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Temperature Influence on Measurement of Unsaturated Soil Hydraulic Conductivity using Minidisk Infiltrometer

Diploma Thesis

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Declaration

I certify that I prepared the diploma thesis on "Temperature Influence on Measurement of Unsaturated Soil Hydraulic Conductivity Using Minidisk Infiltrometer" by myself and I used only the literature sources which I quote and mention in the attached bibliography.

In Prague, **Signature:**

Acknowledgement

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Summary

Unsaturated hydraulic conductivity is one of the most important soil hydraulic properties with great influence on transportation of soil water, nutrient, chemical contaminants and heat. Soil surface temperature changes with seasons of a year and times of a day, and change in weather conditions. These days climate change causes temperature extremes (maximum and minimum) either in years or in days of the same area. As a result the change in temperature causes the change in physical, chemical and biological properties of water and its interaction with soil which could change the rate of water flow in soil. Hence, to collect the best primary data is very essential to reach at good scientific decision. In addition, tension Minidisk Infiltrometer (Decagon Devices, Inc.) is becoming useful hydraulic conductivity near saturation measuring instrument because of its simplicity and handy to carry at field. Therefore, evaluation of temperature influence on measurement of hydraulic conductivity is the focus of the study. The study was conducted at field and in laboratory. The measured data were collected from four replications of field measurement and six replications of laboratory measurement at 10, 25 and 35^{o}C with constant applied tension (2cm). Field measurement was conducted in summer and at the Experimental terrain station of soil moisture dynamics of Department of Water Resources, FAFNR, CULS and laboratory measurement carried out on artificially packed soil columns with the fine earth from the site area. Unsaturated hydraulic conductivity, K(h), was calculated using the approach proposed by Zhang (1997) which is also recommended by the producer. Soil water content change and K(h) was also calculated viscosity and density changes. Comparison of measured K(h) to recalculated K(h) is discussed. Measured K(h) was not significantly varied at field but in the laboratory measured K(h) was significantly varied.

Key words: Temperature, hydraulic conductivity, minidisk infiltrometer, tension.

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1. Introduction

Accurate characterisation of soil hydraulic properties is crucial to solve many hydrological, engineering, and environmental issues linked to soil water storage and transport in the vadose zone (Moret, et al., 2004).

In recent decades unsaturated flow has become one of the most important and active topics of research in soil physics and hydrology. Because an unsaturated soil zone has several functions including (i) storage of water and nutrients, and (ii) transmission of water and other substances and is of importance to the biosphere. It is the transmission zone, which redistributes the water. Therefore, this zone controls the ground water replenishment as well as evaporation from soil surface. The unsaturated zone experiences transport processes of various kinds, chemical reactions, biological activity of roots, rodents, worms, microbiota, and other organisms. It is also a zone of human activity and is used for the cultivation and disposal of waste. This zone is also drastically disturbed by surface mining and construction of civil structures (e.g., buildings, roads, etc). The vadose zone is in direct contact with the atmosphere through gaseous fluxes of water vapor and greenhouse gases (CO_2, CH_4, and N_2O) (Hillel, 1998 and Lal and Shukla, 2004).

According to i.e. Ankeny (1992) unsaturated hydraulic conductivity measurements could provide rapid information on the effects of soil compaction, soil texture, management, and root growth on soil hydraulic properties, and hydraulic conductivity function near saturation is useful in predicting solute movements.
Hydraulic conductivity and soil water retention curve (SWRC) are the most important soil hydrophysical characteristics for sustainable use of water as a natural resource (Mihalikova et al., 2013).

Therefore, accurate and easy measurement instruments to evaluate hydraulic conductivity at field and in the laboratory are very important like Minidisk Infiltrometer (MDI), etc. Introduction of minidisk infiltrometer made repeated measurements much easier owing to the extreme portability of the device, but their practical use rests on the approximation of soil type (Homolák et al., 2009)

Cook (2007) refer that the disc permeameter has become a popular apparatus for measuring in situ sorptivity, S, and hydraulic conductivity, K, of the soil at some prescribed pressure heads of -20, -40, and -100 mm and is now a commonly used method for measuring the hydraulic properties of field soils.

Since minidisk infiltrometer is one widely used instrument to measure unsaturated soil hydraulic conductivity and temperature is everywhere and around the measurement activity that could potentially affect measurement results, the objective of the study focuses on the influence of temperature on practical measurement of unsaturated soil hydraulic conductivity under normal range of operational temperatures using minidisk infiltrometer.

2. Objectives of the study

The main focus of the study area is the evaluation of the practical effect of temperature in measuring unsaturated soil hydraulic conductivity at field and controlled laboratory condition using minidisk infiltrometer at constant tension (2 cm).

To evaluate different parameters which could influence soil hydraulic conductivity such as initial soil water content, soil electrical conductivity and temperature sensitive water physical properties (i.e. viscosity and density), etc.

Hypothesis

The measurement of unsaturated soil hydraulic conductivity using MDI will be influenced or changed significantly by the change in temperature.

3. Literature Review

3.1. Properties of water and temperature

Water has unique physical properties. These properties permit life. Life depends on water. Temperature affects these physical properties of water, including specific heat, viscosity, density, surface tension. Therefore, the matric potential (tension) of water in the soil is affected by temperature changes. The major effect of temperature occurs at the soil surface, where temperature changes are greatest.

The selected properties which are related to this study are listed below (cited from Kirkham, 2005):

Specific Heat is a quantity of heat necessary to raise the temperature of mass (gram) of a substance from T_1 to T_2 in ˚C. Water has the highest specific heat of any known substance except liquid ammonia, which is about 13 percent higher. One sees that the specific heat of water decreases with an increase of temperature up to 35˚C, and then the specific heat increases with further increase in temperature.

Surface Tension: Water has a much higher surface tension than most other liquids because of the high internal cohesive forces between molecules. The value of surface tension at different temperatures is different.

Density: Water has a high density and is remarkable in having its maximum density at 4˚C instead of at the freezing point.

Expansion Up on Freezing: Water expands on freezing, so that ice has a volume about 9% greater than the liquid water from which it was formed. This explains why ice floats and pipes and radiators burst when the water in them freezes. If ice sank, bodies of water in the cooler parts of the world would be filled permanently with ice, and aquatic organisms could not survive. Expansion of freezing water has a fundamental role in soil structure forming processes.

Adsorption: Water tends to be adsorbed, or bound strongly, to the surfaces of clay micelles, cellulose, protein molecules, and many other substances. This characteristic is of great importance in soil and plant water relations.

Viscosity: Water has higher viscosity than acetone liquid and gases such as Helium, Methane, Nitrogen and etc. Fluids possess a definite resistance to change of form and many solids show a gradual yielding to forces tending to change their form. This property of internal friction is called viscosity. Value of the viscosity of water at different temperatures is different.

When a fluid is moved in shear (that is to say, when adjacent layers of fluid are made to slide over each other), the force required is proportional to the velocity of shear. The proportionality factor is called the viscosity. As such, it is the property of the fluid to resist the rate of shearing and can be visualized as an internal friction. The coefficient of viscosity η is defined as the force per unit area necessary to maintain a velocity difference of 1 m/sec between two parallel layers of fluid which are 1 m apart (Hillel, 1998).

The viscosity equation is

$$\tau = F/A = \mu\ \partial u/\partial y \ldots\ldots\ldots\ldots\ldots\ldots 1$$

where τ is the shearing stress (NL^{-2}), F is force acting on an area (N), A is cross sectional area (L^2), μ is the coefficient of dynamic viscosity ($ML^{-1}T^{-1}$) and $\partial u/\partial y$ is the change in velocity gradient perpendicular to the stressed area *A*.

The ratio of the dynamic viscosity of a fluid to its density is called the kinematic viscosity designated as v. It expresses the shearing-rate resistance of a fluid independently of the density. Thus, while the dynamic viscosity of water exceeds that of air by a factor of about 50 (at room temperature), its kinematic viscosity is actually lower.

Fluids of lower viscosity flow more readily and are said to possess greater (which is the reciprocal of viscosity). As shown in Table 3.2, the viscosity of water diminishes by over 2% per 1^{o}C rise in temperature, and thus decreases by more than half as the temperature increases from 5 to 35^{o}C. The viscosity is also affected by the type and concentration of solutes (Hillel, 1998).

The density of liquid water is nearly constant but it varies with the change in temperature and solute concentration. Hence, changes in fluidity are primarily from changes in viscosity. Fluidity varies with composition of the fluid and with temperature. Permeability is property of the porous medium and its pore geometry only. In a stable porous body the same permeability will be obtained with different fluids such as water, air, or oil. However, water interacts with the solid matrix that modify the flow characteristics, so that hydraulic conductivity is dependent on the properties of both water and soil (Hillel, 1998).

The hydraulic conductivity, K, (LT^{-1}) depends on the characteristics of both the soil and of the fluid. The soil characteristics that affect K are total porosity, pore sizes distribution, tortuosity and/or the soil's pore geometry. The fluid characteristics that affect conductivity are density and viscosity.

Therefore, K is the product of two factors that are intrinsic permeability of the soil, k, (L^2) and the fluidity of permeating liquid, f, ($L^{-1}T^{-1}$).

$$K = kf \quad \text{.....2}$$

$$f = \rho g/\mu \quad \text{.....3}$$

$$k = K\mu/\rho g \quad \text{.....4}$$

where μ is dynamic viscosity (NTL^{-2}), ρ is the fluid density (ML^{-3}), and g is gravitational acceleration (LT^{-2}).

Table 3.1. Physical properties of liquid water (Hillel, 1998)

Temperature (^oC)	Density (kg/m^3)	Specific heat (J/kg deg)	Latent heat (Vaporization) (J/kg)	Surface tension (kg/s^2)	Thermal Conductivity (J/m s deg)	Dynamic Viscosity (kg/m s)	Kinematic viscosity (m^2/s)
-10	0.99794×10^3	4.27×10^3	2.53×10^6	-	-	-	-
-5	0.99918×10^3	4.23×10^3	2.51×10^6	7.64×10^{-2}	-	-	-
0	0.99987×10^3	4.22×10^3	2.50×10^6	7.56×10^{-2}	0.561	0.1787×10^{-2}	1.79×10^{-6}
4	1.0000×10^3	4.21×10^3	2.49×10^6	7.5×10^{-2}	0.570	0.1567×10^{-2}	1.57×10^{-6}
5	0.99999×10^3	4.207×10^3	2.49×10^6	7.48×10^{-2}	0.574	0.1519×10^{-2}	1.52×10^{-6}
10	0.99973×10^3	4.194×10^3	2.48×10^6	7.42×10^{-2}	0.587	0.1307×10^{-2}	1.31×10^{-6}
15	0.99913×10^3	4.19×10^3	2.47×10^6	7.34×10^{-2}	0.595	0.1139×10^{-2}	1.14×10^{-6}
20	0.99823×10^3	4.186×10^3	2.46×10^6	7.27×10^{-2}	0.603	0.1002×10^{-2}	1.007×10^{-6}
25	0.99708×10^3	4.18×10^3	2.44×10^6	7.19×10^{-2}	0.612	0.0890×10^{-2}	0.897×10^{-6}
30	0.99568×10^3	4.18×10^3	2.43×10^6	7.11×10^{-2}	0.620	0.0798×10^{-2}	0.804×10^{-6}
35	0.99406×10^3	4.18×10^3	2.42×10^6	7.03×10^{-2}	0.629	0.0719×10^{-2}	0.733×10^{-6}
40	0.99225×10^3	4.18×10^3	2.41×10^6	6.95×10^{-2}	0.633	0.0653×10^{-2}	0.661×10^{-6}
45	0.99024×10^3	4.18×10^3	2.40×10^6	6.87×10^{-2}	0.641	0.0596×10^{-2}	0.609×10^{-6}
50	0.98807×10^3	4.186×10^3	2.38×10^6	6.79×10^{-2}	0.645	0.0547×10^{-2}	0.556×10^{-6}

3.2. Saturated and Unsaturated Soil Water Flow

Liquid water in a soil tends to move from a location where the potential energy is high to a location where the potential energy is low. The potential energy of water in a soil is termed total potential and is composed mainly of matric potential, gravitational potential, and osmotic potential (Miyazaki, 2006).

Hillel (1998) and Dani and Wraith (2002) are briefly discussed that soil water is subject to several force fields whose combined effects result in a deviation in potential energy relative to the reference state called the total potential of soil water (Ψ_T)

$$\Psi_T = \psi_m + \psi_s + \psi_p + \psi_z \text{.............} 5$$

where ψ_m is the matric potential resulting from the combined effects of capillarity and adsorptive forces within the soil matrix.

ψ_s is the solute or osmotic potential determined by the presence of solutes in soil water, which lower its potential energy and its vapor pressure, but osmotic potential is considered as insignificant or almost zero in non-saline soils.

ψ_p is pressure potential defined as the hydrostatic pressure exerted by unsupported water (i.e. saturating the soil) overlying a point of interest.

ψ_z is gravitational potential which is determined solely by the elevation of a point relative to some arbitrary reference point

The difference in chemical and mechanical potentials between soil water and pure water at the same temperature is known as the soil water potential (Ψ_w):

$$\Psi_w = \psi_m + \psi_s + \psi_p \quad \ldots\ldots\ldots\ldots\ldots\ldots 6$$

Note that the gravitational component (ψ_z) is absent in this definition. Soil water potential is thus the result of inherent properties of soil water itself, and of its physical and chemical interactions with its surroundings, whereas the total potential includes the effects of gravity (an external and ubiquitous force field).

The potential energy, which is due to position or internal condition, is of primary importance in determining the state and movement of water in the soil. The potential energy per unit mass of water in the soil varies over a very wide range. Differences in potential energy of water between one point and another give rise to the tendency of water to flow within the soil. In the soil, water moves constantly in the direction of decreasing potential energy. The gradient of potential energy with distance is in fact the moving force causing flow.

3.3. Saturated Water Flow

Understanding movement of water in saturated soil is important in drainage and groundwater studies (Kirkham, 2005).

The water content in a saturated soil system does not change during flow and only positive potentials are the driving force during the water transport (Lal and Shukla, 2004).

$$Q = \frac{V}{t} \text{ or } Q\alpha \ \frac{A(Hi-Ho)}{L} \ldots\ldots\ldots 7$$

$$Q = Ks \ \frac{A(Hi-Ho)}{L} \ldots\ldots\ldots\ldots\ldots 8$$

$$\frac{Q}{A} = Ks \frac{\Delta H}{L} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 9$$

$$q = Ks \frac{\Delta H}{L} \ \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 10$$

where Q is the volumetric flow rate across the column (L^3T^{-1}), volume of water flowing through the soil matrix is *V* (L^3), t is time (T), *A* is the cross-sectional area of flow(L^2), H_i and H_o are the hydraulic head maintained at inlet and outlet of the soil matrix, *L* is the length of flow or soil matrix, ΔH is the hydraulic gradient, *q* is the flow per unit cross sectional area per unit time (LT^{-1}), and is called flux density. The proportionality constant *(K_s)* in Equations (8) to (10) is known as "saturated hydraulic conductivity" of the soil matrix, which has the dimensions of velocity (LT^{-1}).

Henri Philibert Gaspard Darcy, a French hydrologist, described the relationship between the flux density and hydraulic gradient in 1856. The classical Eq. (10) which is the backbone of many steady saturated flow descriptions to date is known as Darcy's law.

3.4. Saturated Hydraulic Conductivity

The saturated hydraulic conductivity *(Ks)* of a porous medium, such as soil, refers to its ability to conduct water when all pores are full of water. It is a compound parameter, which comprises properties of the medium and water at the specified temperature and pressure (Lal and Shukla, 2004).

$$Ks = \frac{VL}{tA\Delta H} \ldots\ldots\ldots\ldots\ldots 11$$

where A is the cross-sectional area of flow through the soil column (L^2), ΔH is the hydraulic head difference as defined in Eq. (6), L is the length of column, and V is volume of water (L^3) flowing across the column in time t.

3.5. Flow of water in unsaturated soil

Unsaturated flow of water is a more commonly prevailing condition in the field than saturated flow. An unsaturated soil zone, or vadose zone, provides a continuum of waterunsaturated subsurface porous media connecting the soil/atmospheric interface and underlying saturated groundwater zone (Lal and Shukla, 2004).

A soil matrix is considered unsaturated, also called the vadose zone, when some of the pores are filled with water, the remaining pores with air and pressure heads are generally negative ($h < 0$). The unsaturated zone of soil refers to that portion of the subsurface above the water table, which contains both air and water in the pores (Dani and Wraith, 2002; Lal and Shukla, 2004)

Hillel (1998) states that most of the processes involving soil-water interactions in the field occur while the soil is in an unsaturated condition.
Flux density, as in the case of saturated flow, is proportional to the driving force (i.e., the hydraulic gradient, $\Delta H/L$). .

$$q = K(\theta)\frac{\Delta H}{L} \qquad \ldots\ldots 12$$

Where $K(\theta)$ is the unsaturated hydraulic conductivity of the porous medium (LT^{-1})

The unsaturated hydraulic conductivity, $K(\theta)$, is dependent on both moisture content and matric potential. Equation (7) can be written in terms of suction (matric potential (ψ_m), or its negative suction head) and gravitational component (ψ_z) as

$$H = \psi_m + \psi_z \qquad \ldots\ldots 13$$

$$q = K(\theta)\frac{\partial(\psi m + \psi z)}{L} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 14$$

$$q = K(\psi m)\frac{\partial H}{L} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 15$$

where H is hydraulic head

Unsaturated flow often entails changes in the state and content of soil water, involving complex relations among such variables as soil wetness (water content), suction, and conductivity, whose interrelations are further complicated by hysteresis (Hillel, 1998).

As a stream of water is passed through the unsaturated soil matrix, the incoming water replaces the air present in the soil pores; it increases the total volume of water inside the soil, thus increasing the moisture content *(θ)* of soil. This agrees with the fundamentals of continuity equation, which states that the difference in the inflow and outflow rate is equal to the change of water storage in soil. The gradient causing flow in unsaturated soils is of negative pressure potential. The flow paths in unsaturated flow are more tortuous as several pores are filled with air. A typical situation of water in an unsaturated soil media forming separate and discontinuous pockets of water the empty pores must be circumvented for the water to flow through the soil matrix. This increases the length of flow path or tortuosity. Since bigger pores drain quickly most of the air-filled pores are more conductive, forcing the water to move through only smaller less conductive pores. The same is also true in aggregated soils, where large inter aggregate spaces empty early leaving the small pores for water flow.

$$\frac{\partial\theta}{\partial t} = -\nabla q \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 16$$

where $\partial\theta$ is the change in water content in a unit volume of soil (L^3), ∂t is change in time (T), $-\nabla q$ is the spatial gradient of flux (LT^{-1}), ∇ is the vector differential operator for three-dimensional gradient.

From the combination of Darcy-Buckingham (Eq. 9) and Continuity (Eq. 11) Richards Equation is derived for unsaturated flow as follows:

$$\frac{\partial\theta}{\partial t} = \nabla[K(\psi m)\nabla H] \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 17$$

where, $\nabla H = \psi m + \psi z$

$$C(\psi m)\frac{\partial \psi m}{\partial t} = \frac{\partial}{\partial L}(K(\psi m)\frac{\partial \psi m}{\partial L} + \frac{\partial K(\psi m)}{\partial L} \ldots\ldots\ldots\ldots\ldots\ldots 18$$

where C is soil-moisture capacity function (L^{-1}), which is equal to the inverse slope of the soil-moisture characteristic curve or $\psi_m(\theta)$, ∂L is the change in lenght .

Hillel (1998) stated that apart from the gravitational force (which is completely independent of soil wetness), the flow of water in the soil is driven by a hydraulic potential gradient. Moreover; the flow rate is affected by the geometric properties of the pore channels through which water moves. These principles apply in unsaturated, as well as saturated, soils. However, the nature of the moving force and the effective geometry of the conducting pores may be very different. Lal and Shukla (2004) also put as in saturated flow ψ_p is a function of L only, however, in unsaturated flow ψ_m is a function of both L and time (t).

In unsaturated flow water will be drawn from a zone where the hydration envelopes surrounding the particles are thicker to where they are thinner, and from a zone where the capillary menisci are less curved to where they are more strongly curved Perhaps the most important difference between unsaturated and saturated flow lies in the hydraulic conductivity. When the soil is saturated, all of the pores are water-filled and conducting. The water phase is then continuous and the conductivity is maximal. When the soil desaturates, some of the pores become air filled so that the conductive portion of the soil's cross-sectional area diminishes. Furthermore, as suction develops, the first pores to empty are the largest ones, which are the most conductive, thus relegating flow to the smaller pores. At the same time, the large empty pores must be circumvented, so that, with progressive desaturation, tortuosity increases. The presence of organic surfactants that adsorb to these surfaces is considered to increase their rigidity or viscosity (Hillel, 1998).

For all of these reasons, the transition from saturation to unsaturation generally entails a steep drop in hydraulic conductivity, which may decrease by several orders of magnitude (sometimes down to 1/1,000,000 of its value at saturation) as suction increases from 0 to 1 MPa. At still higher suctions, or lower wetness values, the conductivity may be so low that very steep suction gradients, or very long times, are required for any appreciable flow to occur at all. As high suctions occur, there may also be a change in the viscosity of the (mainly adsorbed) water, tending to further reduce the conductivity (Hillel, 1998).

Table.3.2. Summary of Differences between Saturated and Unsaturated Flow (from: Lal and Shukla, 2004)

Parameter	Saturated flow	Unsaturated flow
Water content	Constant	Variable over space and time
Air content	Zero (close to zero)	Variable over space and time
Potential gradient	Positive and constant	Negative and variable
Hydraulic conductivity	Maximum, constant	Low and variable
Vapor flow	None	Possible provided temperature gradients also exist
Water flow	Steady	Steady as well as unsteady
Flow paths	Continuous	Tortuous
Continuity equation	Inflow = outflow	Inflow = outflow + source or sink, of water
Flow descriptions	Darcy's law	Darcy–Buckingham equation Richards equation
Flow parameter	K_s	$K(\theta)$

3.6. Unsaturated Hydraulic Conductivity, soil structure and texture

The conductive properties of unsaturated soils depend greatly on their texture and structure (Hillel, 1998).

A comparative study of the hydraulic conductivity as function of volumetric water content K(Θ) for disturbed and undisturbed soil samples carried out by Abrisqueta et al. (2006) showed a clear differences between the $K(\theta)$ functions for disturbed and undisturbed soil samples, pointing to the importance of soil structure in the unsaturated flow of water. For soil moisture values close to saturation (θ=30%), hydraulic conductivity was (0.392 cm h^{-1}) for undisturbed soil, whereas it was lower (0.019 cm h^{-1}) in disturbed soil.

According to Kelishadi, et al. (2013) the averages of soil hydraulic properties including macroscopic capillary length (λc) and unsaturated/saturated hydraulic conductivity values were not significantly affected by soil textural class. The study was carried out on 48 silt clay loam, 38 silt clay and 10 silt loam soil samples. However, 40, 31, 15 and 10 soil samples were taken in pasture land, dry-land farming, irrigation farming and in fallow land uses and the average soil hydraulic properties were mainly varied with land uses significantly independent of their soil texture.

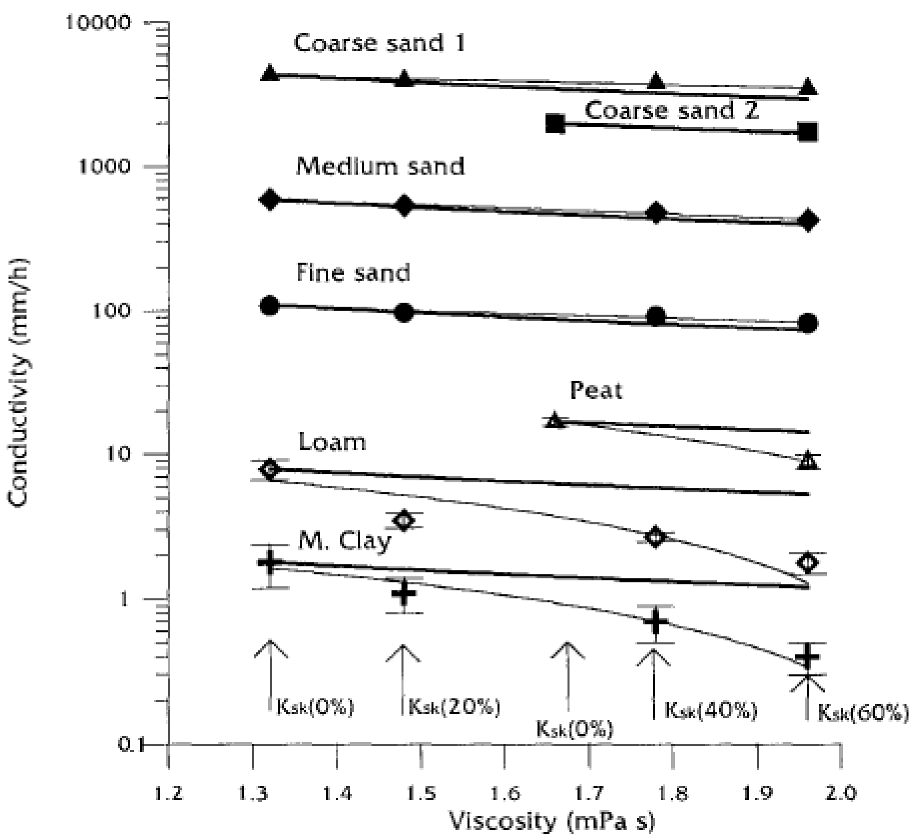

Fig.3.1. **Observed (symbols) and predicted (thick continuous lines) conductivities for kerosene mixtures with different viscosities and chemical composition (from: Jarsj et al., 1996)**

The effect of soil texture is not clear like the effect of soil structure on fluid conductivity as the temperature and viscosity change. For example, in the study of Jarsj et al. (1996) which deals with temperature-induced changes in kerosene viscosity, the resulting saturated conductivity ratios were consistent with the corresponding viscosity ratios. For volatilization induced viscosity changes, which reflect irreversible changes in kerosene chemical composition, showed relatively small deviations from corresponding viscosity ratios for inert, sandy soils (coarse, medium and fine sand). However, for volatilization-induced viscosity changes in soils containing clay or organic matter the observed saturated conductivity ratios deviated significantly. A comparison between four different kerosene mixtures in two montmorillonitic soils furthermore showed that the discrepancies between these two ratios increased with increasing differences in liquid chemical composition. Thus, the results indicate that other properties than the fluid viscosity may be of importance for conductivity of potentially interacting soils with respect to volatile organic liquid mixture residues.

3.7. Hydraulic conductivity and Temperature

According to Zhang et al. (1993) it is evident that the relation between the swelling pressure, Π and the interlayer spacing, λ is essentially independent of temperature in the temperature range of the experiments. Changes in this relation from one temperature to another were within experimental error. The effect of temperature on the relation between Π and ratio of mass of water to mass of clay (m_w/m_c) for the Na-montmorillonite is that Π decreased with temperature, T at any value of mw/mc, i.e., ($\partial \Pi$ / ∂T) was negative, and that m_w/m_c decreased with increasing T at any value of Π.

where λ is the interlayer spacing (L), $\partial \Pi$ is the change in swelling pressure ($ML^{-1}T^{-2}$), ∂T is the change in temperature (^oC).

Romero et al. (2001) states that Retention curves at different temperatures show reduction of total suction with increasing temperatures at constant water content. Temperature influence on water permeability is more relevant at low matric suctions corresponding to bulk water preponderance (inter-aggregate zone). Below a degree of saturation of 75% no clear effect is detected (see fig. 3.3). Experimental data show that temperature dependence on permeability at constant degree of saturation and constant void ratio is smaller than what could be expected from the thermal change in water viscosity (it will be discussed in detail in chapter IV). Thermo-chemical effects altering clay fabric (flocculation or dispersion), porosity redistribution (creating preferential pathways or blocking macropores) and pore fluid chemistry (affecting viscosity) could be relevant. Relative water permeability values at constant void ratio showed no temperature dependence.

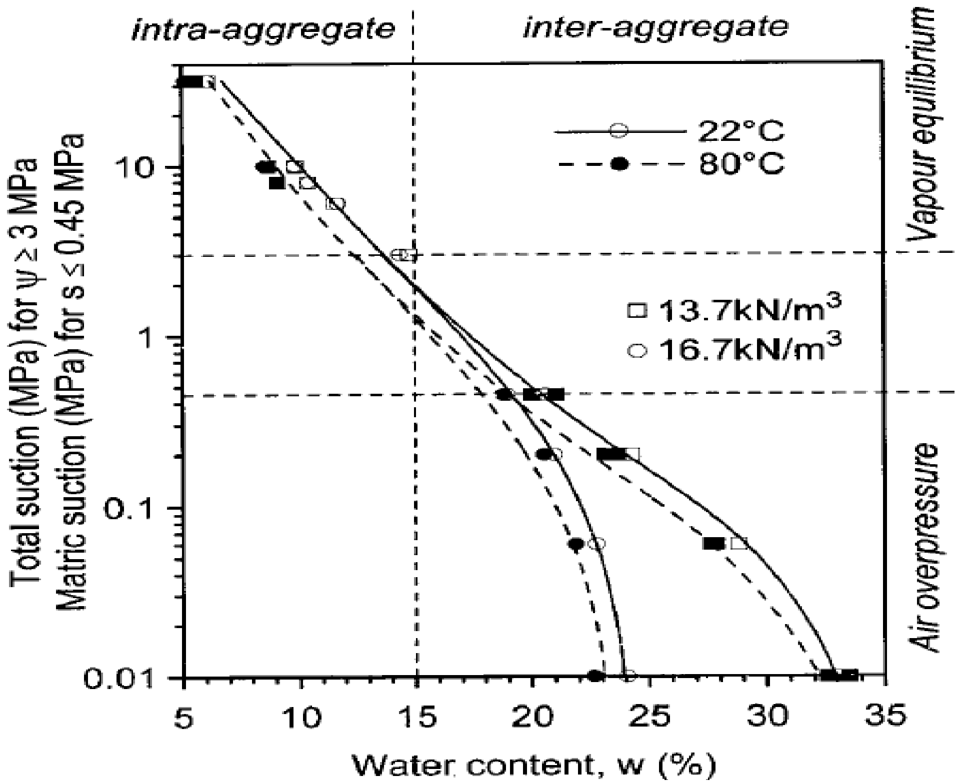

Fig.3.2. **Water Retention Curves for different Temperatures at Fixed Dry Unit Weights (from: Romero et al., 2001)**

Romero et al. (2001) mentioned that temperature effects on water permeability at saturation are usually derived from viscosity changes under free water considerations. This interpretation can be extrapolated to unsaturated states at constant void ratio e and water content w, according to the following expression:

$$\frac{kw(e,w,T)}{k(e,w,Tr)} = \frac{\rho w(T)\mu(Tr)}{\rho(Tr)\mu w(T)} \approx 1 + \beta_T(T\text{-}T_r)\ldots\ldots\ldots\ldots 19$$

where, ρ_w is the density of water, μ_w is the absolute viscosity and β_T = 0.030K^{-1} (for a reference temperature, Tr = 22^{o}C) an empirical coefficient that fits relative viscosity data over a temperature range of 22^{o}C $\leq$ T $\leq$80^{o}C. The slight increase of permeability with temperature at degree of saturation, Sr = 95 $\pm$ 5% of kw (80^{o}C) / kw (22^{o}C) = 1.3 cannot be explained in terms of a reduction of free water viscosity in the same interval of temperature.

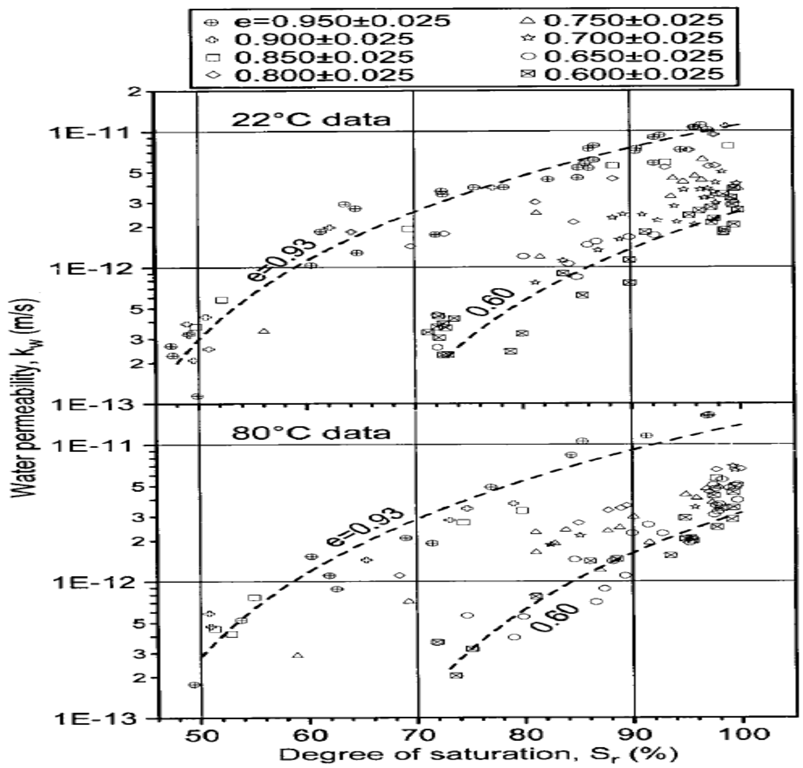

Fig.3.3. **Water Permeability vs. Degree of Saturation at 22^{o}C (upper graph) and 80^{o}C (lower graph) (from: Romero et al., 2001)**

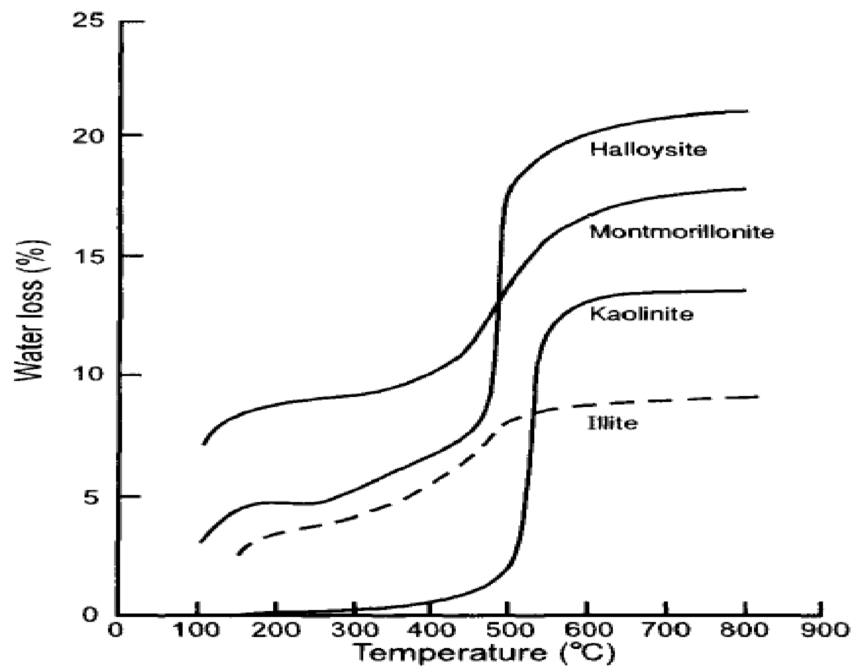

Fig.3.4. **Dehydration Curves of Clay Minerals (After Marshall, 1964) (from: Hillel, 1998)**

On the other hand the hydraulic conductivities of Ca-bentonites with the dry bulk densities of 1.4Mg/m^3, 1.6 Mg/m^3 and 1.8 Mg/m^3 as a function of temperature are presented, hence, the change in viscosity of water with increasing temperature contributes greatly to the increase of hydraulic conductivity, however, the intrinsic permeability is nearly constant. The hydraulic conductivities increase with increasing temperature. The hydraulic conductivities for bentonites with dry bulk densities of 1.4 Mg/m^3 to 1.8 Mg/m^3 at the temperature of 80^{o}C increase up to about three times as those at 20^{o}C (Cho et al., 1999).

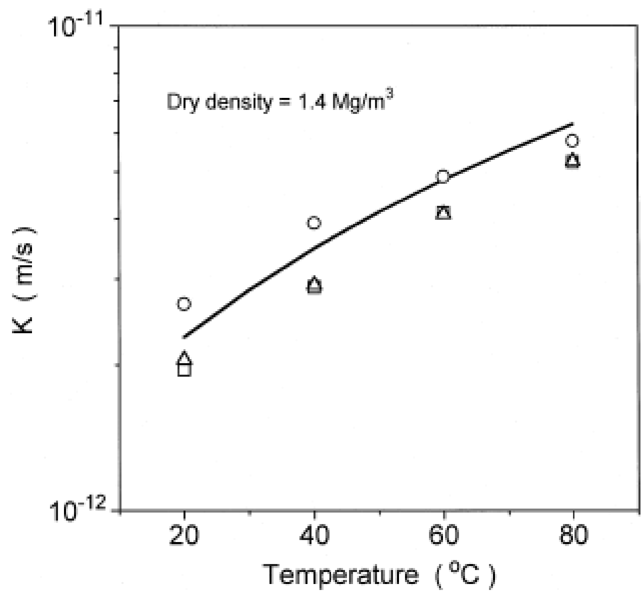

Fig.3.5 **Hydraulic Conductivity of Compacted Bentonite with dry bulk density of 1.4 Mg/m3 at different temperature (*O* , ▭, Δ measured over three different samples -: predicted) (from: W.J. Cho, J.O. Lee and K.S. Chun., 1999).**

Nimmo and Miller (1986) state that model calculations for rapid, transient measurement of hysteretic soil-moisture characteristics as a function of temperature on sand and silt loam soil data indicate that except near saturation, the temperature effect is greater than can be accounted for by the temperature dependence of the surface tension of pure water. The soil water content and water potential curve showed that as the temperature decreases from 50 to 4^{o}C the lower end hysteresis cycle i.e. $\psi(\theta)$ is getting higher. Thus, the results rule out several possible explanations but they support the hypothesis that the concentration and effectiveness of dissolved surfactants increases with temperature.

Eva (1994) explained by providing many scientific articles that temperature has a significant effect on changes in the capillary pressure-saturation relation. Changes in the capillary pressure-saturation relation with temperature are due to changes in the interfacial tension of the fluids with temperature. Contact angle of the system, the drainage and imbibitions are affected differently by changes in temperature and fluid pairs. Different fluid pairs in a given medium, the ratio of their capillary pressures is the same as the ratio of the product of interfacial tension and cosine of the contact angle, which is called the adhesion tension. The mathematical expression can be:

$$\frac{Pc1}{\gamma 1 cos\varepsilon 1} = \frac{Pc2}{\gamma 2 cos\varepsilon 2} \text{...............} 20$$

where P_c is the capillary pressure; γ is the interfacial tension between the fluids; ε 1&2 is the contact angle between the fluid interface and the solid surface where the subscripts 1 and 2 are used to specify the two different fluid pairs.

Philip and de Vries (1957) quantitative scaling of matric potential (ψ_{m}) in proportion to surface tension (σ) is one part of the surface-tension, viscous-flow (STVF) concept of moisture behavior in granular porous media for a given water content (θ) can be expressed as

$$\frac{(\partial\psi\mathrm{m}/\partial T)}{\psi\mathrm{m}} = \frac{(\frac{\partial\sigma}{\partial T})}{\sigma} \text{..............} 21$$

where ∂T is the change in temperature, $\partial\psi_m$ is the change in matric potential, $\partial\sigma$ is the change in surface-tension, ψ_m is matric potential, and σ is surface-tension

Water movement in porous material is a complex phenomenon and also the relationship of soil and water under the frozen conditions is very important; these conditions in some regions can affect the water movement for a significant part of the year. According to Pieter Hoekstra (1966**)** moisture flow in the frozen soil takes place under temperature gradients through the films of unfrozen water. Since the thickness of the unfrozen films decreases with temperature, the rate of water transport also decreases rapidly with decreasing temperature below 0^oC. There is also a free-energy difference in a non isothermal system. The direction of water flow can be predicted in an isothermal system by macroscopic quantities such as tension and water content. In a non isothermal system water content and tension alone are not sufficient to predict the direction in which the water will flow. The free-energy function of soil water can be expressed in terms of three macroscopic variables: water content, temperature, and tension. The limitation of reversible thermodynamics, however, is that it will not predict the rate of flow and in some cases not even the direction of flow. The theory of irreversible thermodynamics, under certain conditions, relates the gradient of a thermodynamic quantity to the rate of flow.

One example related to hydrogeology that could help to understand the relationship between hydraulic conductivity and temperature profile was studied by Su et al. (2004). They found that from six ground water wells observation the two lowest fluxes occur in two wells, which had a nearly constant temperature. The four remaining wells had much higher fluxes compared to those two previously mentioned wells, and they had distinct seasonal temperature profiles because of the increase in advective heat transport. This study paper also will discuss about different temperatures and hydraulic conductivity relations.

Another related example is, due to climate change, hydrological and thermal regimes of rivers are expected to change. However, most of the studies do not deal with changes in water temperature (or water quality in general) (Vliet et al., 2013). Vliet et al. (2013) result show that an increase in the seasonality of river discharge (both increase in high flow and decrease in low flow) for about one-third of the global land surface area for 2071–2100 relative to 1971–2000. Global mean and high (95^{th} percentile) river water temperatures are projected to increase on average by 0.8–1.6 (1.0–2.2) ^{o}C for the (Special Report on Emissions Scenarios) SRES B1–A2 scenario for 2071–2100 relative to 1971–2000.

3.8. Methods of unsaturated soil hydraulic conductivity measurement

Solution of unsaturated problems generally requires experimental determination of the relationship between hydraulic conductivity and water potential or water content. Field methods used to obtain these relationships include the instantaneous profile method, steady flux methods, sorptivity measurements and use of tension infiltrometers. Sample numbers and the extensiveness of a site characterization can be limited, because instantaneous profile and steady-flux techniques require laborious installation or a number of neutron probe access tubes to make hydraulic conductivity measurement (Ankeny, 1991).

Different methods of unsaturated soil water hydraulic conductivity measurement developed by different researchers are presented below (cited and summarized from Marino et al., 2008 and Hillel, 1998):

The measurement methods can be divided in to direct and indirect methods or Field and Laboratory methods.

3.8.1. Direct Methods

3.8.1.1. Steady state methods

In the steady state methods, a constant flow rate or hydraulic gradient is applied under specified average water pressure head (h_{pw}). Steady-state is supposed to occur as soon as the flow rate of the soil sample is equal upstream and downstream and/or if a constant hydraulic gradient is observed through the tested soil sample. The experiment can be repeated for different magnitudes of pressure or water content to yield a $K(\theta)$-h_{pw} relationship. This laboratory method is described by Klute and Dirksen (1986).

The steady state method may be costly, tedious and lengthy in low permeability materials. In order to minimize the long test time required on the traditional steady state methods, the steady state centrifuge method has been used to measure unsaturated water hydraulic conductivity on various porous samples.

3.8.1.2. Unsteady state methods

The unsteady state methods are usually divided in to two primary groups: outflow-inflow methods and instantaneous profile methods.

1. A widely used unsteady state flow method for measuring unsaturated hydraulic conductivity in the laboratory is the outflow method which was described by Gardner (1956). The unsaturated hydraulic conductivity is constant, and relation between the water content and matric suction is linear. The primary advance of these methods is easy to perform.
2. The instantaneous profile method (Watson, 1966) consists of inducing transient flow in a long cylindrical sample of soil and then measuring the resulting water content

and/or pore water pressure profiles at various time intervals. For example the horizontal infiltration technique of (Bruce and Klute, 1956).

3. Direct evaporation technique provides estimates of retention and conductivity curves in the tensiometric range and is quite easy to use and has been intensively studied in recent years. This method is a laboratory method which requires 3-4 weeks per test, not including calibration of psychrometers and or tensiometers, fore-instance (Moore, 1939) laboratory method. The pore water pressure can be measured using tensiometers and /or thermocouple psychrometers and the water content profiles using Time Domain Reflectometry or gamma ray attenuation.

From these considerations, it seems unrealistic to measure the unsaturated hydraulic conductivity of field soil by making laboratory determinations on discrete samples removed from their natural continuum. Such samples are generally dried, fragmented, and repacked into experimental containers so that the original structure is destroyed. Hence it is necessary to devise and test practical methods for measuring soil hydraulic conductivity on a realistic scale in situ (Hillel, 1998) like sprinkling infiltration, infiltration through an impeding layer, internal drainage, sorptivity method, Guelph permeameter and inverse methods.

3.8.2. Indirect methods

Because of experimental difficulties for direct measurement of the hydraulic conductivity of the unsaturated deformable porous medium, it is often estimated using semi-empirical models based on the saturated hydraulic conductivities and SWRC parameters. These Semi-empirical methods are discussed by Brooks and Corey (1964), Mualem (1976), Van Genuchten (1980), Zhang, R. (1997), etc.

For this study Mualem (1976) and Van Genuchten (1980) equations will be discussed to elaborate important Van Genuchten parameters along with direct field and laboratory measurement data and derived equations.

3.9. Equations of unsaturated soil hydraulic conductivity

Van Genuchten (1980) states that the relative conductivity can also be expressed in terms of the pressure head by deriving the equation from the equation of (Mualem, 1976) for predicting relative hydraulic conductivity from the knowledge of soil water retention curve

$$K_r(h) = \Theta^{1/2}\left[\int_0^{\Theta} \frac{\frac{1}{h(x)}dx}{\int_0^1 \frac{1}{h(x)}} dx\right]^2 \ldots\ldots.22$$

Where h is the pressure head, a function of the dimensionless water content or degree of saturation:

$$\Theta = \frac{\theta - \theta r}{\theta s - \theta r} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots23$$

In this equation, s and r indicate saturated and residual values of the soil water content (θ), respectively. To solve Eq. 17, an expression relating the dimensionless water content to the pressure head is needed. An attractive class of Θ(h)-functions, adopted in this study, is given by the following general equation

$$\Theta = \left[\frac{1}{1+(\alpha h)^n}\right]^m \ldots\ldots\ldots\ldots\ldots\ldots..24$$

Where α, *n* and *m* are as yet undetermined parameters. To simplify notation later, h in Eq. 24 is assumed to be positive.

$$K_r(h) = \frac{〚(1-(\alpha h)〛^{(n-1)}\ 〚[1+(\alpha h)〛^{n}]\ 〚-m)〛^{2}}{[1+(\alpha h)^n]^{m}/2} \ \ldots\ldots\ldots25$$

m=1-1/n

Nakhaei and Šimůnek (2014) mentioned that Philip and de Vries (1957) derived the equation to predict soil water pressure head quantitatively from the influence of temperature on surface tension based on capillary theory

$$\frac{\partial h}{\partial T} = \frac{h}{\sigma}\frac{\partial \sigma}{\partial T} \ldots\ldots\ldots\ldots26$$

where T is temperature (K) and σ is the surface tension at the air-water interface (MT^{-2}).

$$hT = \frac{\sigma T}{\sigma ref} href = \alpha h^{*}\, href \ldots\ldots\ldots\ldots 27$$

where hT and $href$ (σT and σref) are pressure heads (surface tensions) at temperature T and reference temperature $Tref$, respectively; and αh^* is the temperature scaling factor for the pressure head. Nakhaei and Šimůnek (2014) cited from Constantz (1982), the temperature dependence of the hydraulic conductivity can be expressed as follows:

$$K(\theta)T = \frac{\mu ref\, \rho T}{\mu T\, \rho ref} k_{ref}(\theta) = \alpha_k K_{ref}(\theta) \ldots\ldots\ldots\ldots\ldots\ldots 28$$

where $Kref$ and KT denote hydraulic conductivities at the reference temperature $Tref$ and soil temperature T, respectively; μref and μT (ρref and ρT) represent the dynamic viscosity ($ML^{-1}T^{-1}$) (density of soil water (ML^{-3})) at temperatures $Tref$ and T, respectively; and αK^* is the temperature scaling factor for the hydraulic conductivity. Nakhaei and Šimůnek (2014) concluded that to estimate or predict selected soil hydraulic and thermal parameters is quite possible using HYDRUS-2D model, but the uniqueness of the optimized values needs further study.

3.10. Tension infiltrometer methods

According to Buttafuoco and Ricca (2013) measurements were made at the surface of ploughed soil and four supply water potentials were used to measure infiltration rates (-15, -10, -5, 0 cm) at the first site and three supply water potentials (-15, -10, -5 cm) at the second site. Hence, they concluded that in-situ tension infiltration measurements have allowed soil hydraulic characterization at two sites with a field method description. Tension disk infiltrometry showed its suitability and simplicity to characterize hydraulic properties in field soils in the particular flow region of saturation and close to saturation. The tension infiltrometer is an instrument used both in the laboratory and the field to characterize water entry into the soil under a range of negative water pressure heads at large negative pressure heads, water flows only through very small pores; as the pressure head is increased towards zero pressure head, larger pores participate to flow.

Li1 et al., (2005) stated that very few methods had been developed for the measurement of infiltration in crusts, which are often distributed on steep slopes where experimental devices are very difficult to install, making measurement difficult. Thus, they had forced to use mini-disk infiltrometers (MDI) in the laboratory to measure steady infiltration rate under nearly saturated condition and tensions in soil crust samples removed from the field. Steady infiltration rate under tensions were performed at three water pressure heads (h=-0.5, -2.0, and -6.0 cm).

3.11. Representative Elementary volume

The term 'representative elementary volume' was first used by Bear (1972). The representative elementary volume can be defined as the minimum volume of a soil sample required from which a given soil parameter measurement becomes independent of the size of the sample (VandenBygaart and Protz, 1998). According to the concept note of Hillel (1998) if soil sample is very small the measured porosity may vary between zero and 100 percent. Measuring the porosity repeatedly at several adjacent points by taking small soil volume, porosity results will fluctuate widely. However, if the scale or volume of each sample increased so as to encompass within it both particles and pores, the fluctuations among repeated measurements at adjacent locations will diminish. Continuous enlarging of soil sample will eventually provide a consistent measurement of average soil porosity.

The representative elementary volume (REV) of a material with respect to a macroscopic property is the minimal volume at which it is reasonable to define this property (Brown et al., 2000).

According to Kaempfer et al. (2005) the representative elementary volume (REV) of the snow (macro-structurally) was deduced with respect to density using four cubic sub-volumes and heat flux in the ice matrix solving numerically the equations for snow volumes with different sizes and at different positions, and related it to parameters obtained by image analysis.

4. Materials and Methods

Two types of experiments were carried out using the Minidisk Infiltrometer (Decagon Devices, Inc.) in this study: Field measurement and laboratory measurement.

4.1. Field measurement

4.1.1. Area of measurement

The field measurement has been conducted at the Experimental terrain station of soil moisture dynamics of Department of Water Resources, located at agricultural experiment field area of of Czech University of Life Sciences Prague during summer in June and July 2013. Soil type of the site was Haplic Chernozem (IUSS Working Group WRB, 2006) and texture class of the soil was silty clay loam and its dry bulk density was 1.55-1.62g/m^3 and porosity was 0.37-0.41%. The study site was flat and the vegetation cover is permanent grass. The weather during field experiment was mostly hot sunny days with sometimes cloud covers and wind drift.

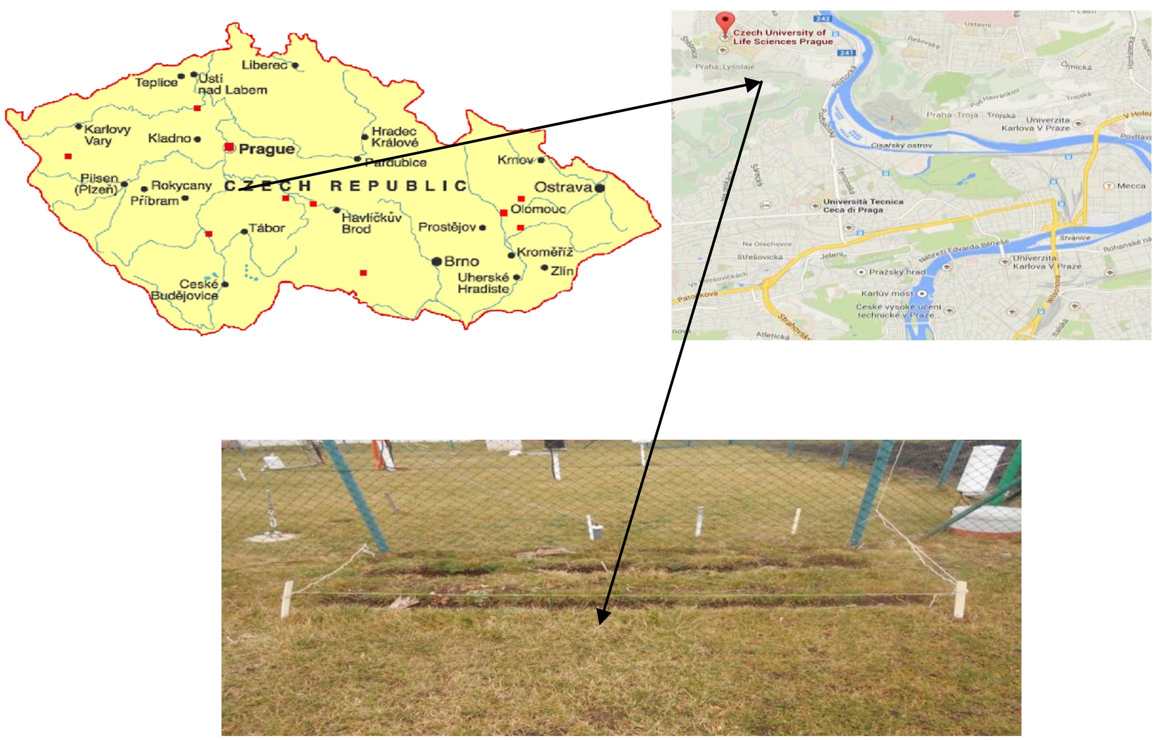

Fig.4.1 **Agriculture field experiment site: near meteorology station, Czech University of Life Sciences**

4.1.2. Field experiment design

In the study site small plot (2.5m x 1.3m) of land was selected for measurements. Inside the plot 2 m x 0.25 m area was cleared from herbs, large plant roots and fine gravels to create suitable smooth surface to place minidisk infiltrometer. . Approximatelly 5 cm of grass root zone was thus removed.

Measurements of unsaturated hydraulic conductivity taken by Minidisk Infiltrometer can be significantly affected by initial water content (i.e. study of Lufinková, 2012). This study is aimed to observe the influence of temperature, thus the other factors with possible influence should be minimised as much as possible. Initial water content during experiments was kept as similar as possible. Therefore, the prepared experimental area was irrigated and water content of the soil was observed. After redistribution (1 or 2 days) the measurements were carried out.

The mini plot was divided in to four spots (0.25 m x 0.25 m) and at a depth of 5 cm for infiltration measurement. Other important measurements were done along with infiltration such as soil water content, soil water temperature, electrical conductivity of the soil before and after water infiltration in the same spot.

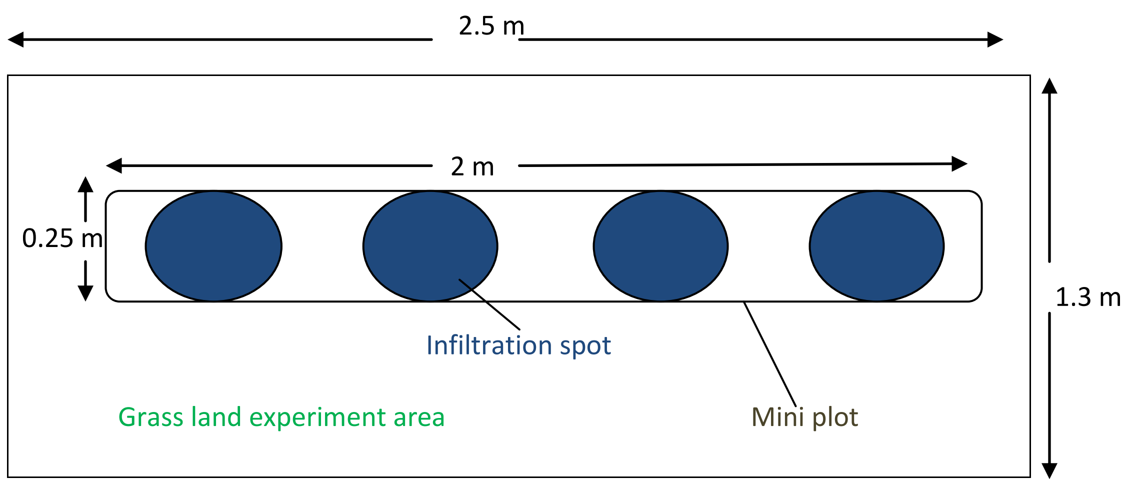

Fig. 4.2 **Schematic representation of prepared area for hydraulic conductivity measurement using minidisk infiltrometer**

During measurement direct contact between minidisk infiltrometer and soil surface was made because according to the Vandervaere et al. (2000), the water initially drained into the sand cover during the initial stages of infiltration influences the cumulative infiltration curve highly, which makes a steep drop in the first seconds of the cumulative infiltration curve.

The infiltration measurements were done at 10^{o}C, 25^{o}C and 35^{o}C water temperature. These values of temperature were chosen based on the idea that they are in the range of “normal operating temperature” of the device, research area, common research season, considering temperature variation of most agricultural climates, and day and night temperature fluctuations.

4.1.3. Field Measuring Materials

Minidisk Infiltrometer (Decagon Devices, Inc.; www.decagon.com) was used for measuring hydraulic conductivity by pouring water to the bubble chamber following to its water reservoir. The suction control tube was set to -2 cm and kept the same for all infiltration experiments. The 2 cm tension was selected according to the Mini Disk Infiltrometer operator’s manual (Decagon Devices, Inc) referred as the most commonly used and because the focus of the measurement was temperature variability and water infiltration.

Minidisk infiltrometer has small dimensions having 32.7 cm total length, 3.1 cm diameter of tube, 4.5 cm diameter and 3 mm thick of sintered stainless steel disc, 21.2 cm water reservoir length, 28 cm mariotte tube length and 10.2 cm suction control tube having 0.5 to 7 cm tension range.

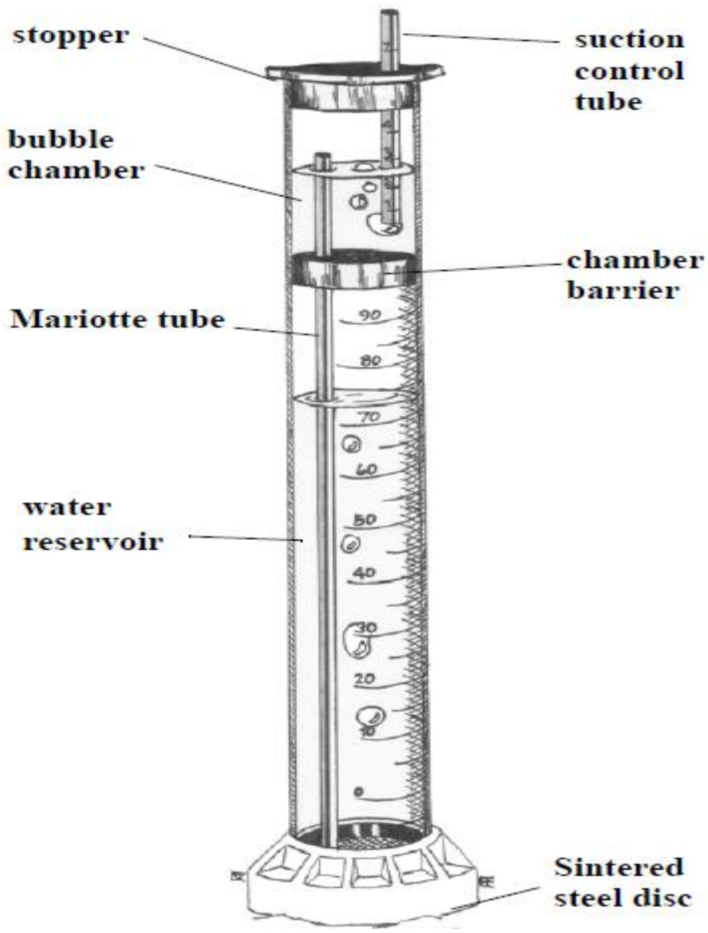

Fig.4.3 **Picture and descriptions of Minidisk Infiltrometer (from: Decagon device, Inc. www.decagon.com)**

The FDR soil moisture sensor 5TE (Decagon Devices, Inc.) was used to measure soil water content (m^3/m^3), soil temperature (^oC) and electrical conductivity (mS/cm) combined with Em50 data logger series. The FDR soil moisture sensor 5TE has an accuracy of $\pm 3\%$, for volumetric water content, $\pm 10\%$ bulk electrical conductivity and $\pm 1^oC$ temperature within the range of 0-100%, 1-23dS/m and -40-50^oC respectively.

The water temperature was adjusted using boiling thermostatted water bath and/or glass mercury thermometer. The water used for infiltration experiments was a common tap water of drinking quality.

Fig.4.4 **Pictures taken during field measurement of unsaturated hydraulic conductivity (June, 2013)**

4.2. Laboratory measurement

Whenever researchers dealing with such an essential soil characteristics like unsaturated hydraulic conductivity, in-situ measurement has particular importance to keep the hydraulic connections of the measured spot with the surrounding soil. However, too many factors influence the in-situ measurements including weather and recurrent seasonal changes, such as changes in root zone condition due to growth of vegetation and the natural soil variability from one measuring spot to another. In order to investigate the temperature influence on Minidisk Infiltrometer measurements the infiltration experiment were carried out under controlled laboratory conditions unlike field measurements to minimize the possible influence of environmental factors and to ensure the highest possible homogeneity of soil conditions.

Three different temperatures were used 10, 25 and 35 °C.

4.2.1. Materials Used in Laboratory

Artificially packed soil samples were prepared (see Fig. 4.2.1). The following equipment were used to carry out the lab infiltration experiments:

Minidisk infiltrometer for hydraulic conductivity measurement, Em50 data logger series for temperature, electrical conductivity and soil water content measurement, transparent plastic container for artificial soil profile compaction, drying oven cabinet, meter for depth measurement, stopwatch for timing, water bath to boil water and thermometer to adjust water temperature, spatula to mix soil matrix and water thoroughly, pestle and mortar for soft earth preparation, rod for soil compaction, etc

Sufficient amount of disturbed soil sample was taken from the same experimental area and the same depth where field measurement had been done. The soil sample was air dried and crushed using pestle and mortar then passed through a 2 mm sieve. 27 kg soft earth was prepared. Using the prepared soft earth six artificial soil profiles were prepared compacting to a depth of 12 cm and dry bulk density of 1.5 mg/cm^3 for each temperature hydraulic conductivity measurement in a cylindrical container. The initial water content of soil was considered and adjusted to 0.2 m^3/m^3.

The temperature of the water was similar to temperature of air during laboratory measurement. The measuring temperature of the air was not controlled at field rather it was very fluctuating due to continues shift from sunny to cloudy, from windy to stable weather while during laboratory experiments air temperature was controlled.

Apart from the change in soil structure due to grinding and artificial packing of the soil, the only difference from field measurements was controlling of air temperature which was kept the same as the temperature of water used for infiltration. Measurement was done (outside laboratory) at air temperature in winter February 2013 for 10^oC water temperature, Laboratory room temperature for 25^oC water temperature and drying oven cabinet which was set to 35^oC for 35^oC water temperature. After packing of soil at the adjusted initial water content, the soil columns were left for several hours to homogenize and stabilize either the water content or the required temperature.

Fig.4.5 **Pictures taken in laboratory weighted soils, soil with 20% water content by volume and compacted soil.**

4.3. Data collection and analysis

Primary data were collected at field and laboratory. The tension applied during measurement was always 2 cm. Cumulative infiltration and time measurements were recorded at the data record form for the Minidisk infiltrometer (Báťková et al., 2012).

Cumulative infiltration to time function was fitted according to Philip (1957) as proposed in Zhang (1997).

$$I = C_1 t + C_2\sqrt{t} \qquad \text{…………29}$$

where C_1 and C_2 are parameters related to hydraulic conductivity and soil sorptivity respectively, I is the cumulative infiltration and t is the time.

Then the unsaturated soil hydraulic conductivity was calculated at the applied tension of 2 cm using the equations below:

$$K(h) = \frac{C1}{A} \text{..............} 30$$

where C_1 is the slope of the curve of the cumulative infiltration and of squared root of time, and A is a value relating to van Genuchten (1980) parameters for the 12 soil texture classes to the radius of the disk and applied tension. The A parameters for silty clay loam were obtained from the table of van Genuchten table below:

Table 4.1. Van Genuchten parameters for 12 soil texture classes and A values for a 2.25cm disk radius and tension values from 0.5 to 7cm (from Decagon device, Inc., 2012).

	radius	**2.25**	**A**							
Texture class	**Alpha(α)**	**n / ho**	**-0.5**	**-1**	**-2**	**-3**	**-4**	**-5**	**-6**	**-7**
sand	0.145	2.68	2.835701	2.40407	1.727908	1.241921	0.892621	0.641565	0.46112	0.331427
loamy sand	0.124	2.28	2.9853	2.786831	2.4286	2.116417	1.844363	1.60728	1.400674	1.220625
sandy loam	0.075	1.89	3.877062	3.887982	3.909913	3.931969	3.954148	3.976453	3.998884	4.021441
loam	0.036	1.56	5.461148	5.717657	6.267384	6.869965	7.530482	8.254505	9.048139	9.918077
silt	0.016	1.37	7.921451	8.177401	8.714378	9.286617	9.896433	10.54629	11.23883	11.97683
silt loam	0.02	1.41	7.102076	7.367933	7.929874	8.534674	9.185601	9.886173	10.64018	11.45169
sandy clay loam	0.059	1.48	3.210664	3.523317	4.242925	5.109507	6.153081	7.409796	8.923184	10.74567
clay loam	0.019	1.31	5.857535	6.10902	6.644845	7.227667	7.861609	8.551155	9.301181	10.11699
silty clay loam	0.01	1.23	7.893227	8.094056	8.511175	8.949789	9.411007	9.895994	10.40597	10.94223
sandy clay	0.027	1.23	3.336287	3.570465	4.089288	4.683501	5.364059	6.143508	7.036218	8.058649
silty clay	0.005	1.09	6.076318	6.169307	6.359575	6.55571	6.757895	6.966316	7.181164	7.402639
clay	0.008	1.09	3.998056	4.096399	4.300401	4.514562	4.739389	4.975412	5.223189	5.483306

RGUI, R version 3.0.2 software (a command which provides basic statistics graphical user interface to analyze experimental data) and Microsoft Office Excel 2007 were used to calculate and analyze data.

4.3 Intrinsic permeability relation method

According to Jarsj et al. (1996) cited from many scientific articles conductivity relations for the water and volatile organic liquid mixture phases are commonly assumed to be related to an intrinsic permeability that is a characteristics property of the porous medium and is independent of the fluid flowing through this medium. Thus, the unsaturated fluid conductivity is then assumed to be related to the intrinsic permeability.

The hydraulic conductivity of the soil was recalculated to show how the change in viscosity and density with temperature affected the result of unsaturated soil hydraulic conductivity measurement, however, the recalculated result could not show the effect of salt in the working water, organic surfactants in the soil, compactness variation among measurement spots, etc. The recalculation was made by considering the most sensitive physical characteristics of water by temperature change such as viscosity and density, and by considering the intrinsic permeability of the soil (which is the same soil matrix) in this paper.

$$K(h)refT = k\rho g/\mu \text{..............31}$$

where K(h)refT is the unsaturated hydraulic conductivity of the porous medium(soil) with respect to the specific fluid at reference temperature(LT^{-1}), k is the intrinsic permeability of the porous medium(soil) (L^{-2}), ρ and μ are the fluid (water) density (ML^{-3}) and dynamic viscosity ($ML^{-1}T^{-1}$),, respectively, and g is the gravitational constant(LT^{-2}).

The soil type in this experiment is one, so that here we can drive a formula to calculate the relative change of hydraulic conductivity due to the change in viscosity and density of the water with temperature. In this experiment room temperature was taken as a reference water temperature i.e. 25^{o}C.

$$K(h)refT = \frac{k\rho g}{\mu}\,,\ K'(h)T = \frac{k\rho\prime g}{\mu\prime}\text{.............32}$$

where K’(h)T is the unsaturated hydraulic conductivity of the porous medium(soil) with respect to the specific fluid at 10 and 35^{o}C water temperature, k is the intrinsic permeability of the porous medium(soil), ρ and μ are the fluid (water) density and viscosity at10 and 35^{o}C water temperature , respectively, and g is the gravitational constant.

$$k = K(h)refT * \frac{\mu}{\rho g} = K'(h)T * \frac{\mu'}{\rho' g} \ldots\ldots\ldots\ldots\ldots 33$$

from equation (34) we can get a simple mathematical relation to calculate the anticipated value of unsaturated hydraulic measurement at 10 and 35^{o}C from the value of reference temperature.

$$K'(h)T = K(h)refT * \frac{\mu\rho'}{\mu'\rho} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 34$$

5. Results and Discussions

5.1. Results and discussion

Unsaturated hydraulic conductivities of a soil were measured using minidisk infiltrometer at three different temperatures at field and in the laboratory. The Field measurements were including 10^{o}C, 25^{o}C and 35^{o}C water temperature variables and laboratory measurements were done at similar water temperature of field measurements. Laboratory measurements of soil hydraulic conductivity were also done at controlled temperature; the temperatures of measuring environment were the same as the temperatures of measuring water.

Table 5.1. Laboratory experiment temperature (T_w), measured values of initial and final water infiltration volume in minidisk infiltrometer reservoir (I_i and I_f), soil moisture content (θ_i and θ_f), electrical conductivity (EC_i and EC_f), soil temperature (T_i and T_f), prepared soil profile depth and unsaturated hydraulic conductivity of at 10, 25 and 35^{o}C temperature

T_w (^{o}C)	I_i (ml)	I_f (ml)	θ_i (m^3/m^3)	θ_f (m^3/m^3)	EC_i (mS/cm)	EC_f (mS/cm)	T_i (^{o}C)	T_f (^{o}C)	D (cm)	K(h) cm/s
8	82	65	0.221	0.25	0.6	0.47	11.9	6.5	10	1.15E-05
	87	74	0.208	0.215	0.41	0.48	12.6	6.8	10	2.30E-05
	90	77	0.248	0.25	0.57	0.63	11.9	8	10	1.15E-05
	88	33	0.227	0.27	0.64	0.45	8.9	8.3	10	-
	84	65	0.224	0.244	0.59	0.51	12.1	8.4	10	1.15E-05
	87	71	0.22	0.239	0.38	0.47	12.2	8.7	10	2.30E-05
25	81	50	0.242	0.285	0.56	0.6	24.5	24	10	5.13E-05
	86	50	0.241	0.281	0.57	0.58	23.9	23.9	10	4.59E-05
	85	49	0.249	0.307	0.59	0.56	23.5	24.6	10	4.59E-05
	85	47	0.208	0.283	0.57	0.62	24.4	24.1	10	4.59E-05
	83	49	0.243	0.295	0.53	0.56	24.3	24.3	10	5.74E-05
	91	44	0.261	0.319	0.56	0.64	24.1	25.8	10	5.74E-05
35	88	57	0.198	0.316	0.66	0.75	26.3	27.3	10	2.30E-05
	87	55	0.204	0.235	0.54	0.59	26	26.8	10	2.30E-05
	88	52	0.207	0.318	0.53	0.67	27.2	27.1	10	4.59E-05
	89	55	0.21	0.291	0.56	0.65	26.4	27.1	10	2.30E-05
	90	54	0.21	0.294	0.6	0.71	26.3	27.2	10	4.59E-05
	89	53	0.201	0.272	0.55	0.65	26.4	27.1	10	2.30E-05

Table 5.1 shows the direct measured parameters and calculated unsaturated hydraulic conductivity at 2 cm tension. The value of K(-2) is ranging from 1.15-2.3E-05 cm/s, 4.59-5.74E-05 cm/s, 2.3-4.56E-05 cm/s at a temperature of 8^{o}C, 25^{o}C and 35^{o}C respectively. The θ_i and θ_f(m^3/m^3) , EC_i (mS/cm) and EC_f (mS/cm), and T_i(^{o}C) and T_f(^{o}C) were continuously recorded and their influence on hydraulic conductivity will be evaluated below.

Table 5.2. Laboratory measured mean values of infiltration volume, soil moisture content, electrical conductivity, soil temperature, soil depth and unsaturated hydraulic conductivity of silt clay loam soil at different temperature

T_w (^{o}C)	I_i (ml)	I_f (ml)	θ_i (m^3/m^3)	θ_f (m^3/m^3)	EC_i (mS/c)	EC_f (mS/cm)	T_i (^{o}C)	T_f (^{o}C)	D (cm)	k(h) cm/s
8	86.33	64.17	0.22	0.24	0.53	0.50	11.60	7.78	10	1.61E-05
25	84.83	48.17	0.24	0.30	0.56	0.59	24.12	24.45	10	5.06E-05
35	88.50	54.33	0.21	0.29	0.57	0.67	26.43	27.10	10	3.06E-05

Mean values from 6 replicates of measurement in laboratory as seen in Table 5.2 such as EC_f and T_f are increased with increasing temperature. The initial soil water contents which were prepared for hydraulic conductivity measurement were measured also using FDR 5TE moisture sensor showed 0.22, 0.24 and 0.21 m^3/m^3 at 10, 25 and 35^{o}C respectively. The FDR 5TE moisture sensor was not much affected by the difference in temperature of water which was mixed with the soft-earth to set the initial water content to 0.2 m^3/m^3 and compact the soil for measurement to a bulk density of 1.5 g/cm^3. However, research study in the department water resources on FDR 5TE moisture sensor shows measuring water content of the soil is varied due to temperature difference i.e. at night and day time it measures different water content at the same measuring spot.

Table 5.3. Soil water stored after infiltration, water out flown from reservoir container, viscosity and density of water, measured and recalculated unsaturated hydraulic conductivity

T_w (oC)	$\theta_{fav.}$-$\theta_{iav.}$ (m^3/m^3)	$I_{iav.}$-$I_{fav.}$ (ml)	μ (Kg/m s)	ρ (g/cm^3)	K(h) (cm/s)	Recal. K(h) (cm/s)
8	0.02	22.17	0.001413	0.99991	1.6E-50	3.19E-05
25	0.054333	36.67	0.00089	0.999971	5.06E-05	5.06E-05
35	0.082667	34.17	0.000719	0.99406	3.06E-05	9.89E-05

Mean values of the data measured in laboratory as seen in Table 5.3 such as μ, ρ are decreasing, but soil water stored after infiltration and recalculated K(h) are increasing decreasing viscosity and density.

The average values of measured and recalculated unsaturated hydraulic conductivity at 10 and 35^{o}C were 1.61E-05 & 3.06E-05 cm/s and 3.19E-05 & 9.8E-05 cm/s respectively. The measured value at reference temperature was 5.06E-05 cm/s.

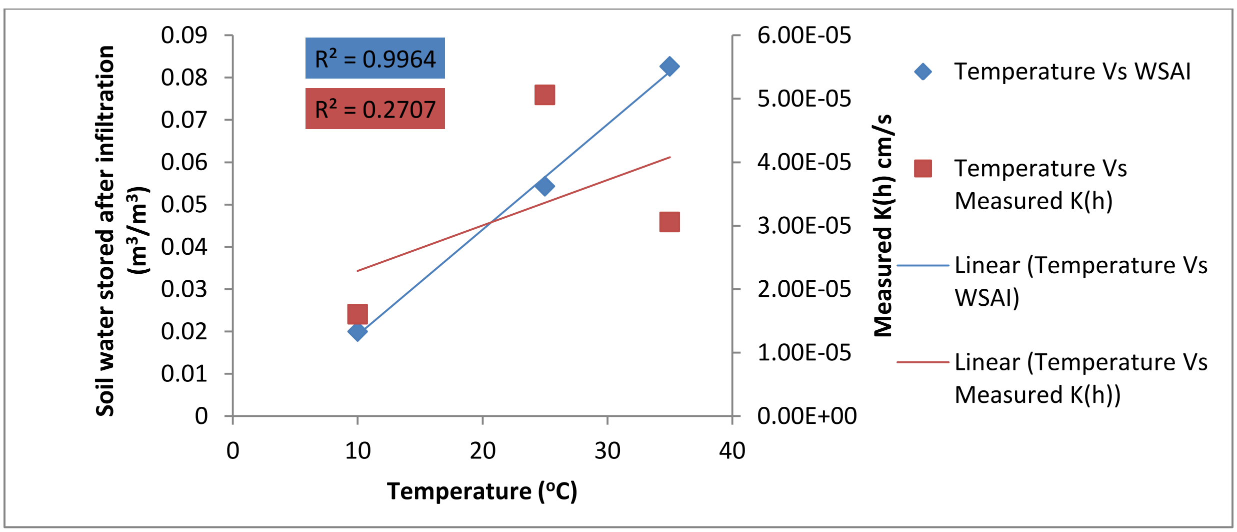

Fig.5.1 **Graph of temperature to soil water stored after infiltration and measured hydraulic conductivity.**

Figure 5.1 shows that the linear relationship between temperature and water stored in the soil after measurement has very high significant dependence or very strong correlation, $R^2 = 99.64\%$. However, temperature and measured hydraulic conductivity has weak significant dependence, $R^2 = 27.07\%$.

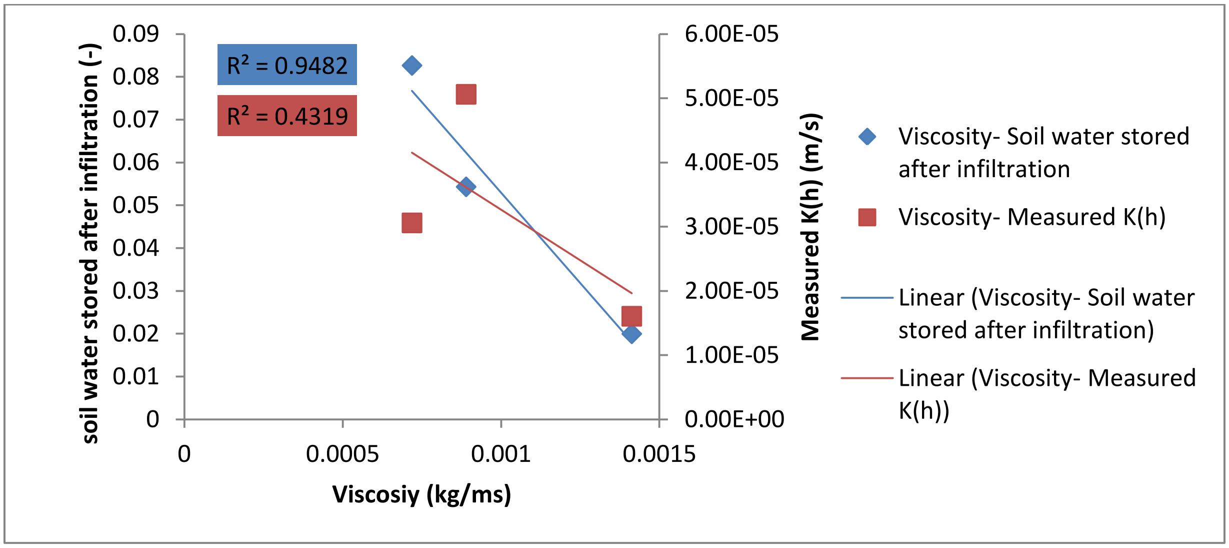

Fig.5.2 **Graph of viscosity, soil water stored after infiltration and measured hydraulic conductivity.**

Figure 5.2 shows that the linear relationship between viscosity and water stored in the soil after measurement has very high significant dependence or very strong correlation, $R^2 = 94.82\%$. However, the measured hydraulic conductivity has moderate dependence with water viscosity, $R^2 = 43.19\%$. There was a significant difference in measured hydraulic conductivity at temperature of 25 and 10°C at p-value = 0.0002766 which is< 0.05 , confidence interval 95%, t=11.9985 and tα = 2.776, df = 4. There was also significant difference between measured hydraulic conductivity at 25 and 35°C though the measurement was biased during experiment, hydraulic conductivity at 35°C seem lower, p-value = 0.008596, confidence interval 95%, t = 4.1871 and tα = 2.776, df = 5. There was no significant difference at temperature of 10 and 35°C, both measured P-value = 0.0516 and t value = 2.7457 were greater than 0.05 and less than 2.776 respectively.

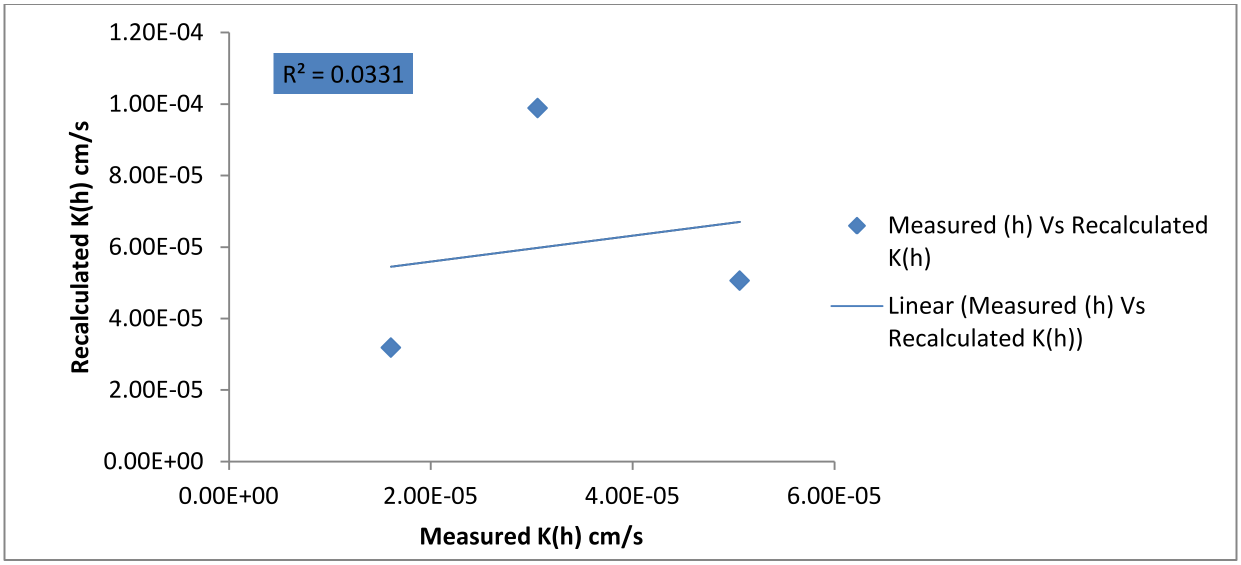

Fig.5.3 **Graph of the relationship between measured and recalculated unsaturated hydraulic conductivity**

Fig.5.3 shows the measured unsaturated soil hydraulic conductivity, K(h), value at (full of air bubble) high temperature measuring condition (35^{o}C) was smaller as compared to the result at (no air bubble) room temperature (25^{o}C). The trend of measured value of K(h) at 10 and 25^{o}C was increasing similar to the trend of the recalculated K(h) but the trend value of the measured and the recalculated K(h) were different at 35^{o}C.

Table 5.4. Field measured values of water temperature, initial and final infiltration water volume, soil moisture content, electrical conductivity, and soil temperature, soil depth and unsaturated hydraulic conductivity of silt clay loam soil at different temperature

As it is described in methods and materials of the study the measurement of parameters and collection of data has been done. The parameters measured and recorded at field experiment are presented in table 5.4 below including calculated results of unsaturated soil hydraulic conductivity.

T_w (^oC)	I_i (ml)	I_f (ml)	θ_i (m^3/m^3)	θ_f (m^3/m^3)	EC_i (mS/cm)	EC_f (mS/cm)	T_i (^oC)	T_f (^oC)	D (cm)	K(h) cm/s
10	34.6	30.8	0.234	0.28	0.37	0.47	89	61	5	5.03E-05
	34.3	30.7	0.242	0.301	0.3	0.47	90	67	5	2.30E-05
	31.6	28.6	0.233	0.287	0.36	0.5	90	60	5	2.30E-05
	31.7	28.6	0.232	0.282	0.23	0.47	89	64	5	3.44E-05
25	30.9	30.3	0.256	0.283	0.36	0.45	88	71	5	1.15E-05
	31.8	30.1	0.258	0.292	0.5	0.49	87	63	5	2.30E-05
	30.4	29	0.229	0.268	0.38	0.51	87	61	5	3.44E-05
	30.2	29.2	0.24	0.261	0.4	0.52	88	61.5	5	2.30E-05
35	30.7	27	0.17	0.273	0.17	0.56	86	49	5	3.44E-05
	30.7	24.5	0.196	0.234	0.26	0.48	81	30	5	1.15E-05
	29.3	25.4	0.196	0.275	0.34	0.51	87	38.5	5	5.76E-05
	27.7	25.6	0.174	0.275	0.31	0.5	86	37	5	5.76E-05

Table 5.4 shows directly measured parameters and calculated unsaturated hydraulic conductivity at 2 cm tension on silty clay loam soil at field experiment site. The value of K(h) is range from 2.2951-5.0271E-05 cm/s, 1.1475-3.4426E-05 cm/s, 1.1475-5.7376E-05 cm/s at a temperature of 8^oC, 25^oC and 35^oC respectively. The θ_i and θ_f (m^3/m^3), EC_i (mS/cm) and EC_f (mS/cm), and T_i (^oC) and T_f (^oC) were continuously recorded and their influence on hydraulic conductivity will be shown below in table 5.5.

Table 5.5. Average field measured values of infiltration volume, soil moisture content, electrical conductivity, soil temperature, soil depth, unsaturated hydraulic conductivity of silt clay loam soil at different temperature and temperature of water in MDI after final infiltration measurement (T_{wai})

T_w	I_i(ml)	I_f (ml)	θ_i (m^3/m)	θ_f (m^3/m^3)	EC_i (mS/c)	EC_f (mS/cm)	T_i(^oC)	T_f (^oC)	K(h) (cm/s)	T_{wai} (^oC)
10	89.50	63.00	0.24	0.29	0.32	0.48	33.05	29.68	3.26	29.6
25	87.50	64.13	0.25	0.28	0.41	0.49	30.83	29.70	2.30	30.83
35	85.00	38.63	0.18	0.26	0.27	0.51	29.60	25.63	4.02	33.05

The field measurements were different from the measurements in laboratory being exposed to direct sun radiation, wind blow, cloud cover and to continuous temperature fluctuations. As it was observed the average measured results of initial and final soil temperature on table 5.5 are higher unlike laboratory measured results. If we take electrical conductivity (sensitive to temperature change) of the soil as an indicator in response to the temperature of measuring infiltration water in the soil, results of electrical conductivity (the measure of soluble salts) in the laboratory were increasing with temperature (see fig. 5.5), change from 0.53-0.50, 0.56-0.59 and 0.57-0.67mS/cm for temperature of 10, 25 and 35^{o}C respectively. However, at field measurement water temperature influence on electrical conductivity is not clear, for instance 10^{o}C in laboratory has lower electrical conductivity but at field similar to 25 and 35^{o}C temperature. Unsaturated hydraulic conductivity and water stored after infiltration result have weak relation to the temperature of water since the temperature of measuring water in minidisk reservoir was shifted to the temperature of its surrounding (i.e. heated by direct sun ray or cooled due to cloud cover and/or wind) during measurement. Measurement result of K(h) at initial water content of 0.18, 0.24 and 0.25 m^3/m^3 were 4.02E-05, 3.26E-05 and 2.3E-05 cm/s respectively which is inversely related to each other.

Though minidisk infiltrometer is easy to use it for measurement, very handy to carry at field and relatively not expensive field measurement of unsaturated hydraulic conductivity, using minidisk infiltrometer at field and in laboratory measurement needs great attention to temperature change. The water in minidisk infiltrometer for measurement of unsaturated soil hydraulic conductivity in laboratory at temperature of 35^{o}C was invaded by air bubbles but not at room temperature (25^{o}C) and lower temperature (10^{o}C). The invasion of air bubbles to the water in MDI at field was not temperature selective but observed in all experiment variables because of hot (summer) weather during measurement.

The measurement of water level height change was visual method, hence, each recorded height change was affected by air bubble noisy that obviously could affect hydraulic conductivity result (water level height change was used for calculation of infiltration rate).

***Fig.5.*4 Pictures taken to show air bubble formed during field measurement due to surrounding high temperature**

As the result at table 5.4 showed, water temperature set to the experiment was 10, 25 and 35^{o}C, however, the water temperature in MDI reservoir chamber and the bubble chamber found to be higher after final water level measurement. The change was from 10^{o}C to 29.6^{o}C, from 25^{o}C to 30.8^{o}C and from 35^{o}C drop to 33^{o}C. The measurement was in summer and at hot temperature that causes the water temperature to be lifted up unlike water temperature in large reservoir. As a result of the temperature sensitivity of water in minidisk infiltrometer hydraulic conductivity measurement using minidisk infiltrometer will be restricted to regions where temperature is not higher or measure in a shed area and stable weather. Hydraulic conductivity measurement in desert areas, mountain plateau and meadows at hot temperature and direct sun radiation will not be applicable. The bubble effect is actually not clear to discuss plainly (further study is required). Air bubble formation at high temperature may be inevitable since dissolved air bubbles are higher in cold water than warm water.

***Fig.5*.5 Pictures taken which show water temperature measurement in MDI reservoir after final infiltration at field**

Table 5.6. Soil water stored after infiltration, water out flown from reservoir container, viscosity and density of water, measured and recalculated unsaturated hydraulic conductivity

T_w (°C)	θf_{av}.- θiav. (m^3/m^3)	I_{iav}-I_{fav} (ml)	μ (kg/ms)	ρ (g/cm^3)	K(h) cm/s	Recal. K(h) cm/s
10	0.05225	26.5	0.001519	0.99991	3.26E-05	1.34E-05
25	0.03025	23.375	0.00089	0.999971	2.29E-05	2.29E-05
35	0.08025	46.375	0.000719	0.99406	4.02E-05	2.82E-05

The results from table 5.6 shows that initial water content has direct influence on the result of unsaturated hydraulic conductivity and water volume stored in the soil after measurement of infiltration at field. Though temperature influence was not linear, the volume of water stored in the soil after infiltration measurement, the volume of water out flowed from minidisk infiltrometer reservoir and unsaturated hydraulic conductivity were consistent.

When we see in table 5.6 the measured value of K(h) at 10 and 35°C were greater than recalculated K(h) at 10 and 35°C. Measured K (h) values were 3.26E-05 and 4.02E-05 cm/s and recalculated K(h) values were 1.34E-05 and 2.82E-05 cm/s.

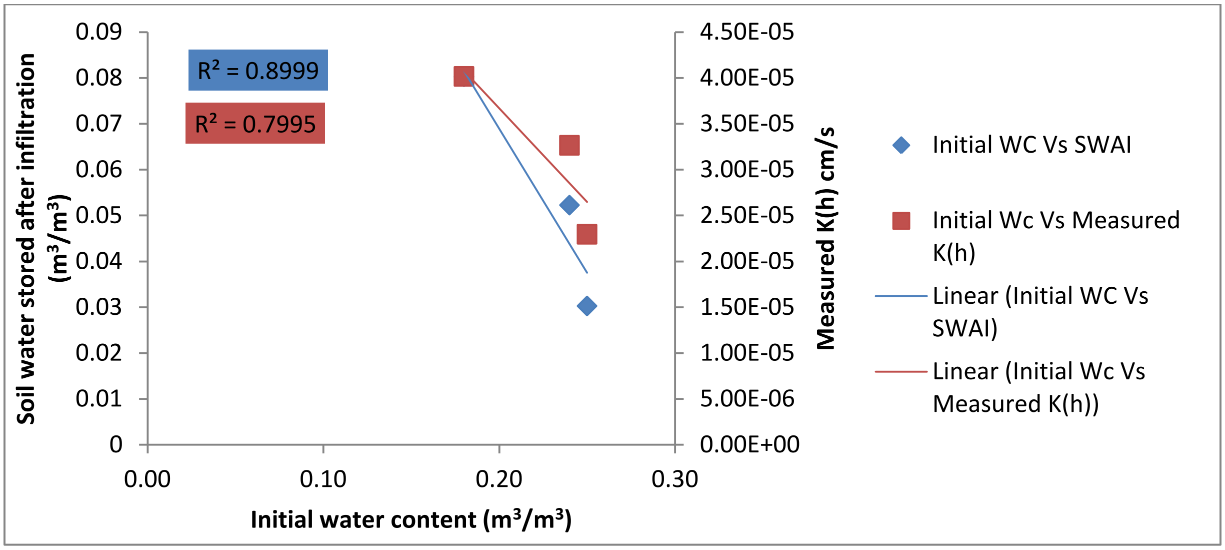

Fig.5.6 **Graph of initial water content, soil water stored after infiltration and measured hydraulic conductivity.**

Fig. 5.6 indicated that inverse correlation of measured initial water content to water stored in the soil after infiltration and measured hydraulic conductivity has high significant dependence, R^2 = 79.95% and very high significant dependence, R^2 = 89.99% respectively.

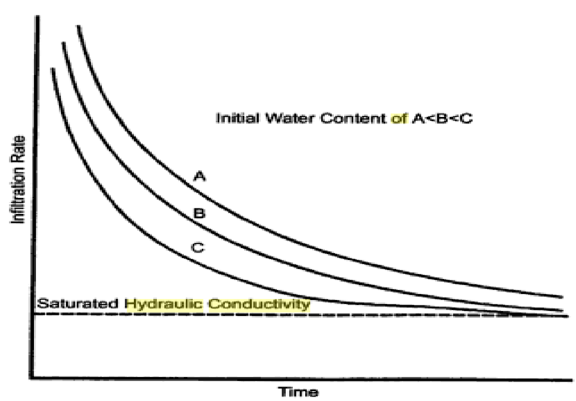

Fig.5.7 **The effect of initial water content of soil on infiltration rates (Evett et al., 1999)**

The graph of fig.5.7 taken from Everett et al. (1993) clearly illustrates that the higher initial water content the lower the infiltration rate, infiltration rate is a measurement recorded at field and laboratory to calculate hydraulic conductivity of a given soil. The experiment conducted on the influence of initial water content to the measurement of hydraulic conductivity in the same experiment site of this study showed that the decreasing trend of hydraulic conductivity with increasing of initial water content (Lufinkova, 2012). 23.9, 31.7 and 35.4% initial water content set to the experiment for 3 cm tension of minidisk infiltrometer had calculated corresponding result of unsaturated hydraulic conductivity of 7.683E-03, 7.417E-03, and 7.233E-03 cm/s.

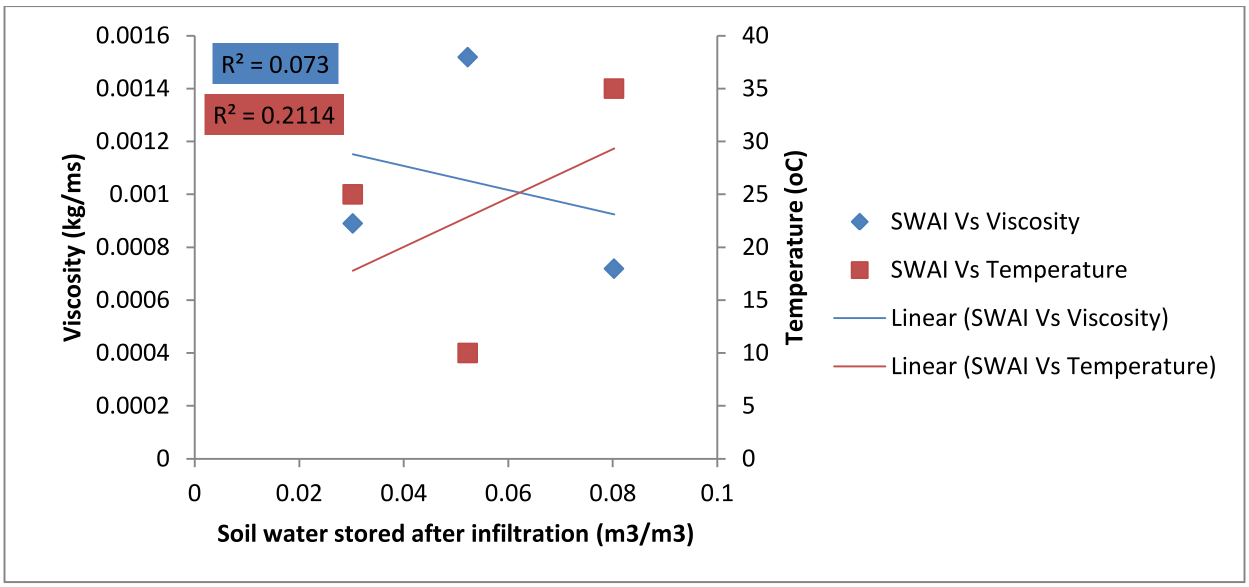

Fig.5.8 **Graph of the relationship between soil water stored after infiltration to viscosity and temperature**

Fig. 5.8 shows that relationship between volume of soil water stored after infiltration and viscosity, and relationship between volume of water stored in soil after infiltration and temperature were weak significant dependence R^2 = 7.3 and 21.14% respectively.

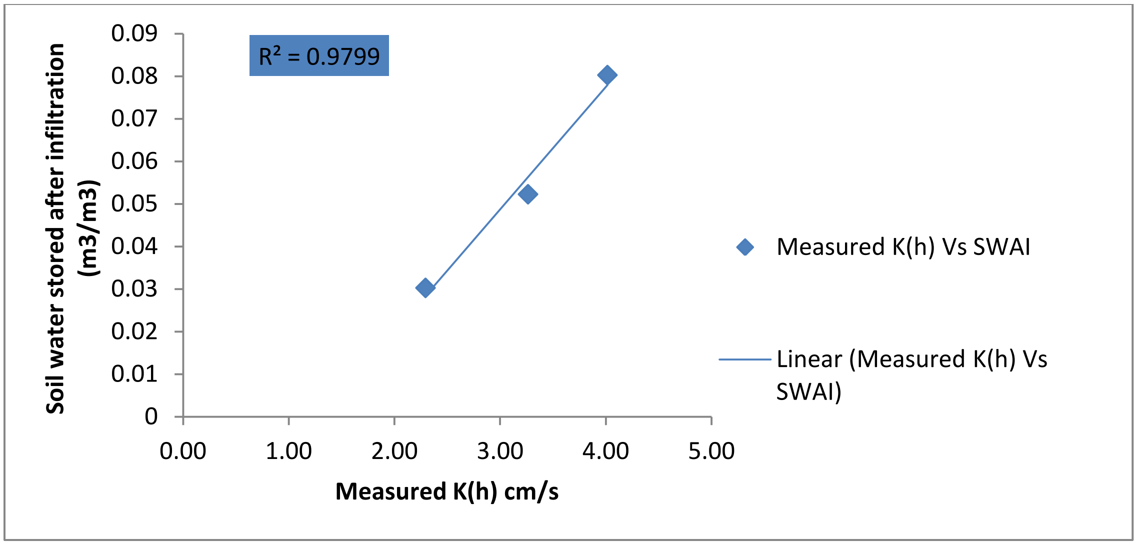

Fig. 5.9 **Graph of the relationship between Soil water stored after infiltration and measured hydraulic conductivity**

Fig. 5.9 The relationship of water stored in soil and hydraulic conductivity was very high significant dependence, $R^2 = 97.99\%$.

There was no significant difference of unsaturated soil hydraulic conductivity between field measurements of all the three temperature variables. The p-values were 0.1823, 0.4354 and 0.5896 which are greater than 0.05, t values were also 0.6021, 0.8979 and 1.7287 which are less than $t\alpha_{(4-1)}$ equals to 3.182 in confidence interval of 95%. Because the temperature of measuring environment was very high that lifted up temperature of small volume water in few minutes of time. Note that the temperature of the water measured after infiltration measurement at field was shifted from 10°C, 25 and 35 to 29.6, 30.83 and 33.06°C respectively because the measurement took place in summer hot temperature. Like the relationship of initial soil water content to unsaturated soil hydraulic conductivity, the relationship between the volumes of water stored after infiltration was inversely proportional to initial soil water content.

Table 5.7 Hydraulic conductivity measurements statistical evaluation at field and in laboratory taking Standard Deviation (SD) and Coefficient of Variance (CV)

Temperature (^oC)	Field Measurements		Laboratory Measurements	
	SD (cm/s)	**CV (%)**	**SD (cm/s)**	**CV (%)**
10	1.29331E-05	35.0	5.75596E-06	39.6
25	9.36971E-06	5.1	2.5965E-06	40.8
35	2.19734E-05	35.1	1.10006E-05	54.7

The variability of measured unsaturated soil hydraulic conductivity data during field experiment was very high at 10 and 35^oC and lower variability at 25^oC. Where as in laboratory measurements unsaturated soil hydraulic conductivity variation was increasing with temperature i.e. 39.6, 40.8 and 54.7% CV at temperature of 10, 25 and 35^oC respectively.

Laboratory measurement variation was higher with temperature and vice versa.

6. Conclusion and Recommendation

6.1. Conclusion

Rates of soil water infiltration and subsurface water movement are important to researchers developing soil management practices to minimize potential ground water contamination from land applied chemicals. Therefore, a simple and rapid field technique of determining field unsaturated hydraulic conductivity would be useful in achieving this objective (Ankeny et al., 1991). In this regard minidisk infiltrometer contribution is very high; however, unsaturated soil water hydraulic conductivity measurement using minidisk infiltrometer is significantly influenced by temperature changes at field and weather condition of the day.

Minidisk infiltrometer is very small to measure unsaturated hydraulic conductivity without changing water temperature in reservoir container and the bubble chamber at field condition. The temperature of water in minidisk infiltrometer is very dynamic that can affect unsaturated hydraulic conductivity actual result due to the change in viscosity of water that can increase or decrease the infiltration of water in to the soil (Hannula and Poeter, 1995). During field measurement the temperature of water was shifted from $10^{o}C$ to $29.6^{o}C$ (i.e. actual temperature of water in the soil is not the same as that of water temperature we used to measure hydraulic conductivity), thus, hydraulic conductivity is a function of fluidity and intrinsic permeability. Fluidity is affected by temperature that changes viscosity. Viscosity affects hydraulic conductivity of the experiment.

Moreover, a lot of bubbles are formed over the entire water in minidisk infiltrometer that might affect accurate water level height change and smoothness of water outflow from the reservoir to the soil. And also may affect the contact between the sintered disk and water in reservoir chamber. Consequently, the estimated result of unsaturated soil hydraulic conductivity will be affected.

The effect of initial water content was higher and more influencing at field than the change in temperature of water on the value of K(h) even though there was an intention to keep it constant as much as possible.

6.2. Recommendations

Field measurement of unsaturated hydraulic conductivity at room temperature (25^{o}C) and below, and at stable (no wind), shade (no direct sun radiation) is safe (create good microclimate).

If there is high temperature and direct sun radiation at the area which can cause air bubble in the water used for measurement, use deaerated water can minimize the problem.

Research activities are needed to improve minidisk infiltrometer to prevent air bubble formation during measurement at high temperature or to neglect bubble noisy and study suitable working temperature range of MDI.

7. References

Abrisqueta, J.M., Plana, V., Ruiz-Canales, A., Ruiz-Sánchez, M. C. 2006. Unsaturated hydraulic conductivity of disturbed and undisturbed loam soil. Spanish Jour. of Agri. Research. 4(1). 91-96.

Ankeny, M.D., Ahmed, M., Kaspar, T.C., Horton, R. 1991. Simple Field Method for Determining Unsaturated Hydraulic Conductivity. Soil Sci. Soc. Am. J. 55(2). 467-470.

Ankeny, M. D. 1992. Methods and Theory for Unconfined Infiltration Measurements. Soil Sci. Soc. of Am. J. 30. 123-141.

Báťková, K., Matula, S., Miháliková, M. 2012. Multimedial Study Guide of Field Hydropedological Measurements [on-line]. English version. Czech University of Life Sciences Prague. Prague. Czech Republic. No pagination. Available at: http://hydropedologie.agrobiologie.cz.

Bear, J. 1972. Dynamics of Fluids in Porous Media. Elsevier, New York, NY, USA, 764 p.

Hillel, D. 1998. Environmental Soil Physics. Academic Press. New York, USA. p. 40, 142, 187. ISBN:0-12-348525-8.

Brooks, R.H. Corey, A. T. 1966. Properties of porous media affecting fluid flow. Proc. Am. Soc. Civ. Eng. J. Irrigation Drainage Div. IR. 2. 61-68.

Brown, G., Hsieh, H., Lucero D.A. 2000, Evaluation of laboratory dolomite core sample size using representative elementary volume concepts, Water Resour. Res., 36(5), 1199–1207.

Bruce, R. R., Klute, A. 1956. The measurement of soil water diffusivity. Soil Sci. Soc. Am. Proc. 20. 458-562.

Buttafuoco, G., Ricca, N. 2013. Tension Disk Infiltrometry to Determine Hydraulic Properties of Coarse Textured Soils on Granites of the Sila Massif (Italy). In: Wöhrle, N., Scheurer, M. (Eds.). Eurosoil 2004. September 4-12, 2004. University of Freiburg, Freiburg, Germany. CD. Available on-line from: <http://www.bodenkunde.uni-freiburg.de/eurosoil-en>

Cho, W.J., Lee, J.O., Chun, K.S. 1999. The temperature Effects on Hydraulic Conductivity of Compacted Bentonite .Applied Clay Science.14.47–58.

Cook, F.J. 2007. Comparisons of Transient Analytical Methods for Determining Hydraulic Conductivity Using Disc Permeameters. In MODSIM07 - Land, Water and Environmental Management: Integrated Systems for Sustainability, Proceedings. International Congress on Modelling and Simulation - Land, Water and Environmental Management: Integrated Systems for Sustainability, MODSIM07, Christchurch, (1212-1218). December 10-13, 2007.

Dani, O., Wraith, M. J. 2002. Soil Physics Companion. In: Arthur W. Warrick (ed.). CRC Press LLC. USA. p. 62. ISBN: 0-8493-0837-2.

Decagon Devices, Inc., 2012. Mini Disk Infiltrometer - User's manual, Decagon Devices, Inc., Pullman WA.

Eva, L. D. 1994. Effect of Temperature and Pore Size on the Hydraulic Properties and Flow of Hydrocarbon Oil in the Subsurface. Journal of Contaminant Hydrology. 16. 55-86.

Evett, S.R., F.H. Peters, O.R. Jones, and P.W.Unger. 1999. Soil hydraulic conductivity and retention curves from tension infiltrometer and laboratory data. In (M.Th. van Genuchten, F.J. Leij, and L. Wu Eds.) Characterization and Measurement of the Hydraulic Properties of Unsaturated Porous Media, Part I. U.S. Salinity Laboratory, USDA-ARS.

Gardner, W. R. 1956. Calculation of capillary conductivity from pressure plate outflow data. Soil Sci. Soc. Am. Proc. 20. 317-320.

Green, R.E., Ahuja, L.R, Chong, S.K. 1986. Hydraulic Conductivity, Diffusivity, and Sorptivity of Unsaturated Soils. In Klute, A. (Ed.) In: Klute, A. (Ed.) Methods of Soil Analysis: Part 1. Physical and Mineralogical Methods, 9. Second Edition. Madison, Wisconsin USA, American Society of Agronomy - Soil Science Society of America. p. FROM-TO. ISBN 0-89118-088-5.

Hannula, S.R., Poeter, E. 1995. Open file report on Temporal and Spatial Variation of Hydraulic Conductivity in a Stream Bed in Golden Colorado. Colorado water resources research institute, Colorado State University. Fort Collins, Colorado 80523. April 11, 1991.

Hillel, D. 1998. Environmental Soil Physics. Academic Press. New York, USA. p. 133. ISBN: 0-12-348525-8.

Hoekstra, P. 1966. Moisture Movement in Soils Under Temperature Gradient with the Cold-Side Temperature Below Freezing. Water Resour. 2. 241-250.

Homolák, M., Capuliak, J., Pichler, V., Lichner, L. 2009. Estimating Hydraulic Conductivity of a Sandy Soil Under Different Plant Covers Using Minidisk Infiltrometer and a Dye Tracer Experiment. Biologia. 64(3). 600-604.

IUSS Working Group WRB. 2006. World reference base for soil resources 2006. World Soil Resources Reports No. 103. FAO, Rome.

Jarsj, J., Destouni, G., Yaron, B. 1997. On the relation between viscosity and hydraulic conductivity for volatile organic liquid mixtures in soils. Elsevier. Jour. of Contam. Hydro. 25. 113-127.

Kaempfer, T.U., Schneebeli, M., Sokratov, S.A. 2005. A microstructural approach to model heat transfer in snow. Geographical Research Letters. 32. L21503. 1-5.

Kelishadi, H., Mosaddeghi, M.R., Hajabbasi, M.A., Ayoubi, S. 2014. Near-saturated soil hydraulic properties as influenced by land use management systems in Koohrang region of central Zagros, Iran. Geoderma. 213. 426–434.

Kirkham, M.B. 2005. Principles of Soil and Plant water Relations. Elsevier Academic Press, Kansas State University, USA. p. 31. ISBN: 92101-4495.

Lal, R., Shukla, M.K. 2004. Principles of Soil Physics. Marcel Dekker, INC. New York. USA. p. 528. ISBN: 0-3247-5324-0.

Li1, X.Y, Gonza´lez, A., Sole´benet, T.A. 2005. Laboratory Methods for the Estimation of Infiltration Rate of Soil Crusts in the Tabernas Desert Badlands. Elsevier Science. Catena. 60. 255–266.

Lufinková, J. 2012. Factors affecting determination of unsaturated soil hydraulic conductivity in situ using tension infiltrometer. Diploma thesis. Czech University of Life Sciences Prague. Faculty of Agrobiology, Food and Natural Resources. Prague, Czech Republic. p.51.

Marshall, C.E. 1964. The Physical Chemistry and Mineralogy of Soils. Wile, New York.In Hillel, D. 1998. Environmental Soil Physics. Academic Press. New York, USA. p. 48.ISBN:0-12-348525-8.

Michelle, T.H., Vliet v., Wietse, H.P., Franssena, J. R., Yearsley, Ludwiga F., Haddelandc, I., Lettenmaier, D. P., Kabat, P. 2013. Global river discharge and water temperature under climate change. Science Direct. Global Environmental Change. 23. 450–464.

Miháliková, M., Matula, S., Doležal F. 2013. HYPRESCZ – Database of Soil Hydrophysical Properties in the Czech Republic. Soil & Water Res. 8. 34-44.

Miyazaki, T. 2006. Water Flow in Soils. CRC Press. 2nd ed. University of Tokyo. Tokyo, Japan Boca Raton, FL 33487-2742. p.31.

Moore, R.E. 1939. Water conduction from shallow water tables. Hilgardia. 2. 383-426.

Moret, D., Lo´pez, M.V., Arru´e, J.L. (2004) TDR application for automated water level measurement from Mariotte reservoirs in tension disc infiltrometers. Journal of Hydrology. 297. 229– 235.

Mualem, Y. 1976. Anew model for predicting the hydraulic conductive of unsaturated porous media. Water Resour. Res. 12.513–522.

Nakhae, M., Šimůnek, J. 2014. Estimating the Soil Hydraulic and Thermal Properties of Unsaturated Porous Media Using HYDRUS-2D. J. Hydrol. Hydromech. 62. 7–15.

Nimmo, J.R., Miller, E.E. 1986. The temperature dependence of isothermal moisture vs. potential characteristics of soils. Soil Sci. Soc. Am. J. 50: 1105-1113.

Philip, J.R., Vries, D.A.de. 1957. Moisture movement in porous materials under temperature gradients. Trans. Am. Geophys. Union 38:222-232. .

Romero, E., Gens, A., Lloret, A. 2001.Temperature Effects on the Hydraulic Behaviour of Unsaturated Clay. Jour. of Geotech. and Geol. Engi. 19. 311-332.

Shukla, M.K., Lal, R. 2004. Principles of Soil Physics. Marcel Dekker, INC. New York. p. ISBN: 0-3247-5324-0.

Su, G. W., Jasperse, J., Seymour, D., Constant, J. 2004. Ground Water Estimation of Hydraulic Conductivity in an Alluvial System Using Temperatures. National Ground Water Association, Ground Water. 42. 890-901.

Vandenbygaart, A.J., Protz. R. 1999. The representative elementary area (REA) in studies of quantitative soil micromorphology. Geoderma . 89. 333–346.

Vandervaere, J.P, Vauclin, M, Elrick D.E. 2000. Transient Flow from Tension Infiltrometers. Part 1. The two-parameter Equation. Soil Sci. Soc. Am. J.64. 1263-1272.

Van Genuchten, M.T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci. Soc. Am. J. 44: 892-898.

Watson, K. K. 1966. An instantaneous profile method for determining the hydraulic conductivity of unsaturated porous materials. Water Resour. Res. 2. 709-715.

Zhang, R. 1997. Determination of Soil Sorptivity and Hydraulic Conductivity from the Disk Infiltrometer. Soil Sci. Soc. Am. J. 61. 1024–1030.

Zhang, Z. Zhang, Z., Low, P. F., Arq, O.C., Roth, B.F. 1993. The Effect of Temperature on the Swelling of Monmorillonite. Clay Minerals. 28. 25-31.

8. Nomenclature, list of abbreviations

8.1 Symbols

A = value relating to van Genuchten parameters (-)

A = the cross-sectional area of flow (L^2)

C_1 = parameter related to hydraulic conductivity (-)

C_2 = parameter related to soil sorptivity (-)

C(ψm) = soil-moisture capacity at a function of matric potential (L^{-1})

D = prepared soil profile depth (L)

e = void ratio (-)

EC_i = measured initial electrical conductivity (mS/cm)

EC_f = measured final electrical conductivity (mS/cm)

F = force acting on an area (N)

f = the fluidity of permeating liquid ($L^{-1}T^{-1}$).

g = acceleration due to gravity (LT^{-2})

h = pressure heads of hydraulic conductivity (L)

hT = pressure heads at temperature(L)

I = cumulative infiltration (L^3T^{-1})

I_i = measured initial water infiltration volume in minidisk infiltrometer reservoir (L^3)

I_f = measured final water infiltration volume in minidisk infiltrometer reservoir (L^3)

k = intrinsic permeability (L^{-1})

K(h)adj. = calculated unsaturated hydraulic conductivity from viscosity and density at measured temperature of water (LT^{-1})

K_r(h) = relative hydraulic conductivity (LT^{-1})

Ks = saturated hydraulic conductivity (LT^{-1})

$K(\theta)$ = unsaturated hydraulic conductivity of the porous medium as a function of water content (LT^{-1})

$K(\psi m)$ = unsaturated hydraulic conductivity of the porous medium as a function of suction (LT^{-1})

$K(\theta)T$ = unsaturated hydraulic conductivity at a temperature (LT^{-1})

Kref = hydraulic conductivities at reference temperature (LT^{-1})

KT = hydraulic conductivities at a temperature (LT^{-1})

$K(h)$ = unsaturated soil hydraulic conductivity at applied suction (LT^{-1})

L = the length of soil column (L)

m = empirical constant affecting the shape of SWRC (-)

m_c = mass of clay(M)

m_w = mass of water(M)

Pc = the capillary pressure

Q = the flow per unit cross sectional ($L^{-2}L^{-2}$)

q = flux density (LT^{-1})

R^2 = coefficient of correlation between measured variables

S_r = degree of saturation (%)

T = temperature (^oC or ^oK)

T_f = measured final soil temperature (^oC)

T_i = measured initial soil temperature (^oC)

Tref = reference temperature (^oC or ^oK)

t = time (T)

T_w = Water temperature for infiltration measurement (^oC)

V = volume of water (L^3)

α = empirical parameter which is the inverse of air entry suction (L^{-1})

αh^* = the temperature scaling factor for the pressure head (-)

αK^* = the temperature scaling factor for the hydraulic conductivity (-)

β_T = empirical coefficient that fits relative viscosity data over a temperature range(-)

γ = interfacial surface tension in soil (MT^{-2})

Θ = Effective or dimensionless water content (-)

θ = volumetric water content (L^3)

θ_f = measured final soil water content (L^3/L^3)

θ_i = measured initial soil water content (L^3/L^3)

θ_r = residual volumetric water content

θ_s= saturated volumetric water content

λ = interlayer spacing (L)

$\lambda_{c\,=}$ macroscopic capillary length (L)

σ = surface tension (MT^{-2})

σT = surface tensions at a temperature(MT^{-2}).

σref = surface tensions at reference temperature(MT^{-2}).

μref = dynamic viscosity at reference temperature ($ML^{-1}T^{-1}$)

μT = dynamic viscosity at a temperature ($ML^{-1}T^{-1}$)

μ_w = the absolute viscosity ($ML^{-1}\ T^{-1}$)

ρref = density at reference temperature(ML^{-3})

ρT = density at a temperature (ML^{-3})

ρ_w = the density of water (ML^{-3})

ΔH = the hydraulic head difference (L)

Π = swelling pressure (Mpa)

Ψ_T = total soil water potential (L)

Ψw = soil water potential (L)

ψ_m = matric potential (L)

ψ_s = the solute or osmotic potential (L)

ψ_z = gravitational potential (L)

τ = shearing stress (NL^{-2})

$\partial L =$ change in lenght (L)

∂t = change in time (T)

$\partial\theta$ = change in water content in a unit volume of soil (L^3)

$\partial u/\partial y$ = change in velocity gradient perpendicular to the stressed area *A*

∇ = vector differential operator for three-dimensional gradient

$-\nabla q$ = spatial gradient of flux (LT^{-1})

8.2 Abbreviations

i.e. - that is

CV- Coefficient of Variance

MDI- Mini Disk Infiltrometer

SD- Standard Deviation

STVF- Surface Tension Viscous Flow

SWRC- Soil Water Retention Curve

VOLM- Volatile Organic Liquid Mixture

9. Appendices

I. Numerical Constant values

$g = 9.81 m/s^2$

ρw = 0.99991 g/cm^3(10^oC), 0.999971 g/cm^3 (25^oC), 0.99406 g/cm^3 (35^oC)

μw = 0.001519(kg/ms) (10^oC), 0.00089(kg/ms) (25^oC), 0.000719(kg/ms) (35^oC)

Appendix II. Data Record Form for the Minidisk Infiltrometer

Location:			
Date:	Weather	θ initial:	Note
Tillage tr.	Temperature: ^{o}C	%	
Sign.			

Time		Reading
min	s	ml

Appendix III. Macro excel sheet for use with the Mini Disk Infiltrometer (From: Decagon Device, Inc., 2012.)

Instructions	
Step 1:	Enter measurement times beginning with zero
Step 2:	Enter corresponding volume measurements
Step 3:	Adjust selection field on graph to fit data

Time (s)	sqrt (t)	Volume (mL)	Infilt (cm)

Step 4:	Select Infiltrometer Type	
Step 5:	Select Soil Type	
Step 6:	Select Suction	

Radius	
alpha	
n/h_o	
Suction	

A	
C1	
K	

11.02317638

Appendix IV. Van Genuchten parameters for 12 soil texture classes and A values for a 2.25cm and 1.6 disk radii and suction values from 0.5 to 7cm (from: Decagon device, Inc., 2012).

	radius	2.25			.223189		A			5.483306
	alpha	n / ho	-0.5	-1	-2	-3	-4	-5	-6	-7
sand	0.145	2.68	2.835701	2.40407	1.727908	1.241921	0.892621	0.641565	0.46112	0.331427
loamy sand	0.124	2.28	2.9853	2.786831	2.4286	2.116417	1.844363	1.60728	1.400674	1.220625
sandy loam	0.075	1.89	3.877062	3.887982	3.909913	3.931969	3.954148	3.976453	3.998884	4.021441
loam	0.036	1.56	5.461148	5.717657	6.267384	6.869965	7.530482	8.254505	9.048139	9.918077
silt	0.016	1.37	7.921451	8.177401	8.714378	9.286617	9.896433	10.54629	11.23883	11.97683
silt loam	0.02	1.41	7.102076	7.367933	7.929874	8.534674	9.185601	9.886173	10.64018	11.45169
sandy clay loam	0.059	1.48	3.210664	3.523317	4.242925	5.109507	6.153081	7.409796	8.923184	10.74567
clay loam	0.019	1.31	5.857535	6.10902	6.644845	7.227667	7.861609	8.551155	9.301181	10.11699
silty clay loam	0.01	1.23	7.893227	8.094056	8.511175	8.949789	9.411007	9.895994	10.40597	10.94223
sandy clay	0.027	1.23	3.336287	3.570465	4.089288	4.683501	5.364059	6.143508	7.036218	8.058649
silty clay	0.005	1.09	6.076318	6.169307	6.359575	6.55571	6.757895	6.966316	7.181164	7.402639
clay	0.008	1.09	3.998056	4.096399	4.300401	4.514562	4.739389	4.975412	5.223189	5.483306

	radius	1.6					A			
	alpha	n / ho	-0.5	-1	-2	-3	-4	-5	-6	-7
sand	0.145	2.68	2.835701	2.40407	1.727908	1.241921	0.892621	0.641565	0.46112	0.331427
loamy sand	0.124	2.28	2.9853	2.786831	2.4286	2.116417	1.844363	1.60728	1.400674	1.220625
sandy loam	0.075	1.89	3.877062	3.887982	3.909913	3.931969	3.954148	3.976453	3.998884	4.021441
loam	0.036	1.56	5.461148	5.717657	6.267384	6.869965	7.530482	8.254505	9.048139	9.918077
silt	0.016	1.37	7.921451	8.177401	8.714378	9.286617	9.896433	10.54629	11.23883	11.97683
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sandy clay loam	0.059	1.48	3.210664	3.523317	4.242925	5.109507	6.153081	7.409796	8.923184	10.74567
clay loam	0.019	1.31	5.857535	6.10902	6.644845	7.227667	7.861609	8.551155	9.301181	10.11699
silty clay loam	0.01	1.23	7.893227	8.094056	8.511175	8.949789	9.411007	9.895994	10.40597	10.94223
sandy clay	0.027	1.23	3.336287	3.570465	4.089288	4.683501	5.364059	6.143508	7.036218	8.058649
silty clay	0.005	1.09	6.076318	6.169307	6.359575	6.55571	6.757895	6.966316	7.181164	7.402639
clay	0.008	1.09	3.998056	4.096399	4.300401	4.514562	4.739389			

Appendix V. Photos from field and lab measurement

Pictures taken during water temperature measurement in minidisk infiltrmeter reservoir after infiltration measurement using thermometer

Pictures that show air bubble formed at field measurement in summer in minidisk infiltrometer reservoir and bubble chamber respectively


Model S

Pictures taken during laboratory hydraulic conductivity measurement at temperature of 10, 25 and 35 ^{o}C.

Appendix VI. Field and laboratory result statistical data relationship output (R 3.02)

> Dataset <- edit(as.data.frame(NULL))

> summary(Dataset)

X10l X25l X35l X10f X25f X35f

Min. :1.15e-05 Min. :4.590e-05 Min. :2.295e-05 Min. :2.30e-05 Min. :1.15e-05 Min. :1.15e-05

1st Qu.:1.15e-05 1st Qu.:4.590e-05 1st Qu.:2.295e-05 1st Qu.:2.30e-05 1st Qu.:2.01e-05 1st Qu.:2.87e-05

Median :1.15e-05 Median :4.859e-05 Median :2.295e-05 Median :2.87e-05 Median :2.30e-05 Median :4.59e-05

Mean :1.61e-05 Mean :5.062e-05 Mean :3.060e-05 Mean :3.27e-05 Mean :2.30e-05 Mean :4.02e-05

3rd Qu.:2.30e-05 3rd Qu.:5.585e-05 3rd Qu.:4.016e-05 3rd Qu.:3.84e-05 3rd Qu.:2.58e-05 3rd Qu.:5.74e-05

Max. :2.30e-05 Max. :5.738e-05 Max. :4.590e-05 Max. :5.03e-05 Max. :3.44e-05 Max. :5.74e-05

NA's :1 NA's :2 NA's :2 NA's :2

> t.test(Dataset$X10f, Dataset$X10l, alternative='two.sided', conf.level=.95, paired=TRUE)

Paired t-test

data: Dataset$X10f and Dataset$X10l

t = 2.2153, df = 3, p-value = 0.1135

alternative hypothesis: true difference in means is not equal to 0

95 percent confidence interval:

-8.003081e-06 4.466387e-05

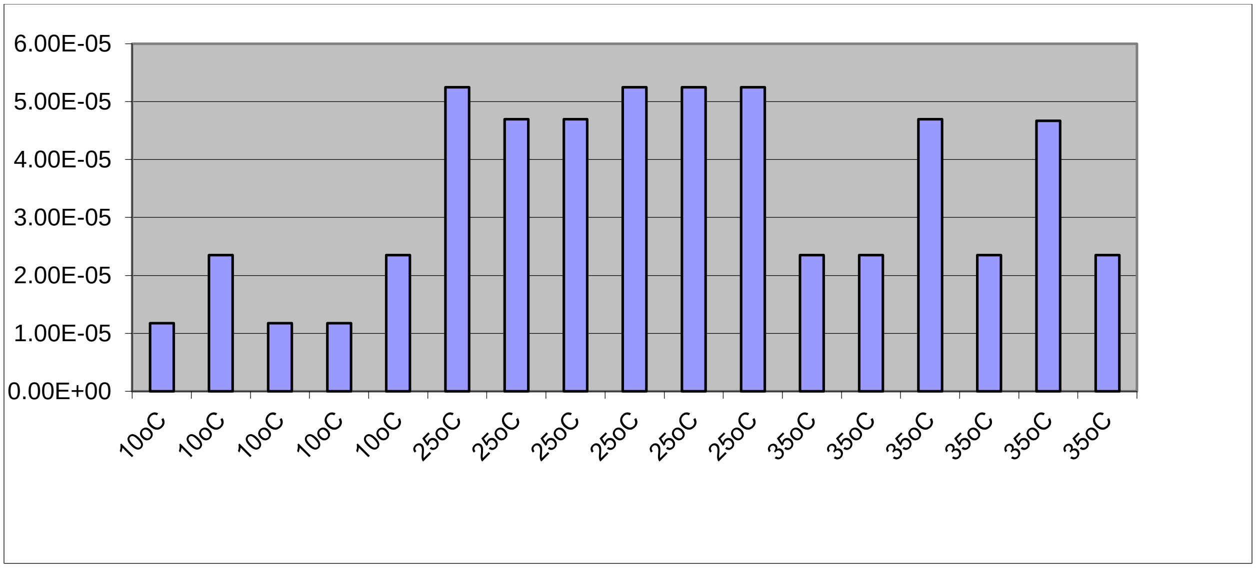
sample estimates:

mean of the differences

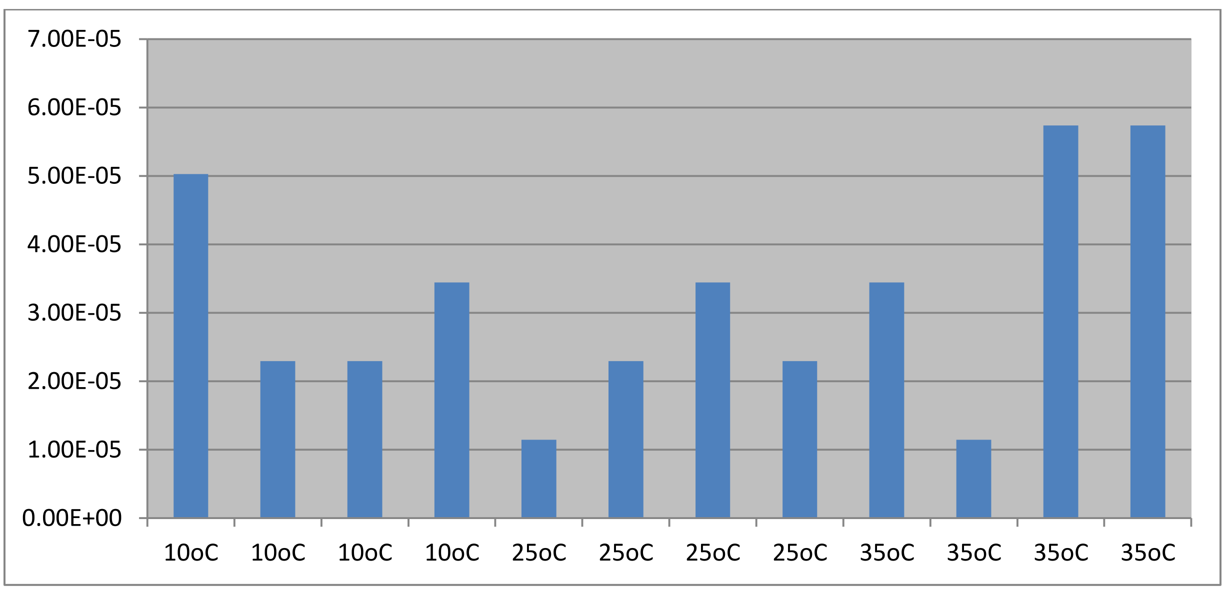
1.833039e-05

```
> t.test(Dataset$X25f, Dataset$X25l, alternative='two.sided', conf.level=.95, paired=TRUE)

            Paired t-test

data:  Dataset$X25f and Dataset$X25l
t = -4.1657, df = 3, p-value = 0.02517
alternative hypothesis: true difference in means is not equal to 0
95 percent confidence interval:
 -4.281276e-05 -5.728943e-06
sample estimates:
mean of the differences
          -2.427085e-05
> t.test(Dataset$X35f, Dataset$X35l, alternative='two.sided', conf.level=.95, paired=TRUE)

            Paired t-test

data:  Dataset$X35f and Dataset$X35l
t = 1.226, df = 3, p-value = 0.3077
alternative hypothesis: true difference in means is not equal to 0
95 percent confidence interval:
 -1.833076e-05  4.130386e-05
sample estimates:
mean of the differences
           1.148655e-05
```

Appendix VII. Measure hydraulic conductivity on spots at field and in laboratory



Graph which shows Temperatures and Unsaturated hydraulic conductivity measurements in laboratory on each prepared soil profile



Graph which shows Temperatures and Unsaturated hydraulic conductivity measurements at field on each spot