1. DESCRIPTION OF NATURAL SURFACES

Earth's solid surface represents an essential part of our ecosystem and because of its diversity a huge number of bio- & geo-physical parameters is required for its detailed description and/or monitoring. An important element of the earth's surface is the soil. Soils can be parameterised in terms of its consistency, and its dielectric and geometric properties. The dielectric properties of soils are expressed - apart from its consistency - primarily by its moisture content. On the other hand, the soil roughness describes the geometric characteristics of the soil surface.

Classical methods in soil physics are based on point measurements to estimate these two surface parameters. The most critical factor, besides the enormous time and working effort, is the extrapolation of point samples to catchment (or sub-regional) scale processes. The greatest error in the parameterisation occurs on the interface between different scales of hydrological to climatic or ecological models. Such mismatch problems have an important implication for the credibility of impact studies driven by the output of models. In order to bridge the scales between different science applications, e.g. climate hydrology and ecology models, methods for both observing and representing sub-grid-scale heterogeneity, as well as linking parameter/state variables across disparate scales, are in great need.

However, remotely sensed (from space or airborne sensors) information provides today new opportunities to hydrologists and ecologists from monitoring extended processes with a spatial resolution from about a meter up to a global scale. Remote sensing techniques provide data for hydrological model parameter estimation, computation of geo-physical parameters and for real–time forecasting. Indeed, remote sensing is capable of quantifying some of the key hydrological parameters for water balance modelling (precipitation, evaporation, snow cover, runoff, soil moisture and surface roughness).

Radar remote sensing with its sensitivity to the dielectric and geometric characteristics of objects, its weather independent imaging capability and its potential to acquire subsurface information, in dependence of the frequency band used, is one of the most promising approaches for surface parameter estimation.

Electromagnetic (EM) wave scattering on rough dielectric surfaces has been the subject of intensive studies for many decades. Many experimental measurements have been accumulated and many approaches have been developed in order to predict and interpret experimental data. Despite the large amount of research efforts the general surface scattering problem is analytically not completely solved. The following approaches can be categorised as approximate solutions, and hence with a more or less restricted applicability or exact but too general to be of practical importance (BECKMANN & SPIZZICHINO 1963).

1.1 Geometrical Properties

Rough surfaces can be divided into two main categories, the deterministic and the randomly rough surfaces. The first are periodic surfaces with a given profile and periodic irregularities. Historically, this class of rough surfaces was first investigated due to the simplicity of a non-statistical approach and thus enabling the simplification of the theory. Natural surfaces are categorised into the class of random rough surfaces, which are characterised by random irregularities on the surface and are best described by the statistical distribution of their deviation from a certain mean level.

Randomly rough surfaces are usually described in terms of their deviation from a smooth 'reference surface'. There are essentially two aspects describing the nature of a randomly rough surface: the spread of heights about the reference surface and the variation of these heights along the surface. A variety of equivalent statistical distributions and parameters may be used to parameterise these two surface properties. In this investigations, the parameter set of the root mean square height RMS, *s* and the surface correlation length, *l,* associated to the surface correlation function are considered the best for the parametric description of the natural surfaces.

1.1.1 Surface Roughness and Surface Correlation Length

The RMS height, *s,* is used to describe the vertical surface roughness and is defined as the standard deviation of the surface height variation in cm,

$$RMS_{\text{height}} = s = \sqrt{\frac{\sum_{i=1}^{n} (z_i - \bar{z})^2}{n-1}} \tag{1}$$

On the other hand, the surface correlation function $\rho(x)$ and the associated correlation length, *l,* are parameters used for the horizontal description of the surface roughness. In the discrete case, the normalised surface correlation function for a spatial displacement $x' = (j-1)\Delta x$ is given by

$$p(x') = \frac{\sum_{i=1}^{N+1-j} z_i z_{j+i-1}}{\sum_{i=1}^{N} z_i^2} \tag{2}$$

where z_{j+i-1} is a point with the spatial displacement from the point x_i (Fung 1994). The surface correlation length is defined as the displacement x' for which $\rho(x')$ between two points inhibits values smaller than 1/e (Euler's Value ≈ 2.7183)

$$p(l) = 1/e \tag{3}$$

Thus, the surface correlation describes the statistical independence of two points on a surface and increases with the correlation between two neighbouring points. For a smooth surface $l = \infty$.

For characterisation of a surface with two parameters, the RMS height, s, and the surface correlation length, l, describe the natural surface only as a two-dimensional stationary randomly rough surface. Both statistical parameters, s and l, are independent from each other according to the single scale roughness theory (BECKMANN & SPIZZICHINO 1963). l can be large or small for a high or low s. Experimental data acquired over natural surfaces show, that most bare soil surfaces are characterised by large spatial variations, making it difficult to determine consistent roughness parameters for modelling and inversion purposes. The most delicate point is the measurement of the correlation length, which is highly variable (MATTIA & LE TOAN 1999).

Two critical points should be mentioned here, which are controversially discussed in the literature: One is the question about how long should be the measurement distance for a accurate estimation of the surface correlation length; the second question is about the shape of the correlation function used for electromagnetic modelling.

Regarding the first question, OH & KAY (1998) demonstrated that the variability of l estimates decreases with profile length, and that a mean estimate of l with a precision of $\pm$ 10 % requires a profile length of 200 l. Their results also illustrate that the correlation length estimates, at short profile lengths, are biased towards values smaller than the true underlying value of l, and that this bias decreases with increasing profile length. Additionally, DAVIDSON et al. (2000) observed for a short profile, of about 1 m, a correlation between s and l, which is not in accordance with the single scale roughness theory. They explained this phenomenon with the relative size of the clods associated with the different tillage states and conclude that this result provides a method of reducing the number of unknowns related to the surface roughness.

The theory of wave scattering from rough surfaces often assumes that surface correlation functions for natural surfaces have mainly two shapes, a Gaussian or an exponential distribution function. Measurements sometimes suggest that surface correlations have an close to the origin exponential shape and change to Gaussian for points further apart (OGLIYV & FOSTER 1989).

Alternatively, to the single scale roughness description, several authors proposes multiscale roughness description. There are two main approaches: the two scale roughness models, with the small and large scale roughness (SHIN & KONG 1984, BAUDOIN et al. 1990) and the excessive number of investigations on random fractals approaches (KELLER et al. 1987, SHEPARD et al. 1995, MATTIA & LE TOAN 1999, DAVIDSON et al. 2000).

1.1.2   Surface Roughness Measurements

The first measurements for the estimation of surface roughness were made mechanically using long and thin steel needles attached on a normalised board. These vertical movable needles are mapping the relief of the ground, which are presented on a scale grid attached on the board (KUIPERS 1957, BURWELL et al. 1963). The distance of the needles was initially varying between 50 to 100 mm. This technique has been improved in order to increase its measurement and operational efficiency. The distance between the needles was reduced down to 5 mm to increase the estimation accuracy, while on the other hand, the digitalisation of

photographs from the needle positions or recording the distance of the vertical movement with a potentialmeter was performed to decrease the evaluation time (CURRENCE & LOVELY 1971). The main advantage of these needle board profilometer techniques is the simple handling and the readily obtained overview of the impact of tillage to the soil surface. The main disadvantage of this technique lies in the disturbance of the soil surface by the needles, preventing the repetition of the measurements. Beyond this, the low spatial resolution, with 5 mm point measurement distance and a vertical resolution of > 1 mm appealed not sufficient for detailed relief investigations. Especially critical is the technical limitation of the length of needle profilometers up to 1.5 m, a fact that makes surface correlation length estimations imprecise.

In the last years, several new methods were developed for measuring surface roughness with high resolution, untouchable and fast. These methods are using optical sensors and are based on photogrammetry or triangulation. The method of photogrammetry consist in transmitting light and measuring its reflection from the surface. These measurement acquisition is fast but the processing of the recorded images and the generation of digital surface elevation models requires a time consuming manual processing (JESCHCKE 1990). Another disadvantage of this method is the disturbing influence of surrounding light as well as the influence of the colour and the moisture content of the soil during the measurement. The laserprofiler method is more robust to surrounding influences, so that the measurements can be collected readily. The transmitted laser pulses are reflected on the soil surface and are recorded with a photodiode. High sampling speed and a high spatial resolution, ranging from 1 to 5 mm with a vertical precision of ± 1.5 mm, are the main advantage of such systems, but the high cost and time effort to construct and install the system are the main disadvantages (HELMING 1992, DAVIDSON 2000). Traditional laser profilers are capable to measure only short profile lengths, from 1 up to 2 m. New generation laser profilers are capable to acquire roughness profiles up to 25 m with a very high location precision (DAVIDSON 2000). Such long profiles are necessary for an accurate estimation of the surface correlation length, and surface characterisation (periodicity, stationarity, homogeneity) while for measuring the RMS height they are of second importance.
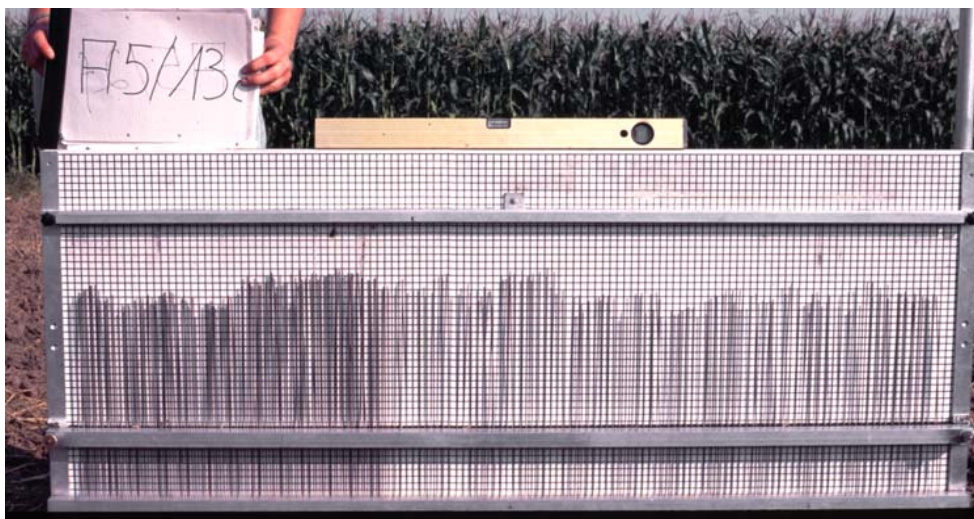
Figure 1 Classical needle profilometer, constructed at the German Aerospace Center, Institute for Radio Technology and Radar Systems, at Oberpfaffenhofen, (DLR-OP-IHR). The needle board is 1.5 m long and 0.60 m high. The distances in between the needles are between 0.5 and 1 cm.

1.2 Material Properties

Soil can be seen as a three phase system of *solid particles*, *soil water* and *soil air* within spatial and time variations of the soil matrix. Soil particles are classified according to the grain size into sand, silt and clay in the descending order. The soil water in the soil matrix represents the porosity part and consists of a portion V_g / V_t (volume of gas/bulk volume of a quantity of soil) occupied by soil air and another V_l / V_t (volume of liquid/total volume of a quantity of soil) occupied by soil water. The amount of water in soil is expressed in

water content, volume fraction $\qquad R = V_l / V_t$ (4)

water content, mass basis $\qquad R_m = m_v / m_s$ (5)

degree of saturation $\qquad S = V_l / (V_l + V_g)$ (6)

Combining (4) and (5), the following relation is obtained for converting from the mass basis to the volume fraction, which is generally more useful in field studies:

$$R_v = R_m \rho_b / \rho_w$$ (7)

In (7), it is assumed that the density of water is unaffected by being adsorbed in soil so that m_v / V_l is equal to ρ_w, the density of pure free water and ρ_b. is the fry bulk density of the soil. The volume fraction, *R*, is equivalent to a depth fraction representing the ratio of the depth of water to the depth of the soil profile that contains it. This form is used when examining gains and loses of water in the field, because precipitation and evaporation are also expressed as depth of waters.

Water content as a volume fraction ranges from zero at oven dryness to a value at pore space saturation. For agronomic and hydrological purposes, two intermediate stages are commonly recognised during the drying of wet soil. The wetter of these is the field capacity, which is the

water content found when a thoroughly wetted soil has drained for about two days. It is determined in the field under conditions that prevent evaporation and allow good drainage. The drier stage, the permanent wilting point, is the water content found when test plants growing on the soil wilt and do not recover if their leaves are kept in a humid atmosphere overnight. Field capacity and the wilting point are used for marking the upper and lower levels of the water of the soil on which water is ordinarily available for plants. As experiments with nonorganic soils show, they both tend to increase with increasing clay in the soil. It could be also shown that at field capacity the degree of saturation of the sandy soil is much lower than that of the clayey soil. This is due to a larger amount of pore space in the sandy soil, which is made up of relatively large pores that drain readily. The conclusion is that the size distribution of pores influences water retention, water movement and aeration, and hence, is often more important than the size distribution of particles (SCHEFER & SCHACHTSCHABEL 1992).

1.2.1   Properties of a Complex Dielectric Constant

The phenomenon by which that non-conducting materials can also be influenced from electrical fields was first observed by M. Faraday, who named them dielectrica. The key parameter that describes the behaviour of a non-conductor in an electrical field is the complex dielectric constant, which is dependent on numerous parameters, such as frequency, temperature, salinity and ferromagnetic substances. Charged particles, under the impact of an external electrical field, are getting out of balance, while the free electrons of a conductor move until the electrical field inside the conductor vanishes. This only partially happens in dieletrica, where the free charges are moving till the back force in the solid body equalises the force affected by the external electrical field. Since, positive and negative particles are contrarily linked, they form electric dipoles. This process is also called dielectric polarisation (MARSHALL et al. 1999). Two main mechanisms can be distinguished, deformation and orientation polarisation, which are responsible for the polarisation of dielectrica. During the orientation of the dipole the energy of a dielectrica is stored. One part of this energy is stored as thermal and the other part is lost due to the internal friction. The dielectrica of a material thus define the amount of electric energy, which can be stored, adsorbed and lead through the medium.

In the electromagnetic wave theory, the real component of the complex dielectric constant is described as refraction or reflection of a wave at the interface between two different media (Snellius Law). The refraction index is a function of the incident angle and the velocity of the transmitted wave, which is related to the refractive angle and the velocity of propagation in the boundary layer of the wave. The refraction index is defined as the square root of the complex dielectric constant of the denser medium and is a dielectric constant, if related to the vacuum or the air. The complex dielectric constant is a measure of the medium response to an electromagnetic wave. This response is composed of two parts, the real and the imaginary (STRATTON 1941, VON HIPPEL 1954), where the complex dielectric constant is given by

$$\varepsilon = \varepsilon' - j\varepsilon''$$

(8)

where ε' is referred to the permittivity of the material, whereas ε'' is referred to the dielectric loss factor of the material and describes the feasibility of a medium to adsorb a wave and to transform its energy into another form. Throughout this dissertation write-up ε' will refers to the average relative dielectric constant of the material. For most natural surfaces $\varepsilon'' \ll \varepsilon'$ (VON HIPPEL 1954).

The attenuation length of electrical field in a given medium is characterised by the imaginary part of its complex dielectric constant. Under the assumption that a propagating wave has exponential attenuation with depth, the penetration depth dp of the wave into the medium is known as the skin depth and is given by (ULABY et al. 1986, SCHANDA 1986):

$$\delta_p = \frac{\lambda \sqrt{\varepsilon'}}{2\pi\varepsilon''} \tag{9}$$

The penetration depth of a slightly lossy medium dp is per definition about 1/e (e = Euler's Number » 2.7183) of the input value attenuated (SCHANDA 1986). From (9), it follows that at longer wavelengths, the penetration depth increases, while at the same time for a fixed wavelength the penetration depth into a medium in general increases with decreasing dielectric constant.

The dielectric constant of most dielectric natural media is between 1 and 6, and increases with increasing water content. Finally, free liquid water (ULABY 1986) has a dielectric constant up to 81 towards low frequencies. This is the reason for the very high sensitivity of microwaves to the moisture contained in the observed bodies, depending on the molecules rotation, e.g. freezing, tight binding to a soil particle, etc., which therefore reduces the e of water.

The behaviour of ε in homogeneous media, such as pure water and ice, are fairly well understood. The frequency dependency of the dielectric constant of pure water is given by the well-known Debye (1929) equation

$$\varepsilon_w = \varepsilon_{w\infty} + \frac{\varepsilon_{w0} - \varepsilon_{w\infty}}{1 + j2\pi f \tau_w} \tag{10}$$

where $\varepsilon_{\omega 0}$ is a static dielectric constant of pure water and $\varepsilon_{\omega\infty}$ presents a high-frequency (or optical) limit of ε_ω, both formulations are dimensionless. τ_ω is the relaxation time of pure water measured in seconds and f the electromagnetic frequency given in Hz (VON HIPPEL 1954).

In case of a dry soil, the real part of the dielectric constant ε', varies over the range between two and four; representative values for the imaginary part lie below 0.05 (ULABY 1986). Since the first water added to dry soil is tightly bound to the surface of the soil particles, it will cause only a small increase of the soil's ε. As more water is added, above the transition value of moisture, the soil's ε rapidly increases due to the behaviour of water molecules. Because the matrix forces, acting on water molecules, decrease rapidly with increasing distance from the soil surface, the water molecules located several layers away from the soil particles are able to move free within the soil (SCHMUGGE 1980).

1.2.2 Relation of Dielectric Constant to Volumetric Soil Moisture

Many empirical and theoretical models have been suggested in the literature to relate the dielectric constant of the mixture to that of its constituents. An extensive investigation in the frequency range between 1.4 and 18 GHz can be found in VON HIPPEL (1954) and in HALLIKAINEN et al. (1985). The behaviour of the experimental measurements have been summarised by polynomial expressions dependent on the volumetric soil water content and the percentage of sand and clay contained in the soil. However, the soil particles have been classified according to USGS (United States Geological Survey) and are, therefore not directly transferable to the European soil classification system. To eliminate the dependence upon adjustable parameters, DOBSON et al. (1985) developed a physical soil model that depends only on measurable soil characteristics and does not require adjustable parameters to fit the experimentally measured data. The model is based on two parameters: the bound water fraction and the free water fraction, according to the pore-size distribution calculated from the particle-size distribution. Studies from TOPP et al. (1980), DE LOOR (1982), SHIVOLA (1989) and STACHEDER (1996) on the permittivity of dielectric mixtures indicate that for a frequency range of 1 to 10 GHz a two component refractive index formula which considers only the volumetric fraction of dry matter and free water is a sufficient good working approach for most soil types. Based on these conclusions, in the following, the developed polynomial relation by TOPP et al. (1980) of the third order was used for the conversion from the volumetric soil water content m_v to the real part of the dielectric constant ε', and vice versa. Measurements and evaluation of the imaginary part ε'' of the complex dielectric constant are not considered in this study because of its almost negligible influence on the total amount of ε.

1.2.3 Soil Water Measurements

There is a large number of methods available to measure directly and indirect the soil moisture content as described in the literature (STACHEDER 1996, PRIETZSCH 1998, MARSHALL 1999). In principle, direct and indirect methods to measure the soil moisture content can be distinguished. The direct methods include all measured processes in which the soil water is removed with evaporation, extraction, or chemical reactions. The indirect methods use the functional relations existing between the physical or chemical properties of the soil matrix and the moisture of the soil. In this study two methods were used, the gravimetric method, which is a direct method, and time domain reflectometry, as indirect approach. Both of them are described in the following.

1.2.3.1 Gravimetrical Method

The most usual direct method to evaluate the water content of a soil sample is to estimate the mass difference before and after drying it in an oven at 150° C until a constant mass is reached. The mass difference m_v corresponds to the water loss of the sample during the drying process. However, the endpoint of the drying processes does not correspond to a completely water-free sample, but to a balanced state between vapour pressure of the material and water vapour partial pressure in the drying region. From this point of view the state for which the

material is called dry, is a subjective term. The water content on a mass basis is described in (5).

The unit of the gravimetric water content is [g/g] and expresses the weight in percent after multiplication with 100 of the soil water [weight %]. If the soil water content is expressed in volumetric percent [vol. %], the bulk density of the soil has to be taken into account. For these, the soil samples are taken with a predefined volume. The water content as a volume fraction R, is obtained from R_m by means of (7), where ρ_b is the dry bulk density of the soil, and ρ_w is the density of the water, which for most applications is expressed as [vol. %] or 1 Mgm^{-3}.
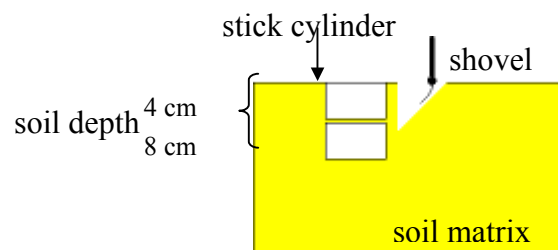

Figure 2   Stick cylinder scheme

This procedure was introduced by GRADNER (1986). Despite arbitrary features of the method, it is the standard with which other estimations of soil moisture content are compared. A disadvantage of the method is that the sampling of experimental areas interferes with continuing experiments. When holes are bored and roots are cut, infiltration and drainage behaviour are effected. A large number of samples may be required because water can be distributed unevenly in the field due to the effect of soil texture and/or structure variability on its retention and movement. Additionally, the method is very time- and work-consuming. Since the amount of water loss, increases with the drying temperature in any inhomogeneous soils that contain clay or organic matter, the oven temperature must be controlled within a range about 100 – 110°C. A variation of about 10 per cent or more has been found in field samples in several experiments as quoted by HOLMES *et al*. (1967). If one consider also the water loss due to evaporation, with a rate of 1-4 mm day $^{-1}$, then the variability of soil makes the moisture sampling approach unattractive.

1.2.3.2   Time Domain Reflectometry

The principle of time domain reflectometry is the measurement of the capacitance of the soil matrix, which depends on the water content of the surrounding medium. The electric capacitance in between two conductors, with the length of 15 cm, placed in the soil depends upon the water content of the soil because the dielectric constant of water (ε' $\approx$ 80) is much larger than that of the dry soil (ε' $\approx$ 5) or the air (ε' $\approx$ 1), which replaces the water as the soil dries. The technique of time domain reflectometry (TDR) circumvents a direct capacitance measurement and is valid for a wide range of soil types (sand and silt). The geometric arrangement for the insertion into the soil is a parallel, two-electrode configuration with the spacing between the insertion rods, that form the transmission line, is 50 mm. Their length determines the depth in the soil to which the pulse is conducted and the result is effectively the volumetric fraction of water in the soil to the chosen depth.

Electric transmission line theory leads to the expression for the propagation velocity of an electromagnetic wave in a slightly long transmission line as

$$V = c[\tfrac{1}{2}\varepsilon'\{1 + (1 + tan^2\delta)^{\frac{1}{2}}\}]^{-\frac{1}{2}} \tag{11}$$

where c is the speed of light, and $tan\ \delta = \{\varepsilon'' + (\sigma_{dc}/\omega\varepsilon_0)\}\ /\varepsilon'$. The remaining symbols are explained as: ε' is the real part of the complex dielectric constant, ε'' is the imaginary part, σ_{dc} is the zero-frequency electrical conductivity of the medium, ω is the angular frequency, and ε_0 is the free-space permittivity. Measurement at a very high frequency makes $tan\ \delta$ tend to zero, so that

$$V = c\ /\ (\varepsilon')^{\frac{1}{2}} \tag{12}$$

This leads to

$$t = 2L\varepsilon^{\frac{1}{2}}\ /\ c\ \text{and} \tag{13}$$

And

$$\varepsilon = (ct\ /\ 2L)^2 \tag{14}$$

Here t is the time taken for the propagation of an electromagnetic pulse launched along a transmission line of length L and reflected back to the origin. The variation of the dielectric constant of soil with the water content, R, can approximately be estimated by adding the three contributions $R\varepsilon$ (water), $\phi\varepsilon$(air) and $[1 - (R + \phi)]\varepsilon$ (soil), where R is the volume fraction of water, ϕ the same for the air-filled pores, and $[1 - (R + \phi)]$ is the volume fraction of the soil. In principle, then it is to be expected that R would have a functional dependence upon ε, as determined for the soil in question. TOPP *et al.* (1980) established such a relationship to be

$$R = -a + b\varepsilon - c\varepsilon^2 + d\varepsilon^3 \tag{15}$$

Volumetric soil moisture content m_v

$$m_v = -5.3\ 10^{-2} + 2.92\ 10^{-2}\ \varepsilon' - 5.5\ 10^{-4}\ \varepsilon'^2 + 4.3\ 10^{-6}\ \varepsilon'^3 \tag{16}$$

Real part ε' of the complex dielectric constant ε

$$\varepsilon' = 3.03 + 9.3\ m_v + 146\ m_v^2 - 76.7\ m_v^3. \tag{17}$$

The experimental values of the variables were determined for ε by pulse travel-times, (14), and for R by gravimetric measurements and the use of dry bulk densities. This empirically determined third order polynomial expression for the dielectric constant is independent of type, bulk density, texture, salinity and temperature of the soil and is confirmed in several investigations (DALTON & VAN GNEUCHTEN 1986, WHALLEY 1993). Thus, up to now this function is used in several studies and is regarded as an universal calibration function even if actual investigations indicate a limited validity for all soil types valid. HERKELRATH *et al.* (1991), for example, received significantly divergent results for soils with organic matter. After intensive investigations with inorganic and organic soil types, ROTH *et al.* (1992) suggest to use two different calibration functions - for inorganic and organic soil.
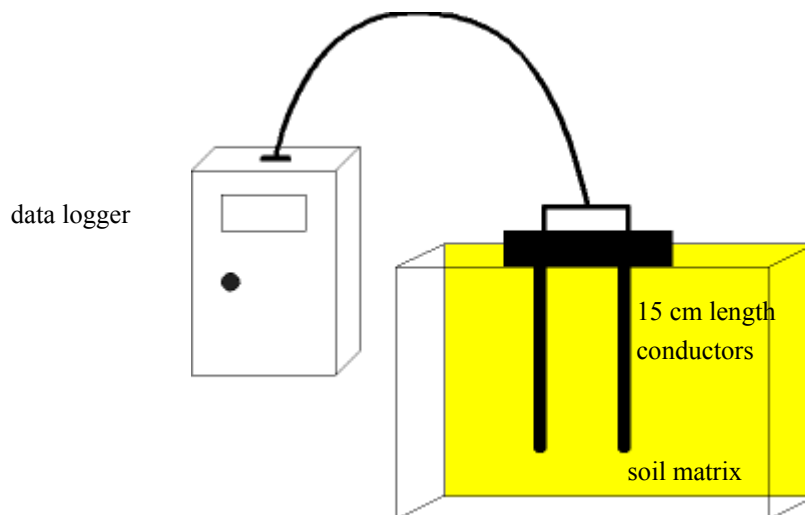
Figure 3 Time domaine reflectometry scheme

The accuracy of the obtained time domain measurements is varying about of ± 1.3 [vol %]. In principle, all studies notice a slight underestimation of soil water content for clay enriched soil and for soil mixed with high organic matter. Some authors explained this phenomenon with dielectric properties of bounded water (DOBSON 1985, HALLIKAINEN 1985). Minerals and swelling clay have a high surface tension and are for this reason able to adsorb a high amount of water. Hydration of the exchangeable cations is largely responsible for this absorption and for the accompanying increase in the interlayer spacing. As consequence of this process, the minerals and clay swell macroscopically. With continuing absorption, it becomes plastic and it is further explained, as follows, by the development of a diffuse layer of ions. If the solution midway between two parallel surfaces has a higher electrolyte concentration than that of an outer solution bathing the clay, water will be attracted osmotically, and the distance between the surfaces will consequently increase. The exchangeable cations, unable to move freely out to the bathing solution, act as being retained by a semi-permeable membrane. The water dipoles are strongly bounded so that they no longer rotate or polarise. Thus, the dielectric constant of bounded water is significantly smaller than of free water and cannot anymore be measured using time domain reflectometry. This is in contrary to the oven drying methods where the bounding forces can be dissolved in the water. Therefore, the use of an universal calibration function may lead to an underestimation of soil water content. Using an universal calibration function for the measurement of soil moisture content should be critically observed. For some soil types, especially mineral and clay rich soil, the use of a soil specific calibration function is needed.