain a saturated water content value and one by one placed to constant head apparatus for measurement. Upper part of the ring was taped to keep the level of constant flooding. Small net was placed on the surface of the sample to prevent clogging of the surface pores of the sample. The outflowing water was registered as an increasing mass of water in a beaker placed on the scale. Water of different temperatures (see Table 7) was placed in the reservoir. Water used for infiltration was tap water of drinking quality, either heated with electronic kettle for higher temperature or cooled in the fridge for lower temperature.

When the measurement was finished, the temperature of sample was registered (by ordinary mercury thermometer) and samples were dried in drying oven at 105°C till constant weight. Every sample was treated with the same measurement procedure.

Table 7 Water temperatures

I set of samples		**II set of sample**		**III set of samples**	
Number of sample	**Temperature of water** (℃)	**Number of sample**	**Temperature of water** (℃)	**Number of sample**	**Temperature of water** (℃)
1	22.5	1	35	1	10
2	23	2	36	2	13
3	23	3	36.5	3	10
4	23	4	35	4	10.5
5	23	5	35	5	10
6	23	6	35	6	13
7	22.5	7	35	7	11
8	23	8	34	8	10.5

3. RESULTS

3.1 Field experiment

Based on field measurements carried out on permanently installed double ring infiltrometer on Experimental terrain station of soil moisture dynamics, the following characteristics were observed. The sensor readings from 5TE sensor in Table 8 were processed before infiltration and after infiltration. Average values of sensor readings are shown in Figures 19, 20 and 21. The sensor measurements were done in outer ring in order not to disturb the infiltration measurements.

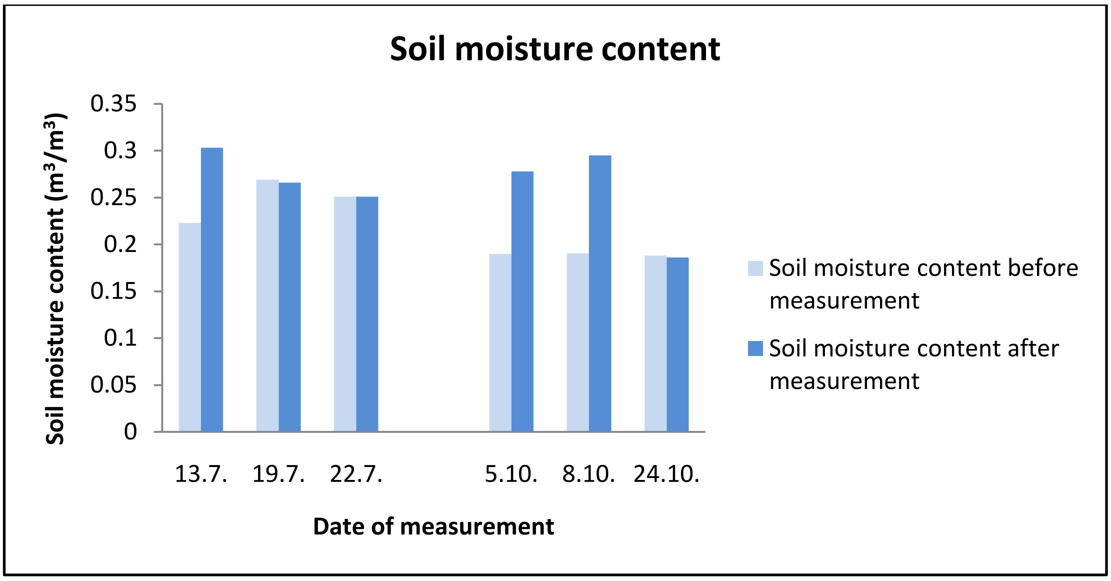


Figure 19 Comparison of soil moisture content before and after measurement

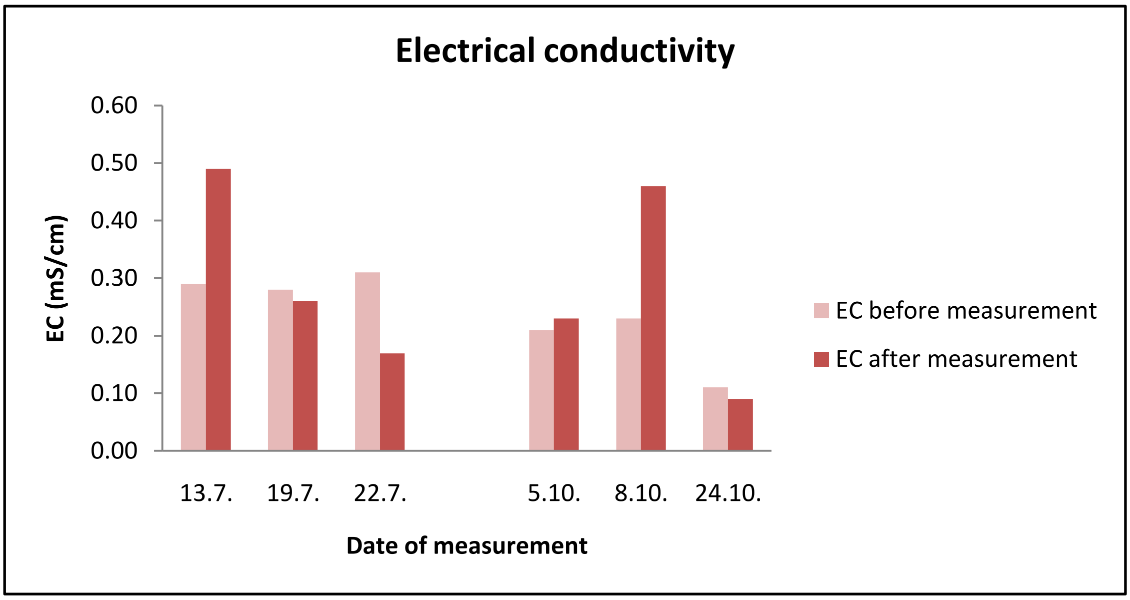


Figure 20 Comparison of electrical conductivity before and after measurement

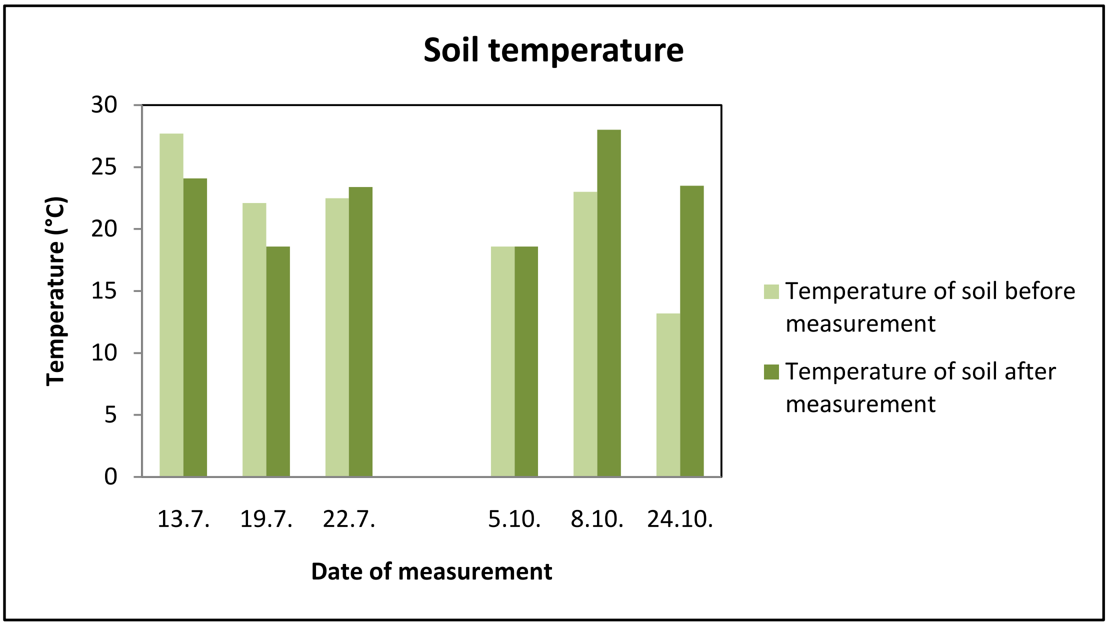


Figure 21 Comparison of soil temperature measured before and after measurement

Based on manual readings (Appendix 4) from field observation, the cumulative infiltration was fitted with Philip's equations (Eq.19) and infiltration rate (Eq. 20) and saturated hydraulic conductivity (Eq. 21) was calculated (see Figure 22). The graphs of cumulative infiltrations are included in Appendix 5.

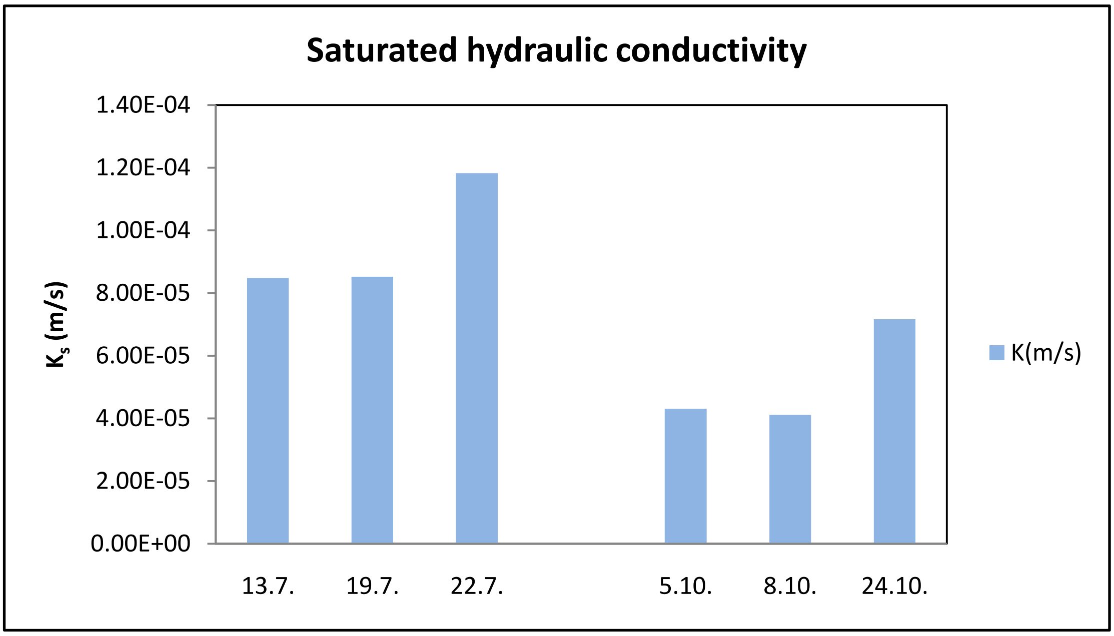


Figure 22 Saturated hydraulic conductivity of field measurement

3.2 Laboratory experiment

Before the measurement all data such as mass of: ring with soil, rubber, textile, watch glass, mass after saturation with data after measurement such as the mass of samples after drying were collected and used for calculation of initial water content, saturated water content and bulk density. For each set of samples individual graphs are shown in Appendix 6. In Figure 23 are shown average values of soil physical properties of 3 sets of 8 samples.

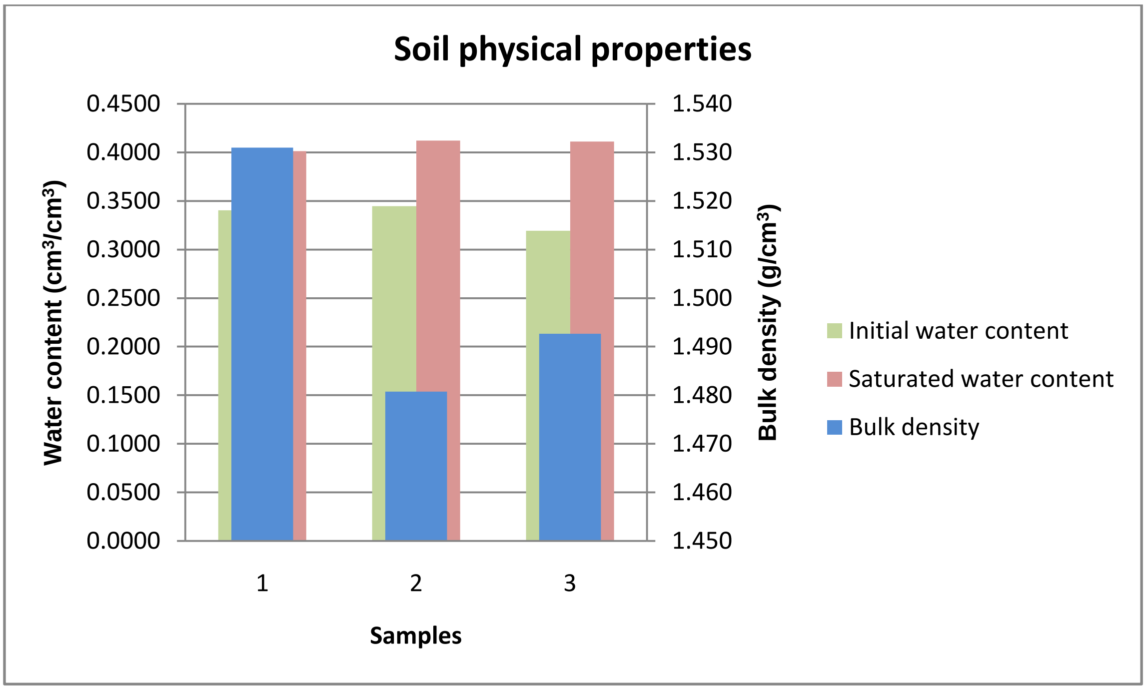


Figure 23 Soil properties of core samples

In Figure 24 we can see the affect of water used for infiltration on soil core samples with concern of the temperature of environment which was around $20°C$.

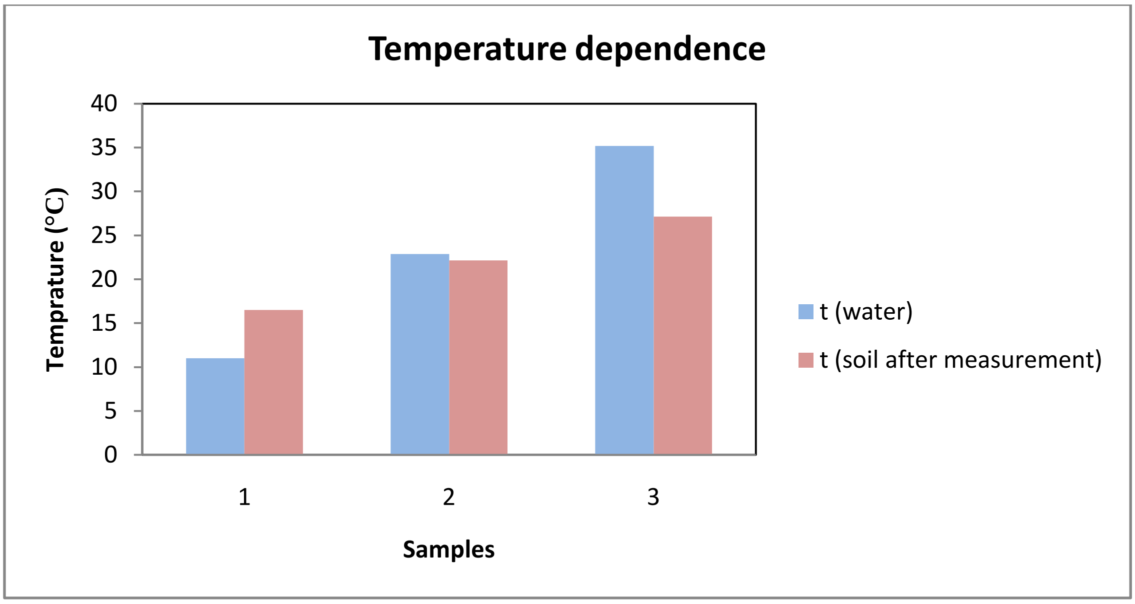


Figure 24 Comparison of soil temperature and water temperature measured before and after measurement

Saturated hydraulic conductivity was calculated in MS Excel with Eq. 11. Results were shown only in m/s due to easier comparison with field measurement (see Figure 25).

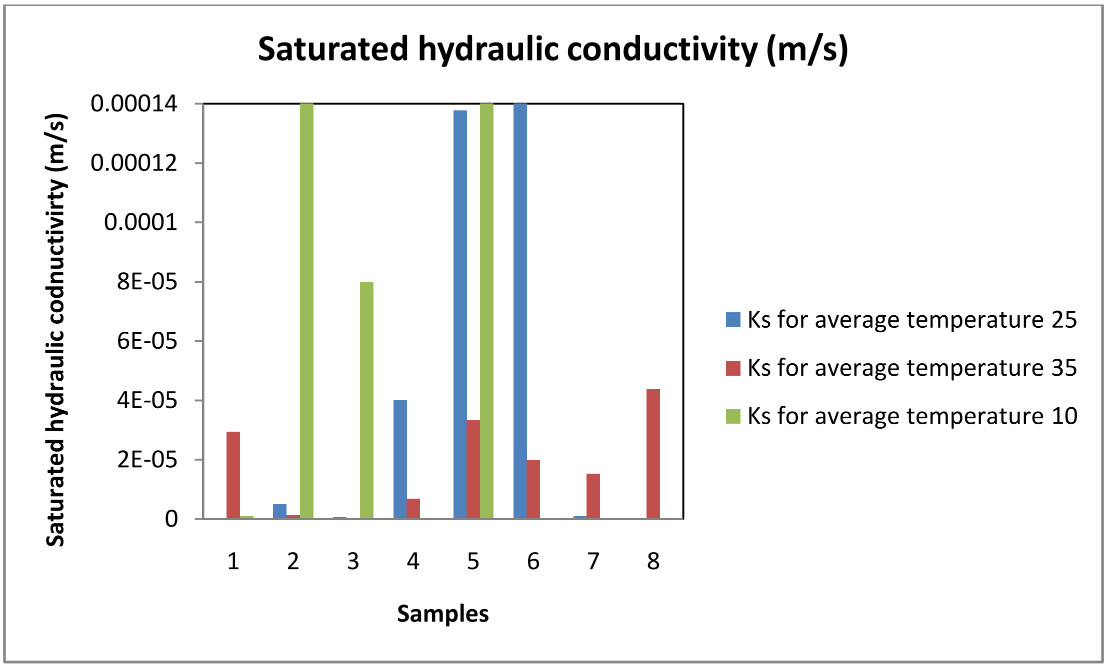


Figure 25 Saturated hydraulic conductivity of laboratory measurement

As you can see from the Figure 25 some of the samples don't have values of calculated saturated hydraulic conductivity. Those samples are 1 and 8 from I set of samples (average temperature 25°C), sample 3 form II set of samples (average temperature 35°C) and 4, 7, 8 from III set of samples (average temperature 10°C) . During measurement of these samples flow didn't occur within several hours so the measurement was stopped.

4. DISCUSSION

The field measurements were carried in two periods, in July and October. The initial water content was supposed to be as similar as possible, in order to limit the influence of it to the value of saturated hydraulic conductivity. However, even if irrigation and waiting for redistribution was applied, it was difficult to keep the same soil moisture content on the whole area of double ring infiltrometer. Thus, the initial soil moisture content is lower in the part measured in October, as shown on Figure 19. The final soil moisture content would be expected in general higher than the initial. However, in three measurements from six, it was almost the same. It can be caused either by measurement error, either by the 5TE sensor sensitivity to bulk density and temperature; the sensor was not calibrated. Values of saturated hydraulic conductivity were also lower in October period, which is in basic agreement with Eq. 4, which says, that with lower temperature the hydraulic conductivity also decreases. Nevertheless, the saturated hydraulic conductivity varies in a range of order 10^{-5} (m/s), only one value was higher in order 10^{-4}.

Based on the field measurements, the following trends can be observed, supported by coefficient of determination (R^2) in the following scatter plots. The coefficient of determination is squared correlation coefficient provides an information about how well the observed values are replicated by the model, as the proportion of total variation of the observed values explained by the model.

Firstly, there is no significant dependence of temperature either of water, either of soil on the values of saturated hydraulic conductivity, see Figure 26. On the other hand, there is strong dependence of K_s on initial soil moisture content, and consequently there is a correlation between initial electrical conductivity and K_s, but weak (see Figure 27). No significant relationship is shown either between final electrical conductivity or final soil moisture content, and K_s (see Figure 28), but there is a strong dependence of final electrical conductivity on final soil moisture content, as could be expected, see Figure 29. Final moisture content seems to be influenced also by temperature of water used, see Figure 30, which is confirmed also by dependence of final electrical conductivity and temperature of water.

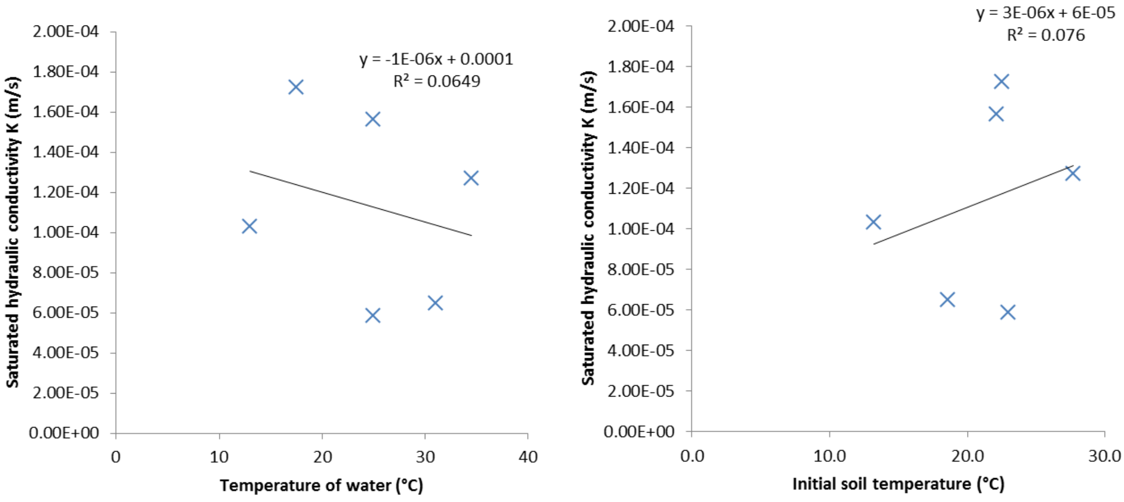


Figure 26 Final correlation between saturated hydraulic conductivity and temperature of water (on the left) or initial soil temperature (on the right).

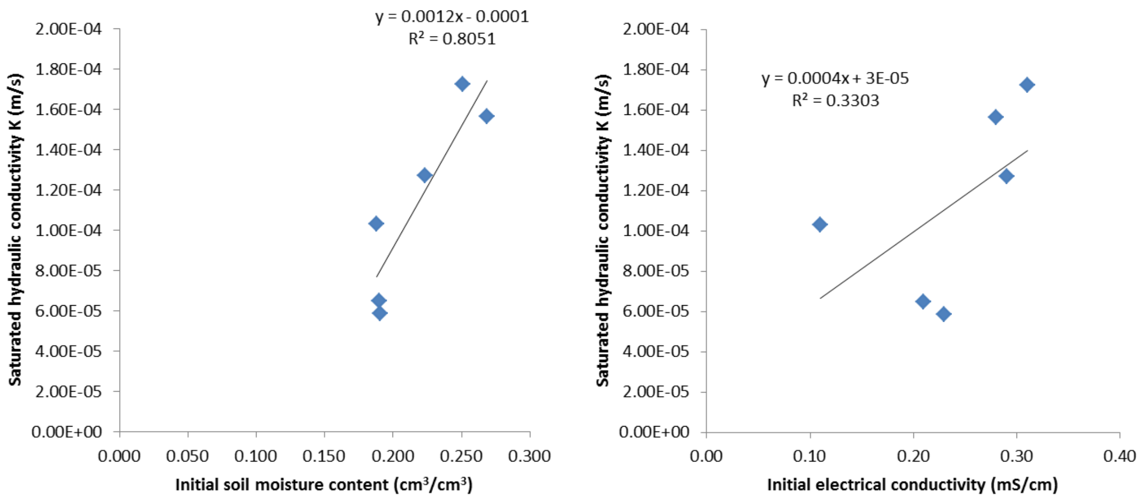
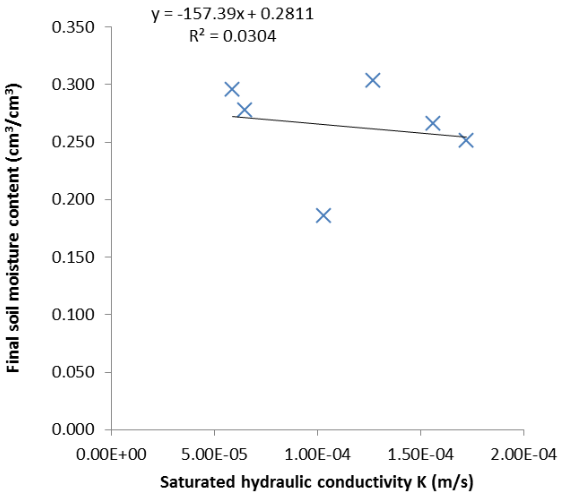


Figure 27 Correlation between saturated hydraulic conductivity and initial soil moisture content (on the left) and initial electrical conductivity (on the right).



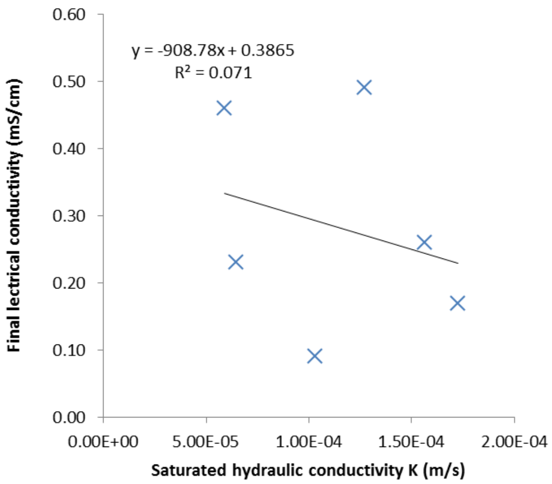


Figure 28 Relationship between final soil moisture content (on the left) and final electrical conductivity (on the right) and saturated hydraulic conductivity.

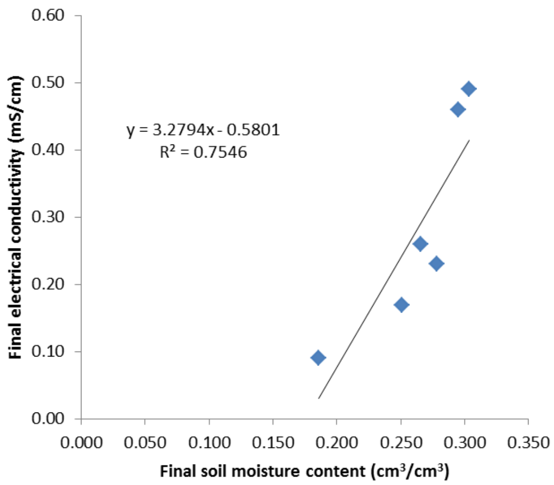


Figure 29 Dependence of final electrical conductivity on final soil moisture content.

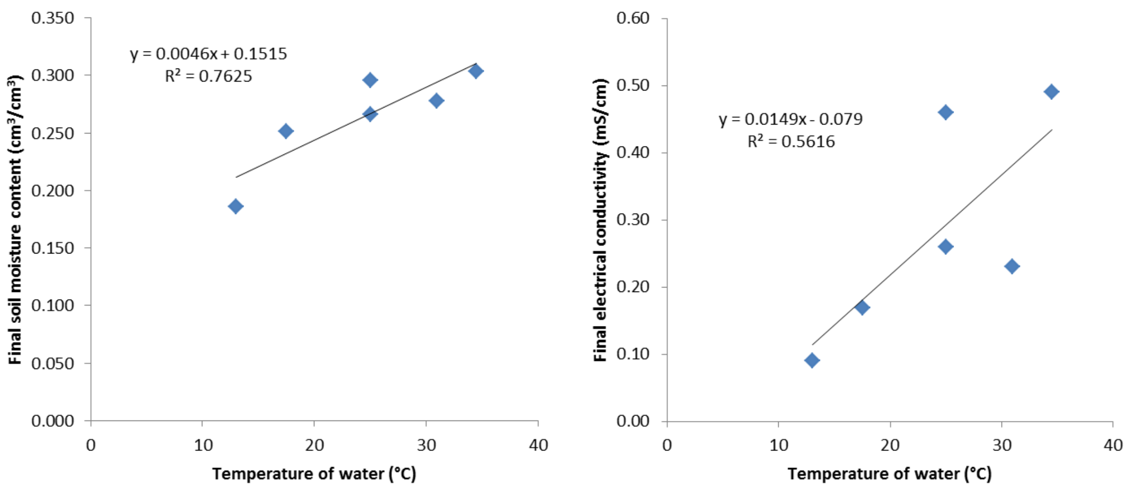


Figure 30 Relationship between final soil moisture content (on the left) and final electrical conductivity (on the right) and temperature of water used for infiltration.

Results were compared to a previous study conducted in the same area. Research from Danaa (2012) shows results of measurement from field experiment of saturated hydraulic conductivity in the range of 0.00005333-0.015167 m/s, which is approximately in the same range compared to this research, although it was measured with another type of infiltrometer. Laboratory measurement by Danaa (2012) showed results from range 0.00000333-0.000375 m/s and 0-0.000048 m/s which match the range obtained in this study; however it is less biased and with smaller variance than described here.

When laboratory measurements of saturated hydraulic conductivity carried out on constant head apparatus, then from the whole set of 24 undisturbed soil samples, 7 samples in total had to be omitted from further calculation. Six samples could not be measured at all, because they were sealed and there was no outflow even after several hours. The samples were probably affected by swelling which occurs in this soil, because when compared the basic physical properties, there is no difference between basic physical properties of those samples which were omitted and those which could be normally measured. Another sample was omitted based on the non-steady outflow from the sample. But in general, the values were very scattered with the range from 10^{-4} to 10^{-7} m/s and no significant trend can be observed from those measurements, because standard deviations from the average values are very high. Three of those cases with no outflow occurred when using the coldest water around 10°C, so it might also take effect on this case.

Generally, the values of saturated hydraulic conductivity obtained during lab measurement were in agreement with those obtained in a field, nevertheless, the lab measurements showed high variance and inconsistency due to insufficient representative elementary volume.

5. CONCLUSION

A relationship between increasing temperature and decreasing dynamic and kinematic viscosity of water is well described and as used in formula relating an intrinsic permeability of porous material to the saturated hydraulic conductivity, a model relationship between increasing temperature and increasing hydraulic conductivity can be easily calculated. However, changing temperature affects also other soil hydraulic, mechanical and biologic properties such as contact angle in pores, solubility of salts, microbiological activity, evapotranspiration etc. Description of all these changing properties in the whole complexity and in a particular soil profile might be rather impossible. Hence, modeling of transport processes in soil is very widely used in present times, and these changes related to changing of temperature are avoided or considered in general as seasonal changes. Studies about seasonal changes in soil hydraulic conductivity are often being carried out. The aim of this study was to investigate, if the particular daily temperature of soil and also water used for infiltration can affect the value of the saturated hydraulic conductivity in range of common operating temperatures, either on field, either on laboratory measurement. Field measurement were carried out on permanently installed double ring infiltrometer and laboratory measurements were carried out on constant head apparatus using 250 cm^3 core samples.

In the field, six infiltration experiments were carried out during July and October on permanent grass field station and with temperature ranging from 13°C to 34.5°C. The 5TE soil moisture sensor (Decagon Devices, Inc.) was employed in order to observe the soil moisture content, temperature of soil, and electrical conductivity. Temperature of water was measure by mercury thermometer. For serious conclusions further study would have to be carried out, but from this preliminary study can be concluded, that for double ring infiltrometer, the values of saturated hydraulic conductivity are not significantly affected by changing the temperature in a normal operational range. It was observed, that initial moisture content has more significant influence on the results even for this big infiltrometer.

In the lab, three sets of undisturbed soil samples were processed, 8 samples in each set. Three different temperatures of water were used, 10°C, 23°C and 35°C. In general, there is again no significant trend of normal operational temperature on the laboratory measurements. However, using of fresh cold tap water should be omitted, because it might cause problem with sealing the porous system of some samples.

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APPENDICES

List of Appendices

Appendix 1. Equipment used for field experiment

Appendix 2. Process of infiltration

Appendix 3. Laboratory equipment

Appendix 4. Manual readings from field measurement

No.	Time	INF.area:1000		Cumulative infiltration(cm)	
	t CUM(S)	i(cm3)	i(cm)	i(measured)	i(calculated)
1	0	0	0	0	0
2	31	500	500	0.5	0.479955351
3	79	500	1000	1	1.105447344
4	126	500	1500	1.5	1.697608457
5	183	500	2000	2	2.403969855
6	228	500	2500	2.5	2.956161434
7	277	500	3000	3	3.55370745
8	311	500	3500	3.5	3.9665725
9	343	500	4000	4	4.354075589
10	375	500	4500	4.5	4.740681656
11	418	500	5000	5	5.258965384
12	456	500	5500	5.5	5.715975169
13	491	500	6000	6	6.136173286
14	525	500	6500	6.5	6.543765882
15	567	500	7000	7	7.046530709
16	604	500	7500	7.5	7.48883826
17	641	500	8000	8	7.930633401
18	678	500	8500	8.5	8.371959915
19	713	500	9000	9	8.789033923
20	757	500	9500	9.5	9.312851624
21	797	500	10000	10	9.788600277
22	835	500	10500	10.5	10.24019524
23	877	500	11000	11	10.73894075
24	922	500	11500	11.5	11.27289307
25	978	500	12000	12	11.93680929
26	1015	500	12500	12.5	12.37515275
27	1055	500	13000	13	12.84877355
28	1097	500	13500	13.5	13.34579665
29	1142	500	14000	14	13.87802268
30	1184	500	14500	14.5	14.37450378
31	1226	500	15000	15	14.8707445
32	1272	500	15500	15.5	15.4139849
33	1316	500	16000	16	15.93336367
34	1364	500	16500	16.5	16.4997023
35	1421	500	17000	17	17.1719007
36	1457	500	17500	17.5	17.59627227
37	1504	500	18000	18	18.15011899
38	1560	500	18500	18.5	18.80974779
39	1610	500	19000	19	19.39846284

No.	Time	INF.area:1000		Cumulative infiltration(cm)	
	t CUM(S)	i(cm3)	i(cm)	i(measured)	i(calculated)
1	0	0	0	0	0
2	34	500	500	0.5	0.288301727
3	103	500	1000	1	0.873384644
4	177	500	1500	1.5	1.500864874
5	253	500	2000	2	2.145304029
6	296	500	2500	2.5	2.509920919
7	385	500	3000	3	3.264593087
8	435	500	3500	3.5	3.688566216
9	485	500	4000	4	4.112539344
10	533	500	4500	4.5	4.519553547
11	590	500	5000	5	5.002882913
12	648	500	5500	5.5	5.494691742
13	699	500	6000	6	5.927144333
14	759	500	6500	6.5	6.435912087
15	833	500	7000	7	7.063392316
16	917	500	7500	7.5	7.775667172
17	980	500	8000	8	8.309873313
18	1031	500	8500	8.5	8.742325904
19	1084	500	9000	9	9.19173742
20	1135	500	9500	9.5	9.624190011
21	1184	500	10000	10	10.03968368
22	1234	500	10500	10.5	10.4636568
23	1284	500	11000	11	10.88762993
24	1333	500	11500	11.5	11.3031236
25	1380	500	12000	12	11.70165834
26	1428	500	12500	12.5	12.10867254
27	1470	500	13000	13	12.46480997
28	1489	500	13500	13.5	12.62591976
29	1520	500	14000	14	12.8887831

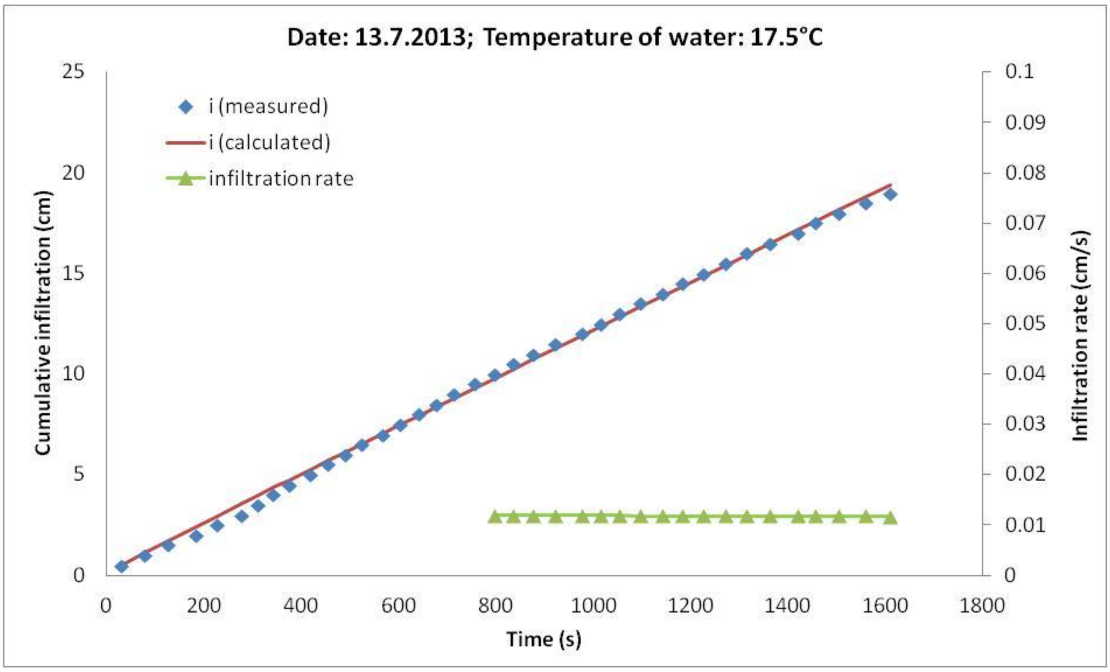
No.	Time	INF.area:1000		Cumulative infiltration(cm)	
	t CUM(S)	i(cm3)	i(cm)	i(measured)	i(calculated)
1	0	0	0	0	0
2	104	500	500	0.5	-0.548517908
3	226	500	1000	1	-0.051013988
4	359	500	1500	1.5	0.708598938
5	494	500	2000	2	1.590730317
6	626	500	2500	2.5	2.519219453
7	730	500	3000	3	3.283205855
8	780	500	3500	3.5	3.65864807
9	835	500	4000	4	4.076929369
10	891	500	4500	4.5	4.507974523
11	952	500	5000	5	4.982867365
12	1013	500	5500	5.5	5.462836126
13	1074	500	6000	6	5.947428599
14	1134	500	6500	6.5	6.428210131
15	1194	500	7000	7	6.912767521
16	1254	500	7500	7.5	7.400819463
17	1314	500	8000	8	7.892117935
18	1374	500	8500	8.5	8.386442938
19	1434	500	9000	9	8.883598249
20	1496	500	9500	9.5	9.400112121
21	1556	500	10000	10	9.902498522
22	1624	500	10500	10.5	10.47470143
23	1690	500	11000	11	11.03278186
24	1750	500	11500	11.5	11.54230981
25	1810	500	12000	12	12.05380636
26	1878	500	12500	12.5	12.63575997
27	1941	500	13000	13	13.1769554

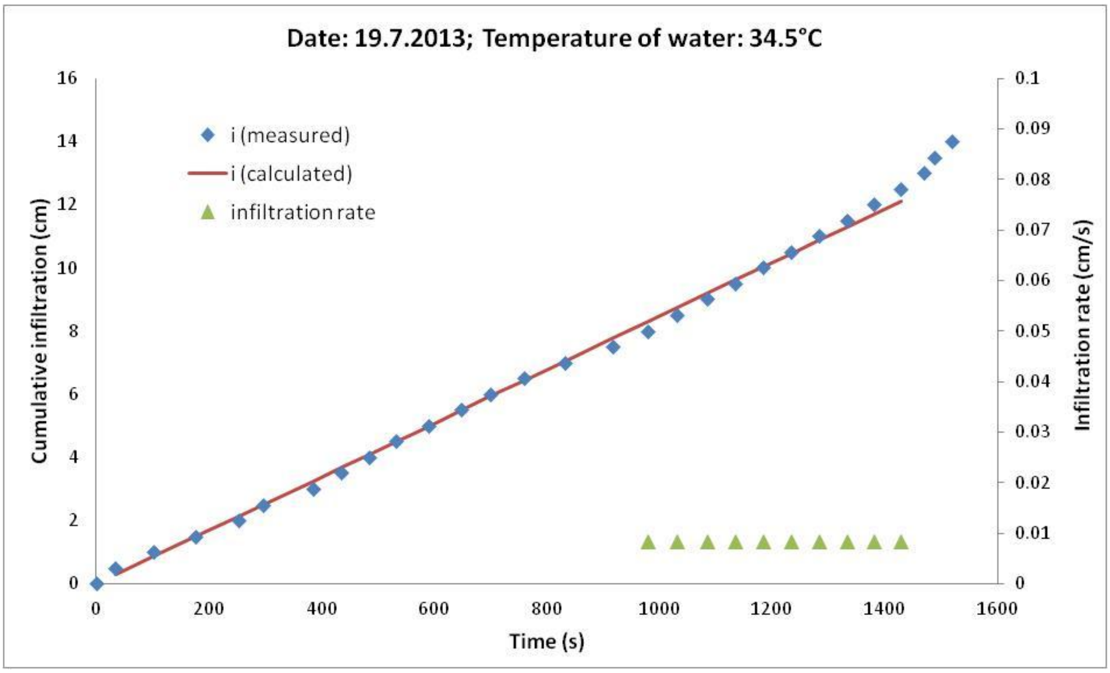
No.	Time	INF.area:1000		Cumulative infiltration(cm)	
	t CUM(S)	i(cm3)	i(cm)	i(measured)	i(calculated)
1	0	0	0	0	0
2	33	500	500	0.5	0.364799754
3	66	500	1000	1	0.648919318
4	124	500	1500	1.5	1.12021462
5	179	500	2000	2	1.552455537
6	303	500	2500	2.5	2.502257816
7	368	500	3000	3	2.992108124
8	430	500	3500	3.5	3.45596156
9	577	500	4000	4	4.546200701
10	565	500	4500	4.5	4.457609587
11	636	500	5000	5	4.980902462
12	704	500	5500	5.5	5.480307406
13	775	500	6000	6	6.000160055
14	844	500	6500	6.5	6.504021931
15	911	500	7000	7	6.992161787
16	982	500	7500	7.5	7.508376829
17	1048	500	8000	8	7.987354501
18	1114	500	8500	8.5	8.465561632
19	1180	500	9000	9	8.943065775
20	1246	500	9500	9.5	9.419925155
21	1314	500	10000	10	9.910613862
22	1380	500	10500	10.5	10.38631321
23	1456	500	11000	11	10.93345907
24	1529	500	11500	11.5	11.45841924
25	1595	500	12000	12	11.93257954
26	1665	500	12500	12.5	12.43502925
27	1737	500	13000	13	12.95138367
28	1809	500	13500	13.5	13.46730864
29	1882	500	14000	14	13.98998685
30	1954	500	14500	14.5	14.50512158
31	2033	500	15000	15	15.06992549
32	2104	500	15500	15.5	15.57718504
33	2178	500	16000	16	16.10554501
34	2248	500	16500	16.5	16.60504739

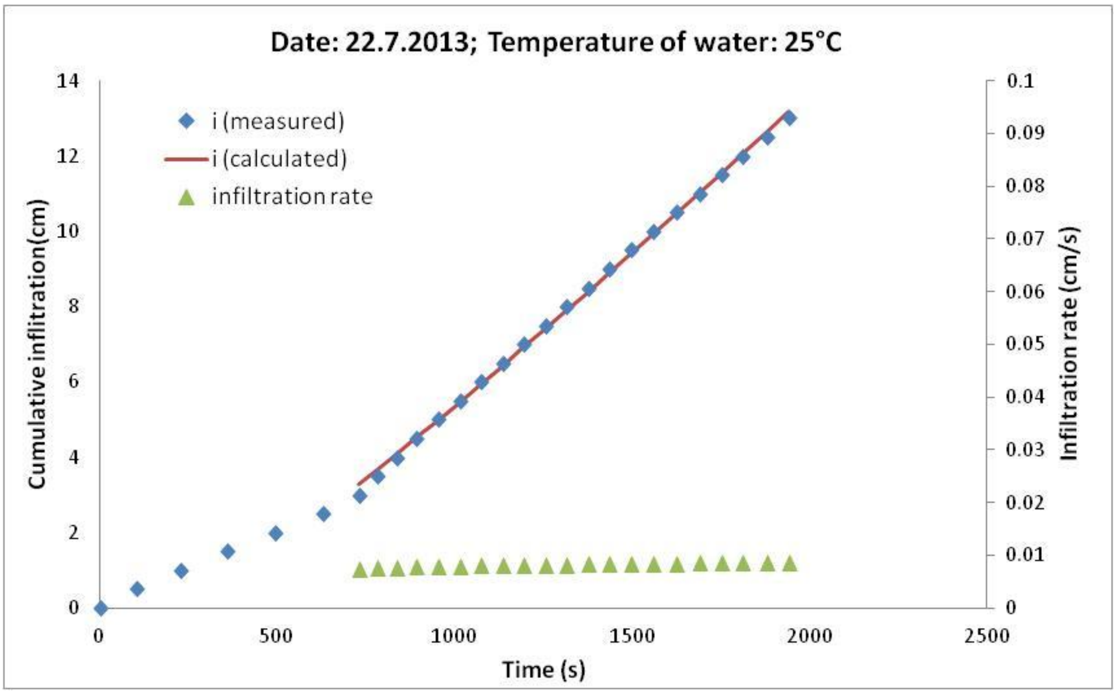
No.	Time	INF.area:1000		Cumulative infiltration(cm)	
	t CUM(S)	i(cm3)	i(cm)	i(measured)	i(calculated)
1	0	0	0	0	0
2	67	500	500	0.5	0.289357626
3	184	500	1000	1	0.793878754
4	304	500	1500	1.5	1.311193191
5	428	500	2000	2	1.845686464
6	549	500	2500	2.5	2.367210952
7	670	500	3000	3	2.888710317
8	795	500	3500	3.5	3.427429913
9	915	500	4000	4	3.944586319
10	1038	500	4500	4.5	4.474659853
11	1190	500	5000	5	5.129696606
12	1314	500	5500	5.5	5.664059486
13	1433	500	6000	6	6.176868848
14	1545	500	6500	6.5	6.659507747
15	1670	500	7000	7	7.198161944
16	1770	500	7500	7.5	7.629081686
17	1885	500	8000	8	8.124635792
18	2001	500	8500	8.5	8.624495503
19	2101	500	9000	9	9.055406412
20	2220	500	9500	9.5	9.568187463
21	2332	500	10000	10	10.05080223
22	2445	500	10500	10.5	10.5377236
23	2558	500	11000	11	11.02464267
24	2670	500	11500	11.5	11.50725061
25	2777	500	12000	12	11.9683117
26	2880	500	12500	12.5	12.41213525
27	2997	500	13000	13	12.91628253
28	3112	500	13500	13.5	13.41181014
29	3226	500	14000	14	13.90302718
30	3340	500	14500	14.5	14.39424268
31	3448	500	15000	15	14.85960337

No.	Time	INF.area:1000		Cumulative infiltration(cm)	
	t CUM(S)	i(cm3)	i(cm)	i(measured)	i(calculated)
1	0	0	0	0	0
2	51	500	500	0.5	0.348173881
3	163	500	1000	1	0.903739224
4	270	500	1500	1.5	1.398957999
5	390	500	2000	2	1.93786009
6	510	500	2500	2.5	2.466743848
7	626	500	3000	3	2.971690534
8	744	500	3500	3.5	3.480719284
9	864	500	4000	4	3.99469444
10	984	500	4500	4.5	4.505705201
11	1104	500	5000	5	5.014280358
12	1224	500	5500	5.5	5.520808002
13	1344	500	6000	6	6.025582949
14	1464	500	6500	6.5	6.5288354
15	1585	500	7000	7	7.034926616
16	1705	500	7500	7.5	7.535641944
17	1830	500	8000	8	8.056090015
18	1945	500	8500	8.5	8.533982458
19	2070	500	9000	9	9.052523324
20	2193	500	9500	9.5	9.561924634
21	2310	500	10000	10	10.04576504
22	2430	500	10500	10.5	10.54134593
23	2550	500	11000	11	11.03630165
24	2670	500	11500	11.5	11.53067587
25	2786	500	12000	12	12.00805468
26	2903	500	12500	12.5	12.48906698
27	3020	500	13000	13	12.96962426
28	3135	500	13500	13.5	13.44154887
29	3255	500	14000	14	13.93357467
30	3380	500	14500	14.5	14.44567321
31	3500	500	15000	15	14.93689862

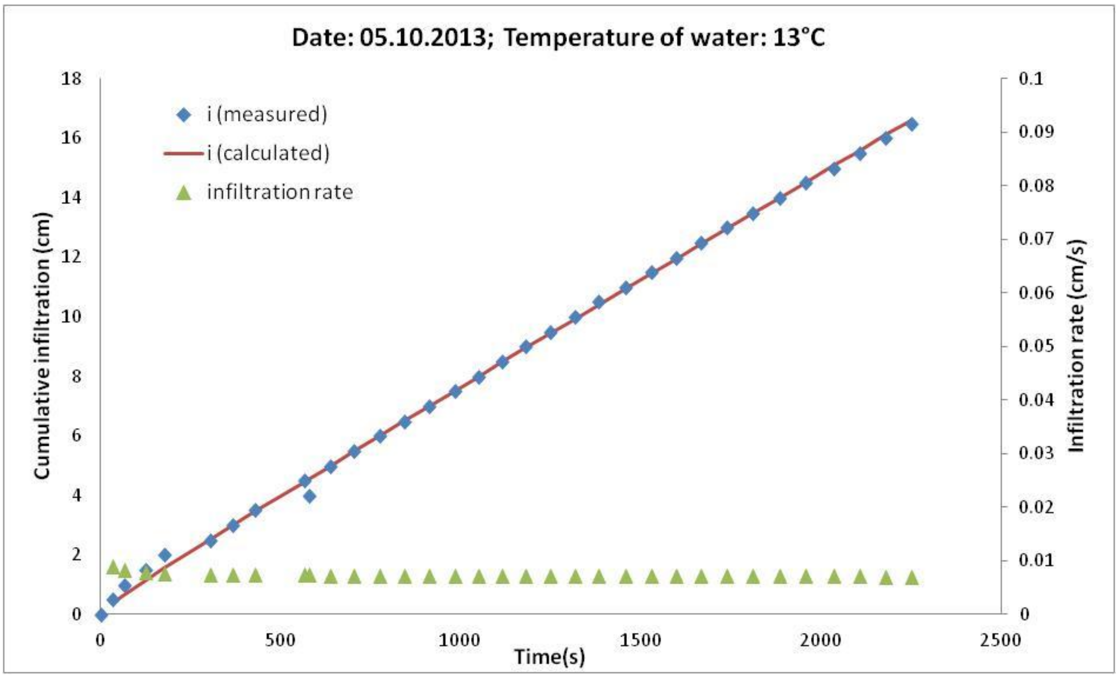
Appendix 5. Cumulative infiltrations from double ring infiltrometer.



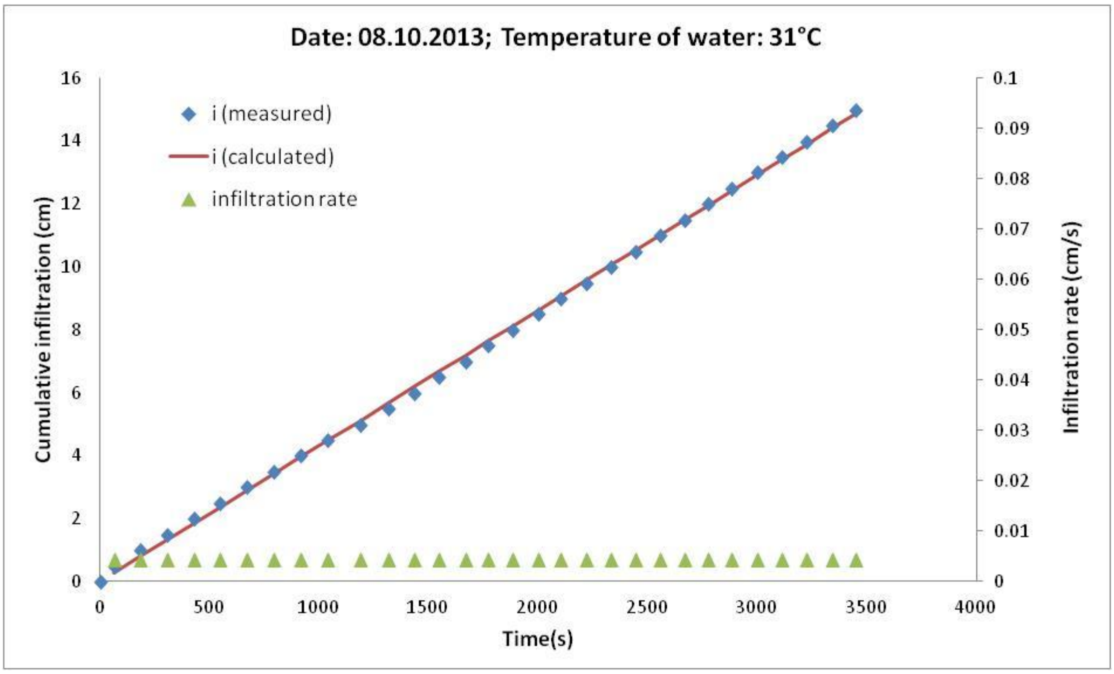


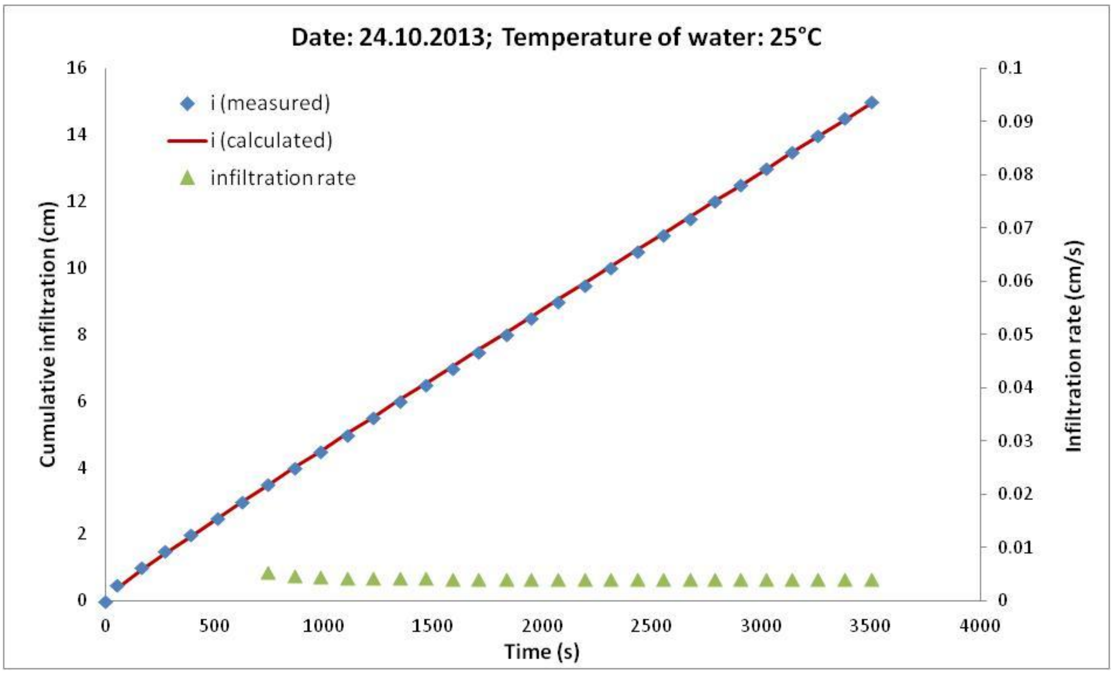


Date: 22.7.2013; Temperature of water: 25°C
i (measured)
i (calculated)
infiltration rate
Cumulative infiltration(cm)
Infiltration rate (cm/s)
Time (s)
0
2
4
6
8
10
12
14
0
0.01
0.02
0.03
0.04
0.05
0.06
0.07
0.08
0.09
0.1
0
500
1000
1500
2000
2500

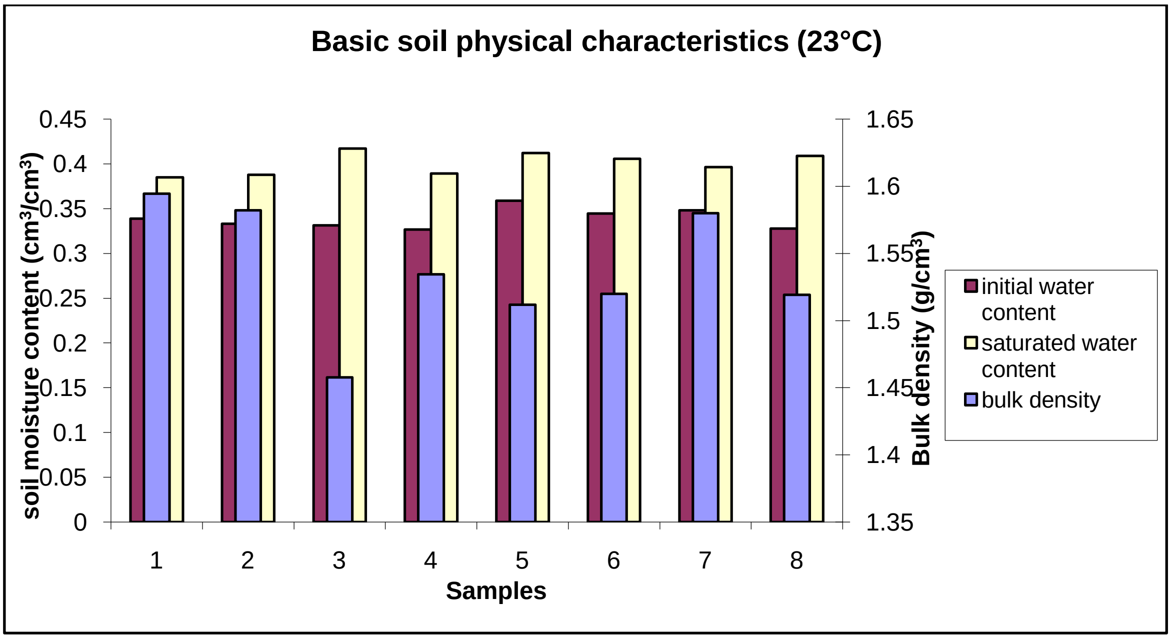


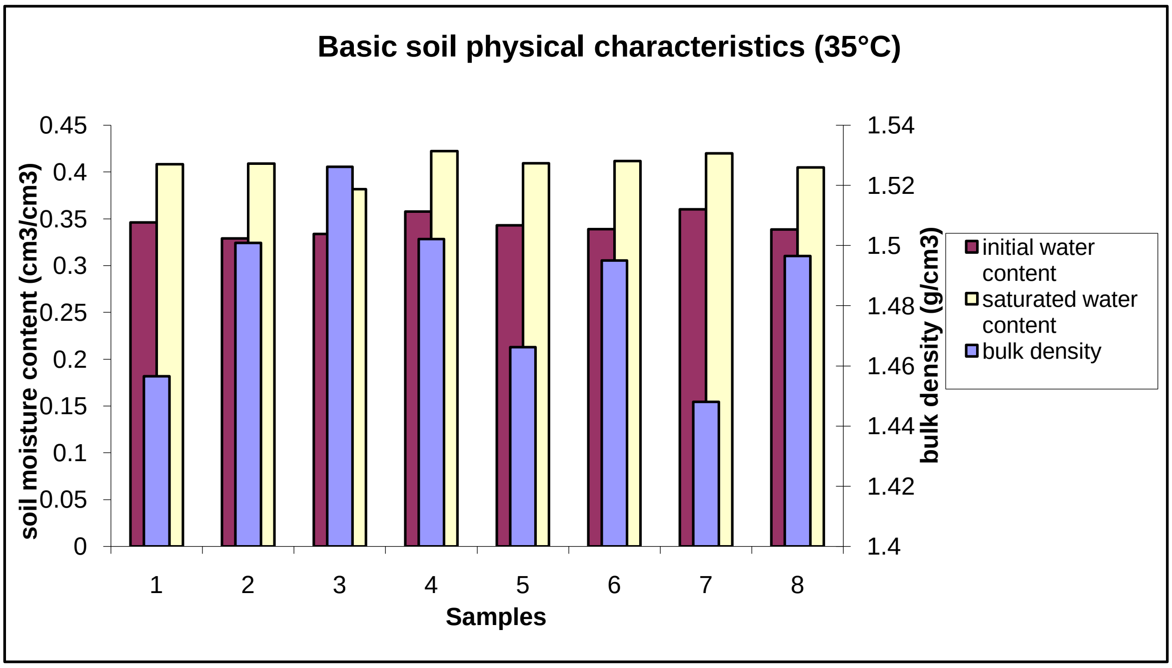
Date: 05.10.2013; Temperature of water: 13°C
i (measured)
i (calculated)
infiltration rate
Cumulative infiltration (cm)
Infiltration rate (cm/s)
Time(s)
0
2
4
6
8
10
12
14
16
18
0
0.01
0.02
0.03
0.04
0.05
0.06
0.07
0.08
0.09
0.1
0
500
1000
1500
2000
2500

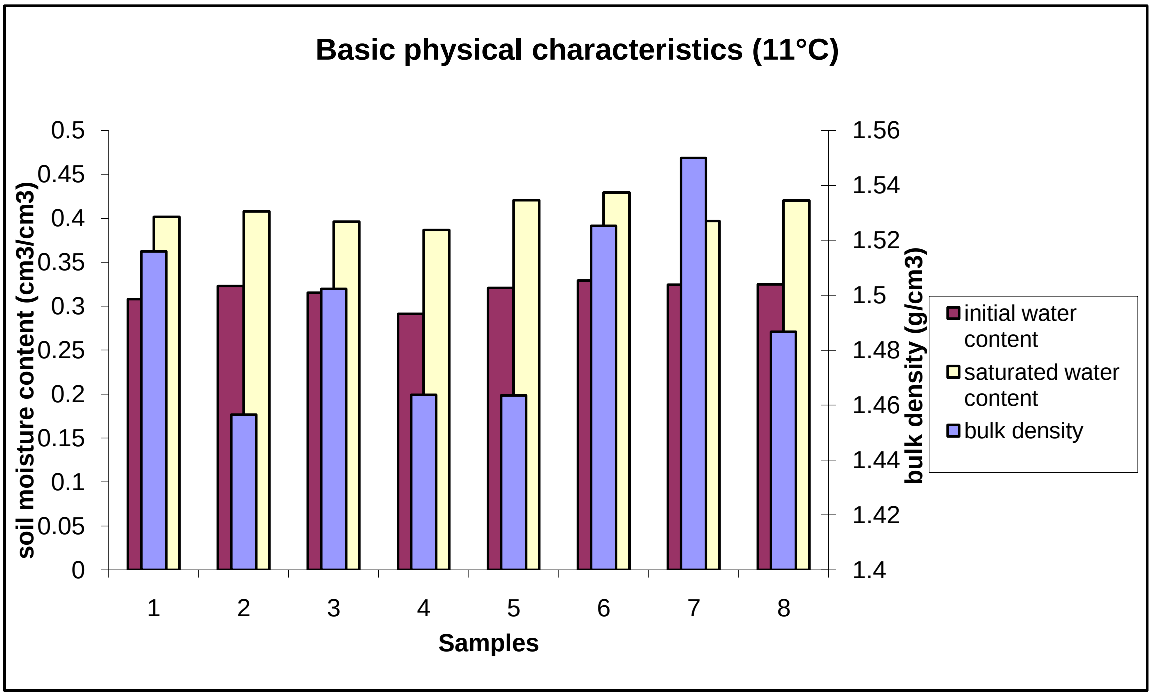

Date: 08.10.2013; Temperature of water: 31°C
i (measured)
i (calculated)
infiltration rate
Cumulative infiltration (cm)
Infiltration rate (cm/s)
Time(s)
0
2
4
6
8
10
12
14
16
0
0.01
0.02
0.03
0.04
0.05
0.06
0.07
0.08
0.09
0.1
0
500
1000
1500
2000
2500
3000
3500
4000


Date: 24.10.2013; Temperature of water: 25°C
i (measured)
i (calculated)
infiltration rate
Cumulative infiltration (cm)
Infiltration rate (cm/s)
Time (s)
0
2
4
6
8
10
12
14
16
0
0.01
0.02
0.03
0.04
0.05
0.06
0.07
0.08
0.09
0.1
0
500
1000
1500
2000
2500
3000
3500
4000

Appendix 6. Basic soil physical properties of undisturbed soil samples used for laboratory measurement of saturated hydraulic conductivity.






Basic physical characteristics (11°C)
soil moisture content (cm3/cm3)
0.5
0.45
0.4
0.35
0.3
0.25
0.2
0.15
0.1
0.05
0
1.56
1.54
1.52
1.5
1.48
1.46
1.44
1.42
1.4
bulk density (g/cm3)
1
2
3
4
5
6
7
8
Samples
initial water content
saturated water content
bulk density