CHAPTER 11

MEASUREMENT OF SOIL MOISTURE

11.1   General

Soil moisture is an important component in the atmospheric water cycle, both on a small agricultural scale and in large-scale modelling of land/atmosphere interaction. Vegetation and crops always depend more on the moisture available at root level than on precipitation occurrence. Water budgeting for irrigation planning, as well as the actual scheduling of irrigation action, requires local soil moisture information. Knowledge of the degree of soil wetness helps to understand the initiation of convective events, and to forecast the risk of flash floods, or the occurrence of fog.

Nevertheless, soil moisture has been seldom observed routinely at meteorological stations. Documentation of soil wetness was usually restricted to the description of the "state of the ground" by means of WMO Code Tables 0901 and 0975, and its measurement was left to hydrologists, agriculturalists and other actively interested parties. Around 1990, the interest of meteorologists in soil moisture measurement increased. This was partly because, after the pioneering work by Deardorff (1978), numerical atmosphere models at various scales became more adept at handling fluxes of sensible and latent heat in soil surface layers. Moreover, newly developed soil moisture measurement techniques are more feasible for meteorological stations than most of the classic methods.

To satisfy the increasing need for determining soil moisture status, the most commonly used methods and instruments will be discussed, including their advantages and disadvantages. Some less common observation techniques are also mentioned. This chapter discusses both in situ and remote sensing soil moisture measurements. Space-based remote sensing is included, complemented by information in Part IV of this Guide.

11.1.1   Definitions

Soil moisture determinations measure either the soil water content or the soil water potential.

Soil water content

Soil water content is an expression of the mass or volume of water in the soil, while the soil water potential is an

expression of the soil water energy status. The relation between content and potential is not universal and depends on the characteristics of the local soil, such as soil density and soil texture.

Soil water content on the basis of mass is expressed in the gravimetric soil moisture content, θ_g, defined by:

$$\theta_g = M_{water}/M_{soil} \qquad (11.1)$$

where M_{water} is the mass of the water in the soil sample and M_{soil} is the mass of dry soil that is contained in the sample. Values of θ_g in meteorology are usually expressed in per cent.

Because precipitation, evapotranspiration and solute transport variables are commonly expressed in terms of flux, volumetric expressions for water content are often more useful. The volumetric soil moisture content of a soil sample, θ_v, is defined as:

$$\theta_v = V_{water}/V_{sample} \qquad (11.2)$$

where V_{water} is the volume of water in the soil sample and V_{sample} is the total volume of dry soil + air + water in the sample. Again, the ratio is usually expressed in per cent. However, recently, the standard to express soil moisture many research communities are adopting now is volumetric water content (m^3/m^3) instead of percentage. The relationship between gravimetric and volumetric moisture contents is:

$$\theta_v = \theta_g \left(\rho_b/\rho_w \right) \qquad (11.3)$$

where ρ_b is the dry soil bulk density and ρ_w is the soil water density.

The basic technique for measuring soil water content is the gravimetric method, described below in section 11.2. Because this method is based on direct measurements, it is the standard with which all other methods are compared. Unfortunately, gravimetric sampling is destructive, rendering repeat measurements on the same soil sample impossible. Because of the difficulties of accurately measuring dry soil and water volumes, volumetric water contents are not usually determined directly. Recently, the standard to express soil moisture many research communities are adopting now is volumetric water content (m^3/m^3) instead of percentage.

Soil water potential

Soil water potential describes the energy status of the soil water and is an important parameter for water transport analysis, water storage estimates and soil-plant-water relationships. A difference in water potential between two soil locations indicates a tendency for water flow, from high to low potential. When the soil is drying, the water potential becomes more negative and the work that must be done to extract water from the soil increases. This makes water uptake by plants more difficult, so the water potential in the plant drops, resulting in plant stress and, eventually, severe wilting.

Formally, the water potential is a measure of the ability of soil water to perform work, or, in the case of negative potential, the work required to remove the water from the soil. The total water potential ψ_t, the combined effect of all force fields, is given by:

$$\psi_t = \psi_z + \psi_m + \psi_o + \psi_p \qquad (11.4)$$

where ψ_z is the gravitational potential, based on elevation above the mean sea level; ψ_m is the matric potential, suction due to attraction of water by the soil matrix; ψ_o is the osmotic potential, due to energy effects of solutes in water; and ψ_p is the pressure potential, the hydrostatic pressure below a water surface.

The potentials which are not related to the composition of water or soil are together called hydraulic potential, ψ_h. In saturated soil, this is expressed as $\psi_h = \psi_z + \psi_p$, while in unsaturated soil, it is expressed as $\psi_h = \psi_z + \psi_m$. When the phrase "water potential" is used in studies, maybe sometimes with the notation ψ_w, it is advisable to check the author's definition because this term has been used for $\psi_m + \psi_z$ as well as for $\psi_m + \psi_o$.

The gradients of the separate potentials will not always be significantly effective in inducing flow. For example, ψ_0 requires a semi-permeable membrane to induce flow, and ψ_p will exist in saturated or ponded conditions, but most practical applications are in unsaturated soil.

11.1.2   Units

In solving the mass balance or continuity equations for water, it must be remembered that the components of water content parameters are not dimensionless. Gravimetric water content is the weight of soil water contained in a unit weight of soil (kg water/kg dry soil). Likewise, volumetric water content is a volume fraction (m^3 water/m^3 soil).

The basic unit for expressing water potential is energy (in joules = kg m^2 s^{-2}) per unit mass, J kg^{-1}. Alternatively, energy per unit volume (J m^{-3}) is equivalent to pressure, expressed in pascals (Pa = kg m^{-1} s^{-2}). Units encountered in older literature are bar (= 100 kPa), atmosphere (= 101.32 kPa), or pounds per

square inch (= 6.895 kPa). A third class of units are those of pressure head in (centi)metres of water or mercury, energy per unit weight. The relation of the three potential unit classes is:

$$\psi \text{ (J kg}^{-1}) = \gamma \cdot \underline{X} \; \psi \text{ (Pa)} = [\psi \text{ (m)}] \, / \, g \quad (11.5)$$

where $\gamma = 10^3$ kg m^{-3} (density of water) and g = 9.81m s^{-2} (gravity acceleration). Because the soil water potential has a large range, it is often expressed logarithmically, usually in pressure head of water. A common unit for this is called pF, and is equal to the base –10 logarithm of the absolute value of the head of water expressed in centimetres.

11.1.3 Meteorological requirements

Soil consists of individual particles and aggregates of mineral and organic materials, separated by spaces or pores which are occupied by water and air. The relative amount of pore space decreases with increasing soil grain size (intuitively one would expect the opposite). The movement of liquid water through soil depends upon the size, shape and generally the geometry of the pore spaces.

If a large quantity of water is added to a block of otherwise "dry" soil, some of it will drain away rapidly by the effects of gravity through any relatively large cracks and channels. The remainder will tend to displace some of the air in the spaces between particles, the larger pore spaces first. Broadly speaking, a well-defined "wetting front" will move downwards into the soil, leaving an increasingly thick layer retaining all the moisture it can hold against gravity. That soil layer is then said to be at "field capacity", a state that for most soils occurs about $\psi_m \approx$ 10 -33 kPa J/kg with a range of values from -1 J/kg for organic soils to -100 J/kg for heavy clay soils. A value of -10 J/kg ((pF $\approx$ 2) can be assigned for a loamy sand soil.(pF ≈ 2). This state must not be confused with the undesirable situation of "saturated" soil, where all the pore spaces are occupied by water. After a saturation event, such as heavy rain, the soil usually needs at least 24 h to reach field capacity. When moisture content falls below field capacity, the subsequent limited movement of water in the soil is partly liquid, partly in the vapour phase by distillation (related to temperature gradients in the soil), and sometimes by transport in plant roots.

Plant roots within the block will extract liquid water from the water films around the soil particles with which they are in contact. The rate at which this extraction is possible depends on the soil moisture potential. A point is reached at which the forces holding moisture films to soil particles cannot be overcome by root suction, and plants are starved of water and lose turgidity: soil moisture has reached the "wilting point", which in most cases occurs at a soil water potential of –1.5 MPa (pF = 4.2). In agriculture, the soil water available to plants is commonly taken to be the quantity between field capacity and the wilting point, and this varies highly between soils: in sandy soils it may be less than 10 volume per cent, while in soils with much organic matter it can be over 40 volume per cent.

Usually it is desirable to know the soil moisture content and potential as a function of depth. Evapotranspiration models concern mostly a shallow depth (tens of centimetres); agricultural applications need moisture information at root depth (order of a metre); and atmospheric general circulation models incorporate a number of layers down to a few metres. For hydrological and water-balance needs – such as catchment-scale runoff models, as well as for effects upon soil properties such as soil mechanical strength, thermal conductivity and diffusivity – information on deep soil water content is needed. The accuracy needed in water content determinations and the spatial and temporal resolution required vary by application. An often-occurring problem is the inhomogeneity of many soils, meaning that a single observation location cannot provide absolute knowledge of the regional soil moisture, but only relative knowledge of its change.

11.1.4 Measurement methods

The methods and instruments available to evaluate soil water status may be classified in three ways. First, a distinction is made between the determination of water content and the determination of water potential. Second, a so-called direct method requires the availability of sizeable representative terrain from which large numbers of soil samples can be taken for destructive evaluation in the laboratory. Indirect methods use an instrument placed in the soil to measure some soil property related to soil moisture. Third, methods can be ranged according to operational applicability, taking into account the regular labour involved, the degree of dependence on laboratory availability, the complexity of the operation and the reliability of the result. Moreover, the preliminary costs of acquiring instrumentation must be compared with the subsequent costs of local routine observation and data processing.

Reviews such as WMO (1968; 1989; 2001) and Schmugge, Jackson and McKim (1980) are very useful for learning about practical problems, but dielectric measurement methods were only developed well after 1980, so too-early reviews should not be relied on much when choosing an operational method.

There are fourfive operational alternatives for the determination of soil water content. First, there is classic gravimetric moisture determination, which is a simple direct method. Second, there is lysimetry, a non-destructive variant of gravimetric measurement. A container filled with soil is weighed either occasionally or continuously to indicate changes in total mass in the container, which may in part or totally be due to changes in soil moisture (lysimeters are discussed in more detail in Part I, Chapter 10). Third, water content may be determined indirectly

by various radiological techniques, such as neutron scattering and gamma absorption. Fourth, water content can be derived from the dielectric properties of soil, for example, by using time-domain reflectometry. Lastly, soil moisture can be ~~measured~~inferred on a global scale ~~by way of remote sensing.~~from remotely sensed measurements of the earth's thermal or reflective properties.

Soil water potential measurement can be performed by several indirect methods, in particular using tensiometers, resistance blocks and soil psychrometers. None of these instruments is effective at this time over the full range of possible water potential values. For extended study of all methods of various soil moisture measurements, up-to-date handbooks are provided by Klute (1986), Dirksen (1999), and Smith and Mullins (referenced here as Gardner and others, 2001, and Mullins, 2001).

11.2   Gravimetric direct measurement of soil water content

The gravimetric soil moisture content θ_g is typically determined directly. Soil samples of about 50 g are removed from the field with the best available tools (shovels, spiral hand augers, bucket augers, perhaps power-driven coring tubes), disturbing the sample soil structure as little as possible (Dirksen, 1999). The soil sample should be placed immediately in a leak-proof, seamless, pre-weighed and identified container. As the samples will be placed in an oven, the container should be able to withstand high temperatures without melting or losing significant mass. The most common soil containers are aluminium cans, but non-metallic containers should be used if the samples are to be dried in microwave ovens in the laboratory. If soil samples are to be transported for a considerable distance, tape should be used to seal the container to avoid moisture loss by evaporation.

The samples and container are weighed in the laboratory both before and after drying, the difference being the mass of water originally in the sample. The drying procedure consists in placing the open container in an electrically heated oven at 105°C until the mass stabilizes at a constant value. The drying times required usually vary between 16 and 24 h. Note that drying at 105°±5°C is part of the usually accepted definition of "soil water content", originating from the aim to measure only the content of "free" water which is not bound to the soil matrix (Gardner and others, 2001).

If the soil samples contain considerable amounts of organic matter, excessive oxidation may occur at 105°C and some organic matter will be lost from the sample. Although the specific temperature at which excessive oxidation occurs is difficult to specify, lowering the oven temperature from 105 to 70°C seems to be sufficient to avoid significant loss of organic matter, but this can lead to water content values that are too low. Oven temperatures and drying times should be checked and reported.

Microwave oven drying for the determination of gravimetric water contents may also be used effectively (Gee and Dodson, 1981). In this method, soil water temperature is quickly raised to boiling point, then remains constant for a period due to the consumption of heat in vaporizing water. However, the temperature rapidly rises as soon as the energy absorbed by the soil water exceeds the energy needed for vaporizing the water. Caution should be used with this method, as temperatures can become high enough to melt plastic containers if stones are present in the soil sample.

Gravimetric soil water contents for air-dry (25°C) mineral soil are often less than 2 per cent, but, as the soil approaches saturation, the water content may increase to values between 25 and 60 per cent, depending on soil type. Volumetric soil water content, θ_v, may range from less than 10 per cent for air-dry soil to between 40 and 50 per cent for mineral soils approaching saturation. Soil θ_v determination requires measurement of soil density, for example, by coating a soil clod with paraffin and weighing it in air and water, or some other method (Campbell and Henshall, 2001).

Water contents for stony or gravelly soils can be grossly misleading. When rocks occupy an appreciable volume of the soil, they modify direct measurement of soil mass, without making a similar contribution to the soil porosity. For example, gravimetric water content may be 10 per cent for a soil sample with a bulk density of 2~~ ~~000 kg m^{-3}; however, the water content of the same sample based on finer soil material (stones and gravel excluded) would be 20 per cent~~,~~ if the bulk density of fine soil material was 1 620 kg m^{-3}.

Although the gravimetric water content for the finer soil fraction, $\theta_{g,fines}$, is the value usually used for spatial and temporal comparison, there may also be a need to determine the volumetric water content for a gravelly soil. The latter value may be important in calculating the volume of water in a root zone. The relationship between the gravimetric water content of the fine soil material and the bulk volumetric water content is given by:

$$\theta_{v,stony} = \theta_{g,fines} \, (\rho_b / \rho_v)(1 + M_{stones}/M_{fines}) \qquad (11.6)$$

where $\theta_{v,stony}$ is the bulk volumetric water content of soil containing stones or gravel and M_{stones} and M_{fines} are the masses of the stone and fine soil fractions (Klute, 1986).

11.3   Soil water content: indirect methods

The capacity of soil to retain water is a function of soil texture and structure. When removing a soil sample, the soil

being evaluated is disturbed, so its water-holding capacity is altered. Indirect methods of measuring soil water are helpful as they allow information to be collected at the same location for many observations without disturbing the soil water system. Moreover, most indirect methods determine the volumetric soil water content without any need for soil density determination.

11.3.1 Radiological methods

Two different radiological methods are available for measuring soil water content. One is the widely used neutron scatter method, which is based on the interaction of high-energy (fast) neutrons and the nuclei of hydrogen atoms in the soil. The other method measures the attenuation of gamma rays as they pass through soil. Both methods use portable equipment for multiple measurements at permanent observation sites and require careful calibration, preferably with the soil in which the equipment is to be used.

When using any radiation-emitting device, some precautions are necessary. The manufacturer will provide a shield that must be used at all times. The only time the probe leaves the shield is when it is lowered into the soil access tube. When the guidelines and regulations regarding radiation hazards stipulated by the manufacturers and health authorities are followed, there is no need to fear exposure to excessive radiation levels, regardless of the frequency of use. Nevertheless, whatever the type of radiation-emitting device used, the operator should wear some type of film badge that will enable personal exposure levels to be evaluated and recorded on a monthly basis.

11.3.1.1 Neutron scattering method

In neutron soil moisture detection (Visvalingam and Tandy, 1972; Greacen, 1981), a probe containing a radioactive source emitting high-energy (fast) neutrons and a counter of slow neutrons is lowered into the ground. The hydrogen nuclei, having about the same mass as neutrons, are at least 10 times as effective for slowing down neutrons upon collision as most other nuclei in the soil. Because in any soil most hydrogen is in water molecules, the density of slow "thermalized" neutrons in the vicinity of the neutron probe is nearly proportional to the volumetric soil water content.

Some fraction of the slowed neutrons, after a number of collisions, will again reach the probe and its counter. When the soil water content is large, not many neutrons are able to travel far before being thermalized and ineffective, and then 95 per cent of the counted returning neutrons come from a relatively small soil volume. In wet soil, the "radius of influence" may be only 15 cm, while in dry soil that radius may increase to 50 cm. Therefore, the measured soil volume varies with water content, and thin layers cannot be resolved. This method is hence less suitable to localize water-content discontinuities, and it cannot be used effectively in the top 20 cm of soil on account of the soil air discontinuity.

Several source and detector arrangements are possible in a neutron probe, but it is best to have a probe with a double detector and a central source, typically in a cylindrical container. Such an arrangement allows for a nearly spherical zone of influence and leads to a more linear relation of neutron count to soil water content.

A cable is used to attach a neutron probe to the main instrument electronics, so that the probe can be lowered into a previously installed access tube. The access tube should be seamless and thick enough (at least 1.25 mm) to be rigid, but not so thick that the access tube itself slows neutrons down significantly. The access tube must be made of non-corrosive material, such as stainless steel, aluminium or plastic, although polyvinylchloride should be avoided as it absorbs slow neutrons. Usually, a straight tube with a diameter of 5 cm is sufficient for the probe to be lowered into the tube without a risk of jamming. Care should be taken in installing the access tube to ensure that no air voids exist between the tube and the soil matrix. At least 10 cm of the tube should extend above the soil surface, in order to allow the box containing the electronics to be mounted on top of the access tube. All access tubes should be fitted with a removable cap to keep rainwater from entering the tubes.

In order to enhance experimental reproducibility, the soil water content is not derived directly from the number of slow neutrons detected, but rather from a count ratio (CR), given by:

$$CR = C_{soil}/C_{background} \qquad (11.7)$$

where C_{soil} is the count of thermalized neutrons detected in the soil and $C_{background}$ is the count of thermalized neutrons in a reference medium. All neutron probe instruments now come with a reference standard for these background calibrations, usually against water. The standard in which the probe is placed should be at least 0.5 m in diameter so as to represent an "infinite" medium. Calibration to determine $C_{background}$ can be done by a series of ten 1 min readings, to be averaged, or by a single 1 h reading. C_{soil} is determined from averaging several soil readings at a particular depth/location. For calibration purposes, it is best to take three samples around the access tube and to average the water contents corresponding to the average CR calculated for that depth. A minimum of five different water contents should be evaluated for each depth. Although some calibration curves may be similar, a separate calibration for each depth should be conducted. The lifetime of most probes is more than 10 years.

11.3.1.2 Gamma-ray attenuation

Whereas the neutron method measures the volumetric water content in a large sphere, gamma-ray absorption scans a thin layer. The dual-probe gamma device is nowadays mainly used in the laboratory since dielectric methods became operational for field use. Another reason for this is that gamma rays are more dangerous to work with than neutron scattering devices, as well as the fact that the operational costs for the gamma rays are relatively high.

Changes in gamma attenuation for a given mass absorption coefficient can be related to changes in total soil density. As the attenuation of gamma rays is due to mass, it is not possible to determine water content unless the attenuation of gamma rays due to the local dry soil density is known and remains unchanged with changing water content. Determining accurately the soil water content from the difference between the total and dry density attenuation values is therefore not simple.

Compared to neutron scattering, gamma-ray attenuation has the advantage of allowing accurate measurements at a few centimetres below the air-surface interface. Although the method has a high degree of resolution, the small soil volume evaluated will exhibit more spatial variation due to soil heterogeneities (Gardner and Calissendorff, 1967).

11.3.2   Soil water dielectrics

When a medium is placed in the electric field of a capacitor or waveguide, its influence on the electric forces in that field is expressed as the ratio between the forces in the medium and the forces which would exist in vacuum. This ratio, called permittivity or "dielectric constant", is for liquid water about 20 times larger than that of average dry soil, because water molecules are permanent dipoles. The dielectric properties of ice, and of water bound to the soil matrix, are comparable to those of dry soil. Therefore, the volumetric content of free soil water can be determined from the dielectric characteristics of wet soil by reliable, fast, non-destructive measurement methods, without the potential hazards associated with radioactive devices. Moreover, such dielectric methods can be fully automated for data acquisition. At present, two methods which evaluate soil water dielectrics are commercially available and used extensively, namely time-domain reflectometry and frequency-domain measurement.

11.3.2.1   Time-domain reflectometry

Time-domain reflectometry is a method which determines the dielectric constant of the soil by monitoring the travel of an electromagnetic pulse, which is launched along a waveguide formed by a pair of parallel rods embedded in the soil. The pulse is reflected at the end of the waveguide and its propagation velocity, which is inversely proportional to the square root of the dielectric constant, can be measured well by actual electronics.

The most widely used relation between soil dielectrics and soil water content was experimentally summarized by Topp, Davis and Annan (1980) as follows:

$$\theta_v = -0.053 + 0.029 \, \varepsilon - 5.5 \cdot 10^{-4} \, \varepsilon^2 + 4.3 \cdot 10^{-6} \, \varepsilon^3 \quad (11.8)$$

where ε is the dielectric constant of the soil water system. This empirical relationship has proved to be applicable in many soils, roughly independent of texture and gravel content (Drungil, Abt and Gish, 1989). However, soil-specific calibration is desirable for soils with low density or with a high organic content. For complex soil mixtures, the De Loor equation has proved useful (Dirksen and Dasberg, 1993).

Generally, the parallel probes are separated by 5 cm and vary in length from 10 to 50 cm; the rods of the probe can be of any metallic substance. The sampling volume is essentially a cylinder of a few centimetres in radius around the parallel probes (Knight, 1992). The coaxial cable from the probe to the signal-processing unit should not be longer than about 30 m. Soil water profiles can be obtained from a buried set of probes, each placed horizontally at a different depth, linked to a field data logger by a multiplexer.

11.3.2.2   Frequency-domain measurement

While time-domain refletometryreflectometry uses microwave frequencies in the gigahertz range, frequency-domain sensors measure the dielectric constant at a single microwave megahertz frequency. The microwave dielectric probe utilizes an open-ended coaxial cable and a single reflectometer at the probe tip to measure amplitude and phase at a particular frequency. Soil measurements are referenced to air, and are typically calibrated with dielectric blocks and/or liquids of known dielectric properties. One advantage of using liquids for calibration is that a perfect electrical contact between the probe tip and the material can be maintained (Jackson, 1990).

As a single, small probe tip is used, only a small volume of soil is ever evaluated, and soil contact is therefore critical. As a result, this method is excellent for laboratory or point measurements, but is likely to be subject to spatial variability problems if used on a field scale (Dirksen, 1999).

11.4   Soil water potential instrumentation

The basic instruments capable of measuring matric potential are sufficiently inexpensive and reliable to be used in field-scale monitoring programmes. However, each instrument has a limited accessible water potential range. For example, tTensiometers work well only in wet soil, while resistance blocks do better in moderately dry soil.

11.4.1 Tensiometers

The most widely used and least expensive water potential measuring device is the tensiometer. Tensiometers are simple instruments, usually consisting of a porous ceramic cup and a sealed plastic cylindrical tube connecting the porous cup to some pressure-recording device at the top of the cylinder. They measure the matric potential, because solutes can move freely through the porous cup.

The tensiometer establishes a quasi-equilibrium condition with the soil water system. The porous ceramic cup acts as a membrane through which water flows, and therefore must remain saturated if it is to function properly. Consequently, all the pores in the ceramic cup and the cylindrical tube are initially filled with de-aerated water. Once in place, the tensiometer will be subject to negative soil water potentials, causing water to move from the tensiometer into the surrounding soil matrix. The water movement from the tensiometer will create a negative potential or suction in the tensiometer cylinder which will register on the recording device. For recording, a simple U-tube filled with water and/or mercury, a Bourdon-type vacuum gauge or a pressure transducer (Marthaler and others, 1983) is suitable.

If the soil water potential increases, water moves from the soil back into the tensiometer, resulting in a less negative water potential reading. This exchange of water between the soil and the tensiometer, as well as the tensiometer's exposure to negative potentials, will cause dissolved gases to be released by the solution, forming air bubbles. The formation of air bubbles will alter the pressure readings in the tensiometer cylinder and will result in faulty readings. Another limitation is that the tensiometer has a practical working limit of $\Psi \approx -85$ kPa. Beyond -100 kPa (≈ 1 atm), water will boil at ambient temperature, forming water vapour bubbles which destroy the vacuum inside the tensiometer cylinder. Consequently, the cylinders occasionally need to be de-aired with a hand-held vacuum pump and then refilled.

Under drought conditions, appreciable amounts of water can move from the tensiometer to the soil. Thus, tensiometers can alter the very condition they were designed to measure. Additional proof of this process is that excavated tensiometers often have accumulated large numbers of roots in the proximity of the ceramic cups. Typically, when the tensiometer acts as an "irrigator", so much water is lost through the ceramic cups that a vacuum in the cylinder cannot be maintained, and the tensiometer gauge will be inoperative.

Before installation, but after the tensiometer has been filled with water and degassed, the ceramic cup must remain wet. Wrapping the ceramic cup in wet rags or inserting it into a container of water will keep the cup wet during transport from the laboratory to the field. In the field, a hole of the appropriate size and depth is prepared. The hole should be large enough to create a snug fit on all sides, and long enough so that the tensiometer extends sufficiently above the soil surface for de-airing and refilling access. Since the ceramic cup must remain in contact with the soil, it may be beneficial in stony soil to prepare a thin slurry of mud from the excavated site and to pour it into the hole before inserting the tensiometer. Care should also be taken to ensure that the hole is backfilled properly, thus eliminating any depressions that may lead to ponded conditions adjacent to the tensiometer. The latter precaution will minimize any water movement down the cylinder walls, which would produce unrepresentative soil water conditions.

Only a small portion of the tensiometer is exposed to ambient conditions, but its interception of solar radiation may induce thermal expansion of the upper tensiometer cylinder. Similarly, temperature gradients from the soil surface to the ceramic cup may result in thermal expansion or contraction of the lower cylinder. To minimize the risk of temperature-induced false water potential readings, the tensiometer cylinder should be shaded and constructed of non-conducting materials, and readings should be taken at the same time every day, preferably in the early morning.

A new development is the osmotic tensiometer, where the tube of the meter is filled with a polymer solution in order to function better in dry soil. For more information on tensiometers see Dirksen (1999) and Mullins (2001).

11.4.2 Resistance blocks

Electrical resistance blocks, although insensitive to water potentials in the wet range, are excellent companions to the tensiometer. They consist of electrodes encased in some type of porous material that within about two days will reach a quasi-equilibrium state with the soil. The most common block materials are nylon fabric, fibreglass and gypsum, with a working range of about -50 kPa (for nylon) or -100 kPa (for gypsum) up to $-1-500$ kPa. Typical block sizes are 4 cm × 4 cm × 1 cm. Gypsum blocks last a few years, but less in very wet or saline soil (Perrier and Marsh, 1958).

This method determines water potential as a function of electrical resistance, measured with an alternating current bridge (usually $\approx 1-000$ Hz) because direct current gives polarization effects. However, resistance decreases if soil is saline, falsely indicating a wetter soil. Gypsum blocks are less sensitive to soil saltiness effects because the electrodes are consistently exposed to a saturated solution of calcium sulphate. The output of gypsum blocks must

be corrected for temperature (Aggelides and Londra, 1998).

Because resistance blocks do not protrude above the ground, they are excellent for semi-permanent agricultural networks of water potential profiles, if installation is careful and systematic (WMO, 2001). When installing the resistance blocks it is best to dig a small trench for the lead wires before preparing the hole for the blocks, in order to minimize water movement along the wires to the blocks. A possible field problem is that shrinking and swelling soil may break contact with the blocks. On the other hand, resistance blocks do not affect the distribution of plant roots.

Resistance blocks are relatively inexpensive. However, they need to be calibrated individually. This is generally accomplished by saturating the blocks in distilled water and then subjecting them to a predetermined pressure in a pressure-plate apparatus (Wellings, Bell and Raynor, 1985), at least at five different pressures before field installation. Unfortunately, the resistance is less on a drying curve than on a wetting curve, thus generating hysteresis errors in the field because resistance blocks are slow to equilibrate with varying soil wetness (Tanner and Hanks, 1952). As resistance-block calibration curves change with time, they need to be calibrated before installation and to be checked regularly afterwards, either in the laboratory or in the field.

11.4.3   Psychrometers

Psychrometers are used in laboratory research on soil samples as a standard for other techniques (Mullins, 2001), but a field version is also available, called the Spanner psychrometer (Rawlins and Campbell, 1986). This consists of a miniature thermocouple placed within a small chamber with a porous wall. The thermocouple is cooled by the Peltier effect, condensing water on a wire junction. As water evaporates from the junction, its temperature decreases and a current is produced which is measured by a meter. Such measurements are quick to respond to changes in soil water potential, but are very sensitive to temperature and salinity (Merrill and Rawlins, 1972).

The lowest water potential typically associated with active plant water uptake corresponds to a relative humidity of between 98 and 100 per cent. This implies that, if the water potential in the soil is to be measured accurately to within 10 kPa, the temperature would have to be controlled to better than 0.001 K. This means that the use of field psychrometers is most appropriate for low matric potentials, of less than –300 kPa. In addition, the instrument components differ in heat capacities, so diurnal soil temperature fluctuations can induce temperature gradients in the psychrometer (Brunini and Thurtell, 1982). Therefore, Spanner psychrometers should not be used at depths of less than 0.3 m, and readings should be taken at the same time each day, preferably in the early morning. In summary, soil psychrometry is a difficult and demanding method, even for specialists.

11.5   Site selection and sample size

There is no standard depth or measurement interval at which soil moisture observations are taken, since this strongly depends on the research objectives for which the sensors are installed. The International Soil Moisture Network (Dorigo et al. 2011) provides an extensive database with harmonised in situ soil moisture time series of networks distributed all over the world. Here data is harmonised to a half-hourly measurement interval whenever possible. Most networks and stations that are incorporated within the ISMN measure soil moisture at several depths starting at 0.05 m and up to a depth of 0.50 m or 1.00 m. By measuring soil moisture at different depths the behaviour of soil moisture at different depths can be compared and used to validate measurements. Measurements of other meteorological parameters is very valuable for soil moisture measurements. For example, precipitation data at the measurement site provides a validation for the soil moisture data.

Representativity of any soil moisture observation point is limited because of the high probability of significant variations, both horizontally and vertically, of soil structure (porosity, density, chemical composition), land cover and relief. It is pivotal that soil moisture and its variability are captured on the scale necessary for studies focussing on hydrological processes and for satellite validation. Gravimetric water content determinations or indirect measurements of soil moisture are only reliable at the point of measurement, making a large number of samples necessary to describe adequately the soil moisture status of the site. To estimate the number of samples n needed at a local site to estimate soil water content at an observed level of accuracy (L), the sample size can be estimated from:

$$n = 4 \, (\sigma^2/L^2) \qquad (11.12)$$

where σ^2 is the sample variance generated from a preliminary sampling experiment. For example, suppose that a preliminary sampling yielded a (typical) σ^2 of 25 per cent and the accuracy level needed to be within 3 per cent, 12 samples would be required from the site (if it can be assumed that water content is normally distributed across the site). A study of Brocca et al. (2007) showed that the minimum number of point samples needed for an area in Central Italy with an extent of around 9 to 8800 m2 varies between 15 and 35 samples. The higher number of samples was needed for sites with more significant relief. Famiglietti et al. (2008) found that a number of 30 samples are sufficient for a footprint of 50 km, taking into account that the data is independent and spatially uncorrelated.

Upscaling of the point measurements obtained by gravimetric water content determination or indirect

measurements with in situ sensors has been the subject of many studies. Upscaling techniques vary to relatively straightforward interpolation and time/rank stability techniques to more complicated techniques such as statistical transformations and land surface modelling. The widely used time/rank stability analysis developed by Vachaud et al. (1985) assesses the skill of a single soil moisture sensor location to estimate the average over the site. A new method was presented by Friesen et al. (2007) and applied by Bircher et al. (2011) where soil moisture sampling was based on landscape units with internally consistent hydrologic behaviour. This method ensures statistically reliable validation via the reduction of the footprint variance and reduces the chance of sample bias.

11.65   Remote sensing of soil moisture

~~Earlier~~As mentioned earlier in this chapter ~~it was mentioned that~~, a single observation location cannot provide absolute knowledge of regional soil moisture~~, but only relative knowledge~~. Soil moisture is highly variable in both space and time, rendering it difficult to measure on a continental or global scale that is needed by researchers (Seneviratne et al., 2010). ~~Space-based r~~Remote sensing of soil moisture accommodates these needs by providing surface soil moisture observations on a global scale every one to two days under a variety of conditions.

In general, remote sensing aims to measure properties of the Earth's surface by analysing the interactions between the ground and electromagnetic radiation. This can be done by recording the naturally emitted radiation (passive systems) or by illuminating the ground and record the reflecting signal (active systems). Soil moisture is usually assessed either through its ~~change, because soils~~effects on the soil's electric or thermal properties. While microwave remote sensing observations are ~~often very inhomogeneous. However, nowadays measurements~~sensitive to the soil's dielectric constant, infrared remote sensing systems are sensitive to its thermal conditions. Part IV of this Guide provides information about Space-based observations, and sections 5.6.2 and 5.6.3 include basic principles of soil moisture observation within the context of many observed geophysical variables. In this section additional detail and practical information is provided.

Over the last decades many soil moisture datasets have been developed from various space-borne instruments using ~~remote~~ different retrieval algorithms (Owe et al., 2001, Njoku et al., 2003, Naemi et al., 2009). Recently, several of these datasets from both active and passive microwave remote sensing observations have been combined generating a global soil moisture dataset covering the last 30 years (Liu et al., 2011).

Although remote sensing has proven to be a valuable tool to measure soil moisture on a global scale, in situ soil moisture measurements are imperative for the calibration and validation satellite-based soil moisture retrievals. The International Soil Moisture Network (ISMN) is a global in-situ soil moisture database mainly developed for validation of satellite products. Many validation efforts have been undertaken to assess the quality of remote sensing products using in situ measurements (Albergel et al., 2012, Matgen et al., 2012, Pathe et al., 2009, Su et al., 2013, Wagner et al., 2008). In addition, many studies have focused on error characterisation of the different soil moisture products (Dorigo et al., 2010, Draper et al., 2013). These studies show that most soil moisture products from remote sensing are capable of depicting seasonal and short term soil moisture changes quite well. However, biases in the absolute value and dynamic range may be large when compared to in situ and modelled soil moisture data.

The next subparagraphs will give an overview of the theoretical background behind the different remote sensing techniques ~~are available for determining soil moisture in the upper soil layer. This allows interpolation at the mesoscale for estimation of evapotranspiration rates, evaluation of plant stress and so on, and also facilitates moisture balance input in weather models (Jackson and Schmugge, 1989; Saha, 1995). The usefulness of soil moisture determination at meteorological stations has been increased greatly thereby, because satellite measurements need "ground truth" to provide accuracy in the absolute sense. Moreover, station measurements are necessary to provide information about moisture in deeper soil layers, which cannot be observed from satellites or aircraft. Some principles of the airborne measurement of soil moisture are briefly given here; for more details see Part II, Chapter 8.~~, space-borne instruments and algorithms that are used.
~~Two uncommon properties of the water in soil make it accessible to~~

11.~~5~~6.1   Microwave ~~r~~Remote ~~s~~Sensing

11.~~5~~6.1.1 Introduction

Microwave remote sensing~~. First, as already discussed above in the context of time-domain reflectometry,~~ uses electromagnetic (EM) waves with wavelengths of 1m to 1cm, what corresponds to the frequency of 0.3 to 300 GHz. As an important quality, these microwaves can travel through the Earth's atmosphere undisturbed and thus allow observations independent from cloud coverage. Furthermore, since not bound to illumination by the

Sun, microwave measurements are operable all-day-round.

When applied to remote sensing of the earth surface, Kirchhoff's Radiation Law states that the emission of a body is equal to one minus its reflectivity. This means, emission and reflection are complementary, yielding that surfaces that are good scatters are weak emitters, and vice versa. As a result, active and passive microwaves systems are influenced inversely by the same physical phenomena on the ground. Fresnel's Reflection Law describes the relationship between the dielectric constant and reflectivity (and thus emissivity), where a higher dielectric constant yields a higher rate of reflection (and smaller emissivity). At microwave lengths the dielectric constant of water is of an order of magnitude larger than that of dry soils at microwave lengths. In remote sensing, this feature can be used either passively or actively (Schmugge, Jackson and McKim, 1980). Passive sensing analyses the natural microwave emissions from the Earth's surface, while active sensing refers to evaluating back scatter of a satellite-sent signal. Therefore, the dielectric constant of soils increases with increasing soil moisture (figure 1). With these physical relations, it is possible to retrieve soil moisture of the earth surface from passive as well as from active microwave remote sensing systems.

The microwave radiometer response will range from an emissivity of 0.95 to 0.6 or lower for passive microwave measurements. For the active satellite radar measurements, an increase of about 10 db in return is observed as soil goes from dry to wet. The microwave emission is referred to as brightness temperature T_b and is proportional to the emissivity β and the temperature of the soil surface, T_{soil}, or:

$$T_b = \beta\, T_{soil} \quad (11.9)$$

where T_{soil} is in kelvin and $\beta\beta$ depends on soil texture, surface roughness and vegetation. Any vegetation canopy will influence the soil component. The volumetric water content is related to the total active backscatter S_t by:

$$\theta_v = L\,(S_t - S_v)\,(RA)^{-1} \quad (11.10)$$

where L is a vegetation attenuation coefficient; S_v is the back scatter from vegetation; R is a soil surface roughness term; and A is a soil moisture sensitivity term.

As a result, microwave response to soil water content can be expressed as an empirical relationship. The sampling depth in the soil is of the order of 5 to 10 cm. The passive technique is robust, but its pixel resolution is limited to not less than 10 km because satellite antennas have a limited size. The active satellite radar pixel resolution is more than a factor of 100 better, but active sensing is very sensitive to surface roughness and requires calibration against surface data.

The second remote-sensing feature of soil water is its relatively large heat capacity and thermal conductivity. Therefore, moist soils have a large thermal inertia. Accordingly, if cloudiness does not interfere, remote sensing of the diurnal range of surface temperature can be used to estimate soil moisture (Idso and others, 1975; Van de Griend, Camillo and Gurney, 1985).
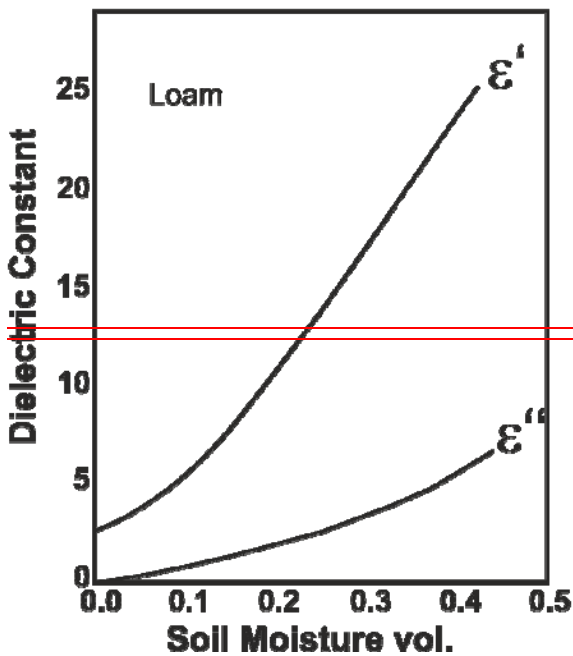
Figure 1: The relationship between the complex dielectric constant (ε' and ε'') and soil moisture for a loamy soil at a frequency of 5 GHz (After Hallikainen et al. 1985).

Microwave beams are able to interact to some extent with volumes of targets since their waves are longer and are not reflected immediately at the surface. Thus, information about inner conditions of vegetation or soils, for example, can be gained. As a rule of thumb, the longer the wavelength, the deeper the radiation penetrates into volumes. Contrariwise, optical waves only interact with surfaces and tell us about the visible colour and brightness.

When observed from above the canopy, vegetation affects the microwave emission in two ways: Firstly, vegetation absorbs or scatters the radiation emitted from the soil. Secondly, the vegetation also emits its own radiation. Under a sufficiently dense canopy, the emitted soil radiation will become totally masked and the observed radiation is mostly due to vegetation. Generally, all frequency bands used in microwave remote sensing of soil moisture are sensitive to vegetation and require some correction in the data for this. Higher frequency bands are more vulnerable to vegetation influences.

11.56.1.2 Multi frequency radiometers

Passive systems like radiometers record the brightness temperature of the Earth's surface. Brightness temperatures are related to the amount of emissivity (and thus reflection) described by the Rayleigh-Jeans approximation of Planck's law. This law states that brightness temperatures are a function of the physical temperature and emissivity. The amount of emission depends on the dielectric constant of the emitting body as described by Fresnel's reflection law.

Since 1978 instruments have been providing global passive data over land and oceans (Figure 11.1), beginning with the Scanning Multichannel Microwave Radiometer (SMMR, 1978-1987), the Special Sensor Microwave Imager (SSM/I, since 1987), the Tropical Rainfall Measuring Mission (TRMM, since 1997) and more recently the Advanced Microwave Radiometers (AMSR-E, 2002-2011 and AMSR-2 since 2012), Coriolis Windsat (since 2003) and the Chinese satellites FengYun III (since 2010). Initially these instruments were not designed for soil moisture observations, but for precipitation, evaporation, sea surface temperatures and cryospheric parameters. However, in the 1970s studies already showed the potential of retrieving soil moisture from brightness temperatures at these frequencies (Schmugge, 1976). The big advantage of radiometers is that data is available from multiple multi-frequency microwave radiometers since 1978, providing a long-term dataset to investigate trends and anomalies. The instruments that are used for soil moisture remote sensing have frequencies varying from 6.6 GHz to 10.7 GHz. It has to be taken into account that a higher microwave frequency leads to less accurate estimates of soil moisture since the attenuation of vegetation increases and penetration ability decreases. Therefore, retrievals from SMMR (6.6 GHz), AMSR-E (6.9 GHz), WindSat (6.8 GHz) and AMSR-2 (6.9 GHz) tend to be more accurate. Another advantage of these sensors is that spatial resolution and radiometric accuracy are much improved. The spatial resolution of AMSR-E is 56 km where the soil moisture products are provided at a spatial resolution of 0.25°.
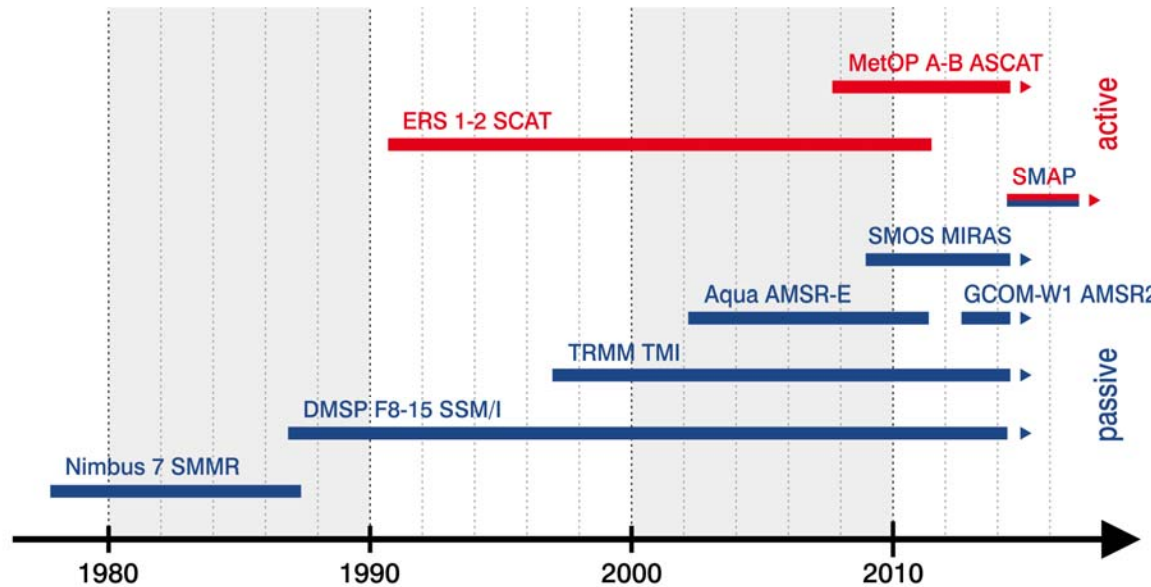
Figure 11.21: Active and passive microwave sensors used for soil moisture retrieval.

11.56.1.3 Scatterometer

A scatterometer is an active microwave instrument that continuously transmits short directional pulses of energy towards the Earth's surface and detects the returned energy. The amount of energy returned to the instrument depends upon geometric and dielectric properties of the surface and is often referred to as normalised radar cross section or backscatter (sigma nought, σ^0). Sacrificing range and spatial resolution, the strength of scatterometers over other types of radars is their higher accuracy and stability to measure the radar cross section of a target. Spaceborne scatterometers were initially developed and designed to derive wind speed and direction over the oceans. Nevertheless, a number of studies acknowledge the capacity of scatterometers for land applications such as monitoring of soil moisture (Magagi and Kerr, 1997; Pulliainen et al., 1998; Wagner et al., 1999). Since European scatterometers operate in longer wavelengths (5.3 GHz) than the US scatterometers (14 GHz) they are more suitable for soil moisture retrieval.

The unique instrument design of the European scatterometers on-board of the European Remote Sensing (ERS) satellites and the Meteorological Operational Platform (MetOp) enables soil moisture retrieval on a global scale with almost daily coverage. Both scatterometers, the Active Microwave Instrument (AMI) in wind mode on board ERS (Attema, 1991) and the Advanced Scatterometer (ASCAT) on board MetOp (Figa-Saldaña et al., 2002) operate in C-band (5.3 GHz) with a wavelength of approximately 5.6 centimetres. The major differences between these two scatterometers are the number of sideways looking antennas and the range of the incidence angles observed. The spatial resolution of the AMI is approximately 50 km with the ASCAT product being provided at spatial resolutions of 25 km and 50 km.

11.56.1.4 Synthetic Aperture Radars (SAR)

Space- or airborne synthetic aperture radars (SAR) are active microwave sensor systems that offer due to advanced signal processing a higher spatial resolution than scatterometers. As imaging side-looking radars, they operate similarly to scatterometers and use the same frequency domain. Besides hydrological applications, SAR systems can be used for the accurate retrieval of three-dimensional geometries, as they enable interferometry.

As an imaging side-looking radar moves along its path on the ground, it accumulates data. The spatial resolution of radars is dependent on the (limited) physical size its antenna, the aperture. Taking advantage of the along-track motion of the carrier, a SAR system simulates a bigger, synthetic aperture as it records amplitude and phase of the ground targets continuously while they are visible to the SAR. These multiple measurements of each target are then summed up coherently, what allows resolving smaller objects on the ground. However, higher energy consumption and a smaller footprint yield a longer revisit time on individual locations and thus the temporal resolution of SARs is inferior to other microwave systems these days.

The higher complexity of soil and surface properties at the scale below 10km introduces additional error and uncertainty sources. As a consequence, SAR systems are yet not employed in operational soil moisture services but instead are used for pre-operational services and scientific products (Doubkova et al., 2009, Pathe et al, 2009). Nonetheless, upcoming SAR satellite missions as ESA's Sentinel-1 programme (Attema et at., 2007) promise improved temporal and radiometric resolution and are suggested to be employed for operational soil moisture

services at a local scale (Hornacek et al, 2012).

11.56.1.5 Dedicated L-Band missions

As stated before, lower frequencies tend to be less sensitive to vegetation interactions, and are therefore thought to be more suitable for soil moisture retrieval. Hence, the first two spaceborne missions specifically designed for soil moisture retrieval operate in the L-Band channel (1.4 GHz). The aim of both SMOS and SMAP is to provide absolute soil moisture with a maximum Root Mean Square Error of 0.04 m^3/m^3.

The Soil Moisture and Ocean Salinity (SMOS) mission of the European Space Agency (ESA) was launched successfully on 2 November 2009. The instrument on-board of the SMOS satellite has a unique instrument design in order to provide the spatial resolution needed to measure soil moisture. The so-called Microwave Imaging Radiometer with Aperture Synthesis (MIRAS) instrument is a 2D interferometric radiometer, where the size of the antenna needed for measuring at the spatial resolution needed is simulated through 69 small antennas. MIRAS provides brightness temperatures with a spatial resolution varying between 30 and 50 km. Global coverage is achieved every 2-3 days.

The second L-Band mission is the Soil Moisture Active and Passive (SMAP) mission by NASA planned for launch in 2014/2015. As for SMOS, the passive microwave instrument operates in L-Band for enhancing the sensitivity to soil moisture. However, the instrument design for SMAP is very different from SMOS. SMAP will use a real aperture antenna in the shape of a large (6m) parabolic reflector antenna that will rotate. Measurements are made with a spatial resolution of 40 km. In addition to the passive measurements, SMAP will also carry a radar making concurrent measurements at a spatial resolution of 1-3 km. By combining the active and passive measurements a soil moisture product will be provided with a spatial resolution of 10 km.

11.56.1.6 Soil mMoisture rRetrieval

For retrieving soil moisture it is necessary to have models that are capable of accounting for vegetation and surface roughness effects on the microwave signal and then convert accordingly the received intensity to soil moisture values. Again, it should be noted, that shorter a wavelength causes inferior performance due to vegetation scattering and lesser penetration depth. Over densely vegetated areas such as the tropical rainforests, soil moisture retrieval is not possible due to the lack of penetration of the L-Band and C-band waves through the vegetation canopy. Additionally, retrieved estimates of soil moisture are only reasonable over snow-free and non-frozen soils.

Passive systems measure the microwave brightness temperature and derive indirectly the emissivity, which then is ingested in a radiative transfer model. Data on soil temperature, roughness, texture and other parameters of the observed area is necessary ancillary information. Data from passive microwave observations is available from AMSR-E using either the VUA-NASA retrieval algorithm that is based on Land Parameter Retrieval Model (LPRM) as described by Owe et al. (2001), the official NASA AMSR-E product (Njoku et al., 2003; Njoku 2004)[1] or the retrieval algorithm from the University of Montana (Jones et al., 2009, Jones and Kimballet al., 201009)[2]. All of these retrieval algorithms are based on radiative transfer equations. However, the retrieval algorithms vary significantly and generate quite different soil moisture values. The VUA-NASA retrieval algorithm solves for Vegetation Optical Depth and the soil dielectric constant simultaneously. Soil moisture is calculated using the Wang-Schmugge mixing model (Wang and Schmugge, 1980). The data can be found at NASA's Global Change Master Directory (National Aeronautics and Space Administration, 2013)[3].

SMOS provides an operational soil moisture product. (Kerr et al., 2012). The SMOS retrieval algorithm uses an iterative approach to minimise the cost function between modelled brightness temperatures and the direct measurements. Hereby the best set of parameters is found, including the soil moisture and vegetation. SMOS Level 2 soil moisture data can be downloaded via ESA Earthnet Online (European Space Agency, 2013)[4].

Active instruments measure the backscattered intensity, which is a function of roughness, incidence angle and dielectric properties of the surface. Again, vegetation and other influences contribute to the signal, which is used to determine the backscatter coefficient. Soil moisture retrieval provided as an operational product from the ASCAT scatterometer, and as a scientific product from the AMI in wind mode relies upon a semi-empirical change detection method. This method, the TU Vienna Change Detection algorithm is tailored to the unique instrument design. Assuming a linear relationship between radar backscatter and soil moisture, in the decibel domain, a relative measure of moisture in the first few centimetres of soil can be obtained, representing the degree of saturation (0-100%). In very dry regions, particularly over sand deserts, the retrieval approach fails believed to be caused by a complex scattering mechanism of surface, volume and sub-surface scattering. Soil moisture data

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[1] http://nsidc.org/data/ae_land.html
[2] http://nsidc.org/data/nsidc_0451.html
[3] http://gdata1.sci.gsfc.nasa.gov/daac-bin/G3/gui.cgi?instance_id=soilmoisture_daily
[4] https://earth.esa.int/web/guest/-/how-to-obtain-data-7329

from the TU Vienna change detection algorithm is freely available via the website of TU Vienna (Vienna University of Technology, 2013)[5] or EUMETSAT (EUMETSAT, 2013)[6].

An overview of operational soil moisture products is given in table 1.

Product Reference	SMOS	AMSR-E	ASCAT
Satellite			
Name	SMOS	Aqua	METOP-A/B
Agency	ESA/CNES	NASA	EUMETSAT/ESA
Launch	2.9.2009	4.5.2002 – 01.10.2011	19.10.2006
Orbit	Polar	Polar	Polar
Altitude	758 km	705 km	850 km
Period	100 min	99 min	100 min
Equator crossing time	6:00 (ascending) 18:00 (descending)	13:30 (ascending) 1:30 (descending)	21:30 (ascending) 9:30 (descending)
Type	Research satellite	Research satellite	Operational (3 satellites)
Sensor			
Name	MIRAS	AMSR-E	ASCAT
Type	Synthetic-aperture radiometer	Multi-frequency real-aperture radiometer	Real-aperture scatterometer
Swath	1000 km	1445 km	2 x 550 km
Scanning principle	Forward looking 2D-interferometer	Rotating parabolic reflector	6 side-looking fan beam antennas
Incidence angle range	0-55°	55°	25–53° (mid-beam); 34–64° (fore- and aft beams)
Frequency	1.4 GHz	6.9, 10.7, 18.7, 23.8, 36.5, and 89.0 GHz	5.3 GHz
Polarisation	H and V (polarimetric mode optional)	H and V	VV
Spatial resolution	30-50 km	75 x 43 km @ 6.9 GHz	25/50 km
Daily global coverage	~82 %	~90 %	~82 %
Retrieval			
Model name	L-MEB	LPRM	WARP
Forward model	Radiative transfer model	Radiative transfer model	Semi-empirical change detection
Model complexity	High	Medium	Low
Inversion approach	Iterative least-square matching	Iterative least-square matching	Direct inversion
Concurrent retrievals	Soil temperature, vegetation optical depth, roughness	Soil temperature, vegetation optical depth	None
Model calibration	None	None	Based on long-term time series
Need for auxiliary data	High	Medium	Low
Error propagation estimates	Not available	Available	Available
Product			
Target quantity	Volumetric soil moisture	Volumetric soil moisture	Degree of saturation
Units	m^3m^{-3}	m^3m^{-3}	0-1 or %
Grid	Fixed ISEA4-9 Discrete Global Grid	Regular Grid	Swath geometry

[5] http://rs.geo.tuwien.ac.at/products/
[6] http://www.eumetsat.int/website/home/Data/Products/Land/index.html

Pixel spacing	15 km	0.25°	12.5 km
Data latency	Within few days after sensing	Irregular updates	Within 130 min after sensing

Table 1: Operational Soil Moisture products and their characteristics.

11.~~56~~.2 Thermal ~~i~~Infrared ~~r~~Remote ~~s~~Sensing

All bodies with a temperature above absolute zero emit electromagnetic energy in the thermal infrared domain. By detecting the thermal properties of the earth surface, soil moisture can be derived based on the distinct differences in thermal properties of soil and water (Idso et al., 1975; Van de Griend, Camillo and Gurney, 1985). Thermal infrared (TI) remote sensing has been used by an increasing number of studies for the derivation of soil moisture. The advantage of thermal infrared remote sensing is that it can provide soil moisture information on a spatial resolution down to a few meters. Furthermore, it can provide soil moisture information over dense vegetation, which is one of the limitations of microwave remote sensing. A disadvantage of TI remote sensing is the inability to measure soil moisture when cloud cover is present and that it is considerably affected by atmospheric effects. Therefore, complex noise removal mechanisms are needed in most cases. TI remote sensing of soil moisture is not as straightforward as microwave remote sensing since there is no direct link between temperature data and soil moisture (Jackson et al., 1975). Nevertheless, several approaches exist to indirectly retrieve soil moisture data using thermal infrared observations from GOES, AVHHR, MODIS, Landsat and others.

The first approach is called triangle approach and is based on the empirical relationship between soil moisture, soil temperature and fractional vegetation cover. This relationship was demonstrated by Price et al. (1990) and resulted in a triangular scatterplot of surface temperatures and remotely sensed NDVI. The triangle approach was later used in several studies to estimate soil moisture, by Sandholt et al. (2002) and Carlson, Gillies and Perry (1994) among others.

The second approach makes use of the differences in thermal properties between water and soils. Water differentiates from many other matters with its relatively large heat capacity and thermal inertia. Thermal inertia is defined as the resistance of an object against its heating for 1K. The thermal inertia of water is relatively high, which indicates a high resistance to temperature changes. It has been shown that the behaviour of land surface temperature in the morning is strongly depending on soil moisture in the soil, since the water will heat up more slowly. One of the approaches that utilises this behaviour is the calculation of the Apparent Thermal Inertia (ATI). When measuring the difference between maximum and minimum temperatures over one day the ATI can be measured. It is described as

$$ATI = (1-A)/\Delta T \qquad (11.11)$$

where A is the albedo of the pixel in the visible band and ΔT the difference between the minimum and maximum temperature. Many studies have already assessed the potential of ATI to describe soil moisture and its spatial and temporal variability (e.g. Verstraeten et al., 2006, Van Donninck et al., 2011).

Another method to retrieve soil moisture using thermal infrared remote sensing is by integrating the data into land surface models. Soil moisture controls latent heat fluxes by way of both evaporation and transpiration, where wet soil conditions lead to increased evaporation and transpiration. The Atmosphere Land Exchange Inversion model (ALEXI) uses the relationship between evaporation, transpiration and soil moisture to derive soil moisture data. All major components, including the latent heat flux, of the energy budget are estimated from net radiation and vegetation parameters retrieved from AVHHR and GOES. Accordingly, soil moisture can be derived from latent heat fluxes by using a soil water stress function (Anderson et al., 1997, 2007, Hain et al., 2011). An intercomparison of soil moisture retrieved from microwave remote sensing and ALEXI showed that the two datasets are complementary, where ALEXI is better at estimating soil moisture over dense vegetation and microwave remote sensing shows more reliable results over low to moderate vegetation (Hain et al., 2011).

~~11.6 Site selection and sample size~~

~~Standard soil moisture observations at principal stations should be made at several depths between 10 cm and 1 m, and also lower if there is much deep infiltration. Observation frequency should be approximately once every week. Indirect measurement should not necessarily be carried in the meteorological enclosure, but rather near it, below a sufficiently horizontal natural surface which is typical of the uncultivated environment.~~

~~There is no standard depth or measurement interval at which soil moisture observations are taken, since this~~

strongly depends on the research objectives for which the sensors are installed. The International Soil Moisture Network (Dorigo et al. 2011) provides an extensive database with harmonised in situ soil moisture time series of networks distributed all over the world. Here data is harmonised to a half-hourly measurement interval whenever possible. Most networks and stations that are incorporated within the ISMN measure soil moisture at several depths starting at 0.05 m and up to a depth of 0.50 m or 1.00 m. By measuring soil moisture at different depths the behaviour of soil moisture at different depths can be compared and used to validate measurements. Measurements of other meteorological parameters is very valuable for soil moisture measurements. For example, precipitation data at the measurement site provides a validation for the soil moisture data.

Representativity of any soil moisture observation point is limited because of the high probability of significant variations, both horizontally and vertically, of soil structure (porosity, density, chemical composition). Horizontal variations of), land cover and relief. It is pivotal that soil water potential tend to be relatively less than such variations of soil water content.moisture and its variability are captured on the scale necessary for studies focussing on hydrological processes and for satellite validation. Gravimetric water content determinations or indirect measurements of soil moisture are only reliable at the point of measurement, making a large number of samples necessary to describe adequately the soil moisture status of the site. To estimate the number of samples *n* needed at a local site to estimate soil water content at an observed level of accuracy (*L*), the sample size can be estimated from:

$$n = 4\,(\sigma^2/L^2) \qquad (11.1112)$$

where σ^2 is the sample variance generated from a preliminary sampling experiment. For example, suppose that a preliminary sampling yielded a (typical) σ^2 of 25 per cent and the accuracy level needed to be within 3 per cent, 12 samples would be required from the site (if it can be assumed that water content is normally distributed across the site). A study of Brocca et al. (2007) showed that the minimum number of point samples needed for an area in Central Italy with an extent of around 9 to 8800 m2 varies between 15 and 35 samples. The higher number of samples was needed for sites with more significant relief. Famiglietti et al. (2008) found that a number of 30 samples are sufficient for a footprint of 50 km, taking into account that the data is independent and spatially uncorrelated.

A regional approach divides the area into strata based on the uniformity of relevant variables within the strata, for example, similarity of hydrological response, soil texture, soil type, vegetative cover, slope, and so on. Each stratum can be sampled independently and the data recombined by weighing the results for each stratum by its relative area. The most critical factor controlling the distribution of soil water in low-sloping watersheds is topography, which is often a sufficient criterion for subdivision into spatial units of homogeneous response. Similarly, sloping rangeland will need to be more intensely sampled than flat cropland. However, the presence of vegetation tends to diminish the soil moisture variations caused by topography.Upscaling of the point measurements obtained by gravimetric water content determination or indirect measurements with in situ sensors has been the subject of many studies. Upscaling techniques vary to relatively straightforward interpolation and time/rank stability techniques to more complicated techniques such as statistical transformations and land surface modelling. The widely used time/rank stability analysis developed by Vachaud et al. (1985) assesses the skill of a single soil moisture sensor location to estimate the average over the site. A new method was presented by Friesen et al. (2007) and applied by Bircher et al. (2011) where soil moisture sampling was based on landscape units with internally consistent hydrologic behaviour. This method ensures statistically reliable validation via the reduction of the footprint variance and reduces the chance of sample bias.

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