3. Water cycling

S Reinsch & D Robinson (editors)

Within ecosystems, the exchange of water and energy between the soil, plants, and the atmosphere is often termed the soil–plant–atmosphere continuum (SPAC). Understanding how SPAC works is a major research challenge, of interest to climate-change research and wider research on environmental change. We need to know what enters the system via precipitation and how this water is partitioned between runoff, interception, and infiltration. Of the water that infiltrates soil we need to know whether it drains to groundwater, is evaporated, or is transpired by plants. Studies in this area have uncovered a range of processes that enable ecosystems to thrive, often under challenging environmental conditions. Active research is underway regarding processes such as hydraulic redistribution (Ryel et al., 2002), preferential soil water flux (Clothier et al., 2008), alternative soil moisture states (Robinson et al., 2016), plant-induced wetting patterns (Franz et al., 2011), and the development and persistence of soil water repellency and how it alters infiltration (Doerr et al., 2000). This research is important because soils and plants provide feedbacks to climate. We need to understand these feedbacks to ensure that ecosystem management strategies, interventions, or policy regulations do not exacerbate environmental pressures (Robinson et al., 2019).

In this chapter, we focus on general soil measurements for site characterisation and soil physical measurements that are relevant to understanding SPAC. The major focus is on soil hydraulic measurements, which include soil moisture, hydraulic conductivity, water retention, and water potential. The methods included are a starting point for determining parameters that link to, or are used in, modelling SPAC. In addition, we include some measurements used to track the progress of water through the plant to the atmosphere (for other protocols relevant to the SPAC also see 5.8 Psychrometry for water potential measurements, 5.9 Pressure-volume curve – TLP, ε, Ψo, 5.10 Maximum leaf hydraulic conductance, 5.13 Stable isotopes of water for inferring plant function and 5.16 Leaf hydraulic vulnerability to dehydration). The purpose of this chapter is not to provide a fully comprehensive review of methods, but to identify some critical ones that will get a research team started with measurements for their work in climate-change studies. The methods may also be applied to other types of studies investigating global-change drivers (e.g. nutrients, land-use change). The extensive references are intended to provide the reader with links to a wider literature selection.

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How to cite a protocol:

E.g. To monitor soil moisture we used the method described in protocol 3.1 Soil moisture in the Supporting Information S3 Water cycling in Halbritter et al. (2020).

Halbritter et al. (2020) The handbook for standardised field and laboratory measurements in terrestrial climate-change experiments and observational studies (ClimEx). *Methods in Ecology and Evolution*, *11*(Issue) pages.

How to cite an updated protocol version:

E.g. To monitor soil moisture we used the method described in protocol 3.1 Soil moisture in the Supporting Information S3 Water cycling in Halbritter et al. (2020), using the updated protocol version, Date, available in the online version: www.climexhandbook.uib.no.

Halbritter et al. (2020) The handbook for standardised field and laboratory measurements in terrestrial climate-change experiments and observational studies (ClimEx). *Methods in Ecology and Evolution*, *11*(Issue) pages.

Table of contents

3.1 Soil moisture

Authors: Robinson DA[1], Jones SB[2]

Reviewer: Reinsch, S[1]

Measurement unit: $m^3\ m^{-3}$; Measurement scale: plot; Equipment costs: €€; Running cost: €; Installation effort: moderate; Maintenance effort: moderate; Knowledge need: moderate; Measurement mode: data logger

Soil moisture is the amount of water in the soil (Robinson et al., 2008; Vereecken et al., 2008). It provides the biological moisture pool for microbial activity and plant transpiration supporting terrestrial life. Soil moisture is a key variable in hydrological, agricultural, and climate- and global-change research. Soil moisture dynamics are likely to respond in different ways to climate change, depending on whether it leads to drought, warming, or excess rainfall (Seneviratne et al., 2010). This will directly affect the biologically available moisture pool and oxygen levels in the case of wet soils. Moreover, because soil moisture controls microbial activity, carbon and nutrient cycling will be affected, as will greenhouse gas fluxes of CO_2, CH_4, and N_2O.

Research is now emerging that suggests extreme events such as drought can lead to unforeseen feedbacks in moisture dynamics and potential soil moisture state shifts through structural alteration (Robinson et al., 2016) or the development of water repellency (Goebel et al., 2011). We are only just beginning to identify the occurrence and importance of such phenomena. At continental scales we now understand the important role of soil moisture for the energy balance and how soil moisture deficits contribute to increasing the magnitude of heatwaves (Seneviratne et al., 2006). This is why efforts to create community databases, such as the international soil moisture network (Dorigo et al., 2013) and the global database on soil infiltration data (Rahmati et al. 2018), are important.

3.1.1 What and how to measure?

Soil moisture sensors

The most commonly used methods for soil moisture determination are dielectric sensors with time domain reflectometry (TDR; Ferré & Topp, 2002; Robinson et al., 2003). A range of sensors are available and compared in Blonquist et al. (2005a). Given the different performance characteristics arising from measurement frequency, the new generation of digital time domain transmission (TDT; Blonquist et al., 2005b) or TDR sensors are less susceptible to errors experienced by lower frequency capacitance and impedance-based sensors. Most sensors require low power and are affordable with their own data loggers. The signal processing is generally performed on a microprocessor in the head of the sensor, so for TDR, cable length is not a major limitation. The digital TDR and TDT family of sensors has well-characterised sampling volumes, larger than other low-frequency sensors. Moreover, because they work in the GHz frequency range they are the least susceptible to electrical loss due to solution electrical conductivity. The addition of a temperature sensor in the head of most sensors today gives another important measurement parameter for understanding soil physical behaviour. The preferred location of soil moisture sensors is discussed in protocol 1.5. Meteorological measurements.

Other methods of soil moisture measurement

The standard method for soil moisture determination is taking a volumetric core and oven drying. Although important for calibration, the destructive manual nature of the method renders it unsuitable for most climate-change and long-term monitoring experiments. A range of techniques are described in Evett et al. (2008).

Where to start

Choosing a sensor can be challenging: price is not always a good guide. Sensors determine water content from permittivity (Robinson et al., 2003). See the paper by Blonquist et al. (2005a) to compare performance and electrode length characteristics of a number of commonly used sensors. The new generation of digital sensors offers low-cost robust measurements (Blonquist et al., 2005a). For monitoring you will then need to decide if you want to measure vertically to determine moisture storage, or horizontally to obtain a definitive depth. Sensors are often installed horizontally by digging a small trench and then inserting them into the soil layer of interest (see Evett et al., 2008 for more information on installation).

3.1.2 Special cases, emerging issues, and challenges

Horizontal installation at multiple depths is helpful in determining the water balance and convenient for comparison with models such as HYDRUS-1D (Šimůnek et al., 2016). Calibration from permittivity to soil moisture content can be done using a standard calibration for mineral soils (e.g. Topp et al., 1980) and many sensors now incorporate something similar. For clay soils and organic soils more specific calibration may be required. Placing three sensors in the soil is ideal: for example, one near the soil surface at around 5 cm, one at the depth of maximum rooting density, and one below the roots (see Figure 1.5.1 in protocol 1.5. Meteorological measurements).

3.1.3 References

Theory, significance, and large datasets

D'Odorico & Porporato, (2004), Jung et al. (2010), Rodriguez-Iturbe (2000), Seneviratne et al. (2010), Taylor et al. (2012)

More on methods and existing protocols

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3.2 Soil hydraulic conductivity

Authors: Robinson DA[1] and Marshall M[1]

Reviewers: Jones SB[2]

Measurement unit: cm day^{-1}; Measurement scale: plot; Equipment costs: €€; Running costs: €; Installation effort: low; Maintenance effort: low; Knowledge need: moderate; Measurement mode: manual

Soil hydraulic conductivity depends on soil structure and water content. Hydraulic conductivity controls the water-flow rate through soil and the time to ponding when water at the soil surface is partitioned between infiltration and runoff. Hydraulic conductivity is treated as a parameter in most models, however, it is increasingly recognised that changes to structure brought about by tillage, vegetation, and climate and land-use change make it a variable and subject to feedbacks (Robinson et al., 2019). Climate change, especially extreme events such as flood and drought, can change the soil structure and hence alter the speed at which water enters the soil and affect the partitioning. This is important because any soil water which runs off, or is directly evaporated, is transferred to the biological soil moisture pool and is not available for transpiration. Typical values of soil hydraulic conductivity are 500 cm day^{-1} for sand, 50 cm day^{-1} for loam, and 5 cm or less day^{-1} for clay. These values are important input parameters for soil hydraulic models such as HYDRUS-1D (Šimůnek et al., 2008).

3.2.1 What and how to measure?

Soil infiltrometers

Soil hydraulic conductivity can be measured in the field or in the laboratory. The preferred method is in the field because this is minimally invasive as it does not require the removal of a soil core. Saturated soil hydraulic conductivity measurements (made in the lab) are generally used as a reference value, although soils are rarely saturated except below the water table. Measurement of hydraulic conductivity near saturation can be highly variable depending on the presence of macropores: many researchers use tension-infiltrometers to eliminate macropore flow from the measurement (Angulo-Jaramillo et al., 2000). Field estimates of hydraulic conductivity using infiltrometers are not typically fully saturated due to trapped air, providing only approximate values of saturated hydraulic conductivity. Furthermore, long-term infiltration rates are generally not well-captured by infiltrometers with small water reservoirs. Models are used to fit infiltration data to hydraulic parameters such as infiltration rate, sorptivity, and soil hydraulic conductivity (Mualem, 1986; Šimůnek et al., 2003). A range of disc tension-infiltrometers is available, with mini-disc infiltrometers proving optimal for plot-level work as they have a small disc that causes minimal impact on soils. Mini-disc infiltrometers can be read manually or attached to pressure transducers for multiple simultaneous measurements (Madsen & Chandler, 2007). Each measurement typically takes 30 minutes and will depend on the soil type. Multiple measurements per plot will be required to account for spatial variability.

Where to start

Mini-disc infiltrometers are arguably the easiest place to start - Madsen & Chandler (2007) (see Figure 3.2.1). These simple and easy-to-use devices can provide estimates of unsaturated hydraulic conductivity at tensions between 1 and 6 cm. The pdf guide is helpful (https://www.metergroup.com/environment/products/mini-disk-infiltrometer/). Some background to the measurements can be found in Angulo-Jaramillo et al. (2000), Mualem (1986), and Šimůnek et al. (2003, 2008).

Figure 3.2.1 Setup of infiltration measurements in a forest (left) and on a grassland soil (right) using mini-disk infiltrometers and stop watches. Photo: Francis Parry (Centre for Ecology & Hydrology).

3.2.2 Special cases, emerging issues, and challenges

Ensure the base is clean and unobstructed to water flow before priming the infiltrometer. Create a level surface of native soil or fine sand, perhaps 2 mm thick, to ensure a fully hydraulically conducting interface across the entire area of the infiltrometer. To work properly, good hydraulic contact must be maintained and the infiltrometer should be positioned vertically and not leaning.

3.2.3 References

Theory, significance, and large datasets

Durner (1994), Jarvis et al. (2013), Mualem (1976), Nielsen et al. (1973), Porporato et al. (2004), Robinson et al. (2016), Schaap & Leij (1998), Schaap et al. (2001), van Genuchten (1980)

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Amoozegar & Warrick (1986), Decagon, (2017), Mualem (1986)

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3.3 Soil water retention

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Reviewers: Jones SB[2], Reinsch S[1]

Measurement unit: volumetric water content (m^3 m^{-3}) v. soil matric potential (kPa); Measurement scale: plot; Equipment costs: €€€; Running costs: €; Installation effort: medium; Maintenance effort: medium; Knowledge need: moderate; Measurement mode: manual

Drought and flooding, as well as changes to land use can cause changes to soil structure, which in turn will alter the hydraulic response of the soil and the soil's ability to retain water (Robinson et al., 2019). Soil hydraulic properties primarily control the partitioning of water between what infiltrates into the soil and what runs off. The soil water retention curve or, water release curve (WRC), defines the relationship between the soil volumetric moisture content (Θ) and the pressure head (h). Hydraulic conductivity is a function of the soil saturated hydraulic conductivity (Ks), and can also be defined in terms of h or Θ (Ramos et al., 2006; Sakai et al., 2015). The relationships Θ(h), K(h), and K(Θ) define hydraulic properties of soil within the context of the Richards equation, which describes unsaturated water movement through soil (Rashid et al., 2015). These relationships are highly non-linear and vary as a function of soil texture (van Genuchten, 1980), but are also influenced by biotic and abiotic factors (Marshall et al., 2009). Accurate estimates of soil hydraulic properties are critical for simulating a range of soil hydrological processes, including water infiltration, surface runoff, evaporation, transpiration, profile soil moisture content, and solute transport (Ritchie et al., 1972; Gupta & Larson, 1979; Novak et al., 2000; Siyal et al., 2013; Sakai et al., 2015). These properties also have a direct impact on gaseous movement and heat transfer with subsequent impacts on many biogeochemical properties and processes. Soil hydraulic properties are used to address problems across a range of disciplines including ecology, environmental sciences, biogeochemistry, and agriculture (Durner & Flühler, 2005) and are a requirement to evaluate water and energy balances between the land surface and atmosphere within global circulation models (Mohanty et al., 2002). Robinson et al. (2016) measured climate change-induced shifts in soil hydraulic properties of an Atlantic heath resulting from drought-induced changes to soil structure; this led to a change in soil moisture dynamics and a change in the ability to retain soil moisture.

3.3.1 What and how to measure?

Soil water retention measurement

Numerous methods exist for quantifying soil hydraulic properties either directly through laboratory or field measurements (Šimůnek & van Genuchten 1996; Shao & Horton, 1998; Dane & Topp, 2002; Fujimaki & Inoue, 2003) or indirectly through the use of pedotransfer functions (PTFs) or inverse solutions of the Richards flow equation (Neuman, 1973). PTFs are equations used for the indirect estimation of a soil from more easily measured properties such as bulk density and texture (Parasuraman et al., 2007; Vereecken et al., 2010; also see protocol 1.3 Soil types and physical characteristics). Much research has focused on developing a function that can describe the WRC across all soil types but as yet no single function can be described as generic. Inverse solution methods have been used to estimate WRC by minimising the sum of squared deviations between observed and predicted transient flow data (Rashid et al., 2015; Hopmans et al., 2002). However, this requires data describing soil moisture dynamics that are both representative and accurate.

To determine the impact of climate change on soil hydraulic properties we need to measure parameters that are relevant, indicative of change, and sensitive to the treatments applied. The evaporation method first considered in the 1960s, and simplified by Schindler & Müller (2006), is a commonly used laboratory procedure allowing accurate characterisation of soil hydraulic properties at the wet end of the WRC. The method works by calculating the WRC and unsaturated hydraulic conductivity simultaneously from measuring the changes in weight and tension of a saturated soil sample dried by evaporation. Tensiometer measurement limitations mean that the method only works between 0 to ~-80 kPa (Schindler et al., 2010).

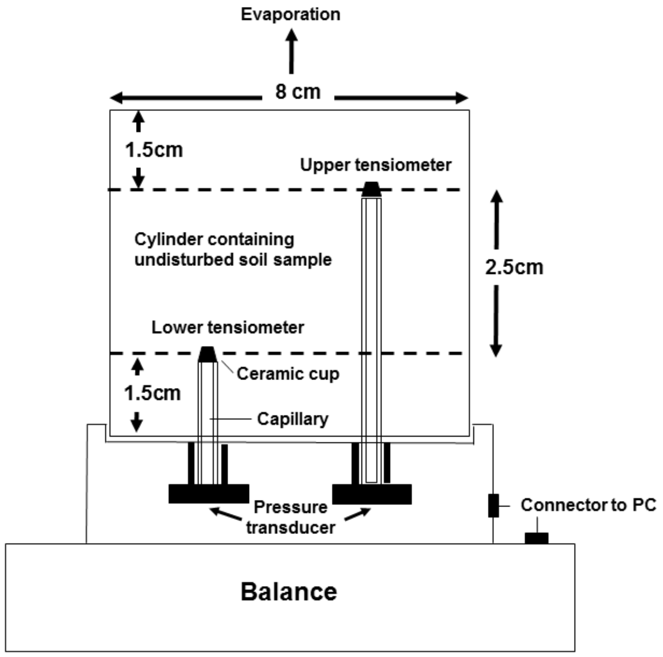

Figure 3.3.1 Illustration of the HYPROP device (modified from Shindler et al., 2010).

At higher suctions, the dry end of the WRC can be evaluated by determining the relative humidity (Rh) of soil air considered to be in equilibrium with the soil water phase using the chilled mirror method (Jensen et al., 2015). Samples are equilibrated within a sealed chamber containing a controlled temperature mirror with the means of detecting condensation. Decagon Devices Inc. (Pullman) offer two commercially available apparatuses that determine soil moisture relationships across both the wet and dry ends of the WRC. The HYPROP® system performs the evaporation method using two tensiometers and a weighing balance (Figure 3.3.1 and 3.3.2) to provide estimates of ϴ(h) and K(h) (Shindler et al., 2010). The WP4 (models T and now C) Water Potentiameter is a hygrometer that performs the chilled mirror dew-point method to measure soil moisture potential at the dry end of the WRC. The gravimetric moisture content, ϴg, of the WP4 samples are then determined in the standard method and converted to volumetric moisture content, ϴv, to determine ϴ(h) at the dry end.

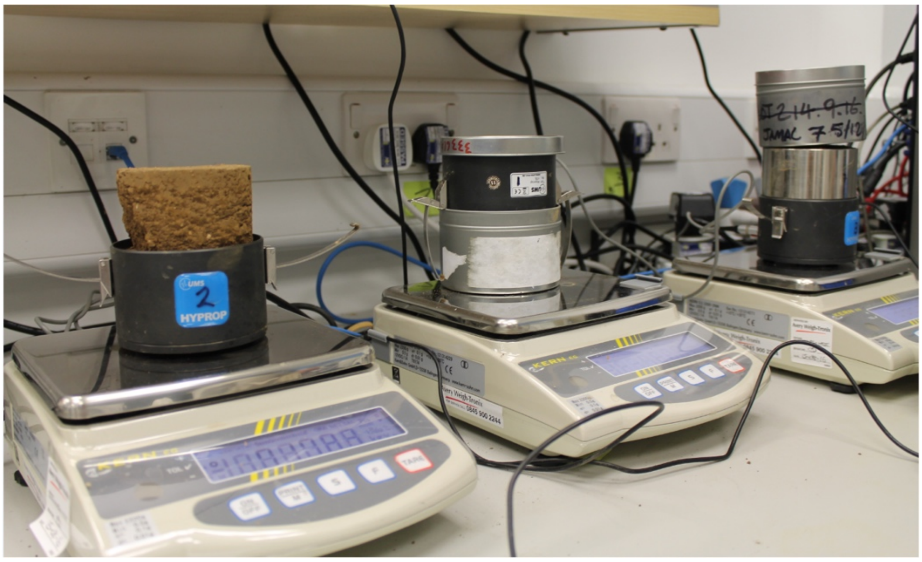

Figure 3.3.2 A laboratory setup for demonstration showing three balances with the HYPROP pressure transducers and the soil cores on top (bottom) Note: soil core remains in the cylinder during water loss but was removed for demonstration. Photo: Inma Lebron (Centre for Ecology & Hydrology).

Where to start

If using the evaporation method (Schindler & Müller, 2006; Schindler et al., 2010) read the instructions book. More detail can be found in the following articles: Dane & Topp (2002), Durner & Flühler (2005).

3.3.2 Special cases, emerging issues, and challenges

Retention measurements are challenging to do correctly. If using the evaporation method some practice will help. Soil cores can be taken from any soil depth of interest considering that the HYPROP rings are 5 cm in height. If soil cores are taken from deeper down in the profile, a soil pit needs digging with enough space to allow the space for the equipment and to lever the soil core out. Ensure any cores taken in the field are completely filled with soil and no gaps are left by stones etc. Soil protruding the ring edge can easily be sliced off using an ordinary bread knife. Cores can be kept moist by wrapping them in cling film. In the lab always saturate from the bottom up, this is ensured by clearly marking the top and the bottom of the core in the field. Try to keep the temperature relatively constant by best storing the cores in a fridge (e.g. 4°C). No storage limit is known but cores need to be retained at field moisture without perturbation. If possible process cores quickly to avoid disturbances. Be careful using tensiometers – they are fragile and easily broken – do not over tighten or apply excess force (i.e. via syringe) to the pressure transducers, exceeding their capacity and breaking the fragile membrane.

3.3.3 References

Theory, significance, and large datasets

Cosby et al. (1984), Peters & Durner (2006), Robinson et al. (2016), Schaap & Leij (1998a, 1998b), Schaap et al. (2001), van Genuchten (1980)

Methods and protocols

Schindler & Müller (2006), Schindler et al. (2010)

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3.4 Soil water potential

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Reviewer: Robinson DA[2]

Measurement unit: soil water potential – total soil water potential: μ in J kg^{-1} (energy per unit mass), ψ in N m^{-2} (energy per unit volume) or *h* in m (energy per unit weight) v. volumetric water content (m^3 m^{-3}). Although only μ has the unit of potential in the strict sense, the various expressions are equivalent under the assumption of constant density of water and all expressions are generically used in soil and plant sciences; Measurement scale: plot; Equipment costs: €€€; Running costs: none; Installation effort: moderate; Maintenance effort: low; Knowledge need: moderate; Measurement mode: data logger

Whereas soil moisture indicates the amount of water in the soil, the soil water potential provides information on the force with which that water is held by the soil matric forces (soil suction) in addition to chemical forces at play (e.g. salts): in other words how soil water is retained. Root water uptake by vegetation is determined by the difference in the total water potential between soil and root xylem (Steudle & Peterson, 1998). While the soil water potential may consist of several terms (e.g. Jury et al., 1991), the dominant term in unsaturated soil is the matric potential which describes water retention as a result of capillary forces in plants and soil, in addition to molecular imbibition forces associated with cell walls in plants and colloidal surfaces binding some of the soil water. Although soil water retention is harder to measure than soil water content, it is essential in dry soils where soil moisture sensors may fail to pick up small changes in soil moisture. While only small changes in soil moisture occur in dry soil, these reflect a large change in the soil water potential and thus have large consequences for the ability of microbes and vegetation to take up water. Root-sourced signals in response to soil water availability appear to play a key role in regulating stomatal aperture (e.g. Bacon, 2004). Moreover, as plants may exhibit adaptivity to changing environmental conditions (Rodriguez et al., 2008; von Arx et al., 2012) and exhibit differences in drought sensitivity (Engelbrecht et al., 2007; Harnett et al., 2013), the protocol can also be of use in other study types considering climate change and land-use change.

3.4.1 What and how to measure?

Soil water potential measurement

Tensiometers filled with a polymer solution (called osmotic, polymer, or full-range tensiometers) are currently the only instruments capable of measuring soil water matric potential directly with adequate accuracy and low maintenance under field conditions (Bakker et al., 2007; van der Ploeg et al., 2010). The measuring range is between 0 and 2 MPa (1 Pa = 1 Nm^{-2}) soil suction with 0.1% full scale accuracy for the pressure sensor and 0 to 40 °C with 0.01 °C accuracy for the temperature sensor. The osmotic properties of the polymer ensure that the tensiometer recovers negative pressure measurement ability automatically after drying below -2 MPa followed by rewetting of the soil. The measurement volume is around the volume of the employed ceramic interface between soil and polymer solution. The sensors can be installed as stand-alone with a data logger and lithium battery, or employed as part of a multi-port data logger with battery.

Alternatively, water-filled tensiometers are widely used to monitor the soil water matric potential and have been used for almost 100 years (Or, 2001; Young & Sisson, 2002). However, measurements are limited to about 1 atmosphere (e.g. 0.1 MPa) soil suction under field conditions at sea level. Porous matrix sensor methods are available that employ dielectric methods to measure water content in a ceramic material with known water retention characteristics and in equilibrium with the surrounding soil (e.g. Whalley et al., 2007). Comparison under laboratory conditions of a porous matric sensor type with a measurement range of 0.01

to 0.5 MPa soil suction with polymer tensiometers shows good reliability, although possible temperature effects on performance are unknown (Degré et al., 2017).

Where to start

Bakker et al. (2007) describe the principles of the polymer tensiometer; Degré et al. (2017) present a comparison of polymer tensiometers with two alternative porous metric sensor types; Or (2001) provides a historic overview of tensiometry; van der Ploeg et al. (2010) show how polymer tensiometers perform in two soil types; Whalley et al. (2007) explain the principle of a dielectric porous matrix sensor; Young & Sisson (2002) provide a technical overview of conventional tensiometry.

3.4.2 Special cases, emerging issues, and challenges

Sensors considering measurement of the soil water potential are installed below the soil surface, preferably inserted horizontally to prevent preferential flow of water along the data cables that lead to the data logger. The sensors need to be in good contact with the surrounding soil. Compensating for temperature effects in the shallow subsoil is necessary for all soil water potential sensors, but especially for polymer tensiometers, which therefore generally have an integrated temperature sensor.

3.4.3 References

Theory, significance, and large datasets

Bolt et al. (1976), Corey & Klute (1985), Grant & Backmann (2002), Groenevelt & Bolt (1969), Nitao & Bear (1996)

More on methods and existing protocols

Young & Sisson (2002)

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3.5 Soil temperature

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Measurement unit: °C; Measurement scale: plot; Equipment costs: €€; Running costs: €; Installation effort: medium; Maintenance effort: medium; Knowledge need: medium; Measurement mode: data logger

Soil temperature is a measure of the intensity of heat present in soil (Buchan, 2001), which impacts critical processes taking place, including the germination of seeds, microbial activity (Hanson et al., 2000; Fierer et al., 2006), chemical reactions, carbon sequestration (Kirschbaum, 1995), gas production and emissions (Schaufler et al., 2010), the growth and maintenance of plant roots (Atkin et al., 2000), soil evaporation (Kalma et al., 2008), plant transpiration, and freeze/thaw cycling.

Atmospheric conditions drive near-surface soil temperature, which is directly affected by climate change and position in the landscape (e.g. along gradients). Soil temperature is perhaps one of the most common soil measurements and reliable long-term monitoring stations (e.g. Rothamsted, UK; 1931–present) provide valuable information on temperature trends associated with climate change and other global-change drivers. An initiative building a global database of soil temperatures (https://soiltemp.weebly.com/) may help to provide global soil temperature records in the future.

3.5.1 What and how to measure?

Soil temperature measurement – direct contact measurements

Subsurface temperatures are more commonly measured using a thermocouple or a thermistor, and increasingly are incorporated into other sensors, such as soil moisture sensors. Thermocouples employ a bi-metal junction where the gradient in voltage is directly proportional to the gradient in temperature (Seebeck effect) and must be read using an advanced circuit common in multi-meter and data-logging devices. A thermistor is a type of negative coefficient resistor, whose resistance is dependent on temperature and which can be read by a simpler voltage reading circuit. Today's thermistors and thermocouples can resolve 0.01 °C but absolute temperature calibration can be costly to achieve at this resolution. More accurate temperature measurements are available using a resistance temperature detector (RTD), which uses a more expensive metal-based temperature-dependent resistance circuit. Subsurface temperature sensors are generally epoxy-coated and may be embedded in stainless steel tubing for protection from corrosion and water damage. Advances in microelectronics have facilitated heat pulse probes (Campbell et al., 1991) that employ temperature rise measurements in soil to infer soil thermal properties as well as other processes of interest (e.g. soil heat- and water-flux density; Yang et al., 2013). Because of the low cost of temperature-sensing circuits, most environmental sensors today include a temperature measurement in addition to other properties (e.g. soil moisture, electrical conductivity).

Soil temperature measurement – non-contact measurements

Measurement of a bare soil surface using contact thermometry is prone to serious errors (i.e. contact loss, surface disturbance). Remotely measured soil temperature relies on surface-emitted infrared radiation sensed by a distant thermometer (infrared-spot thermometer, pyrometer, or detector array). Early instruments employed fine-wire thermocouples but advances in microelectronics have produced faster response and more accurate detectors for infrared thermometry applications. Infrared radiation emitted from a surface is a function of the surface emissivity as well as surface temperature. Although most surfaces including soils have high emissivity of 0.9 to 0.97 (Fuchs & Tanner, 1968), which is a key assumption of today's low-cost radiometers, measurement errors arise when lower emissivity surfaces (e.g. highly reflective) are detected without adjusting for the assumed high emissivity value.

Installation

Soil sensors are often used to monitor seasonal changes in the soil environment. Daily changes occur, but can often be considered noise against the slower seasonal signal. Rather than measuring at fixed depths, it is of interest to know the temperature near the soil surface, to obtain the upper boundary condition for modelling temperature, and in the upper horizon (e.g. within 5–15 cm) at a point corresponding with moisture measurements, such as at maximum root density (see Figure 1.5.1 in protocol 1.5 Meteorological measurements). In the best of all worlds, a second set of sensors would be placed at greater depth, perhaps near the bottom of the root system. Such positioning would capture the rare drought or snowmelt events that deplete or refill the whole soil profile. In boreal forests, these deeper sensors are often placed at 50 cm below the surface. As much as we would like to standardise these depths, the variation in diurnal/seasonal cycles and in root water depletion depths prevents convergence on a single recommendation.

When installing sensors in a plot it is generally best to install them horizontally (to measure temperature at the desired height) which can be achieved by excavating a small trench from outside the plot into the plot (see Figure 1.5.1 in protocol 1.5 Meteorological measurements). This prevents preferential flow of water along the cables.

Generally, in areas with rodents, it can be useful to protect the wire of a sensor with PVC tubes to prevent damage. In alpine areas, where there is a lot of snow in spring, it can be advisable to protect the wire higher up, because rodents can climb up on the snow.

For each of the sensors, the optimal **sampling interval** is every minute and should be reported in the form of half-hourly to hourly averages. Modelling of ecosystem gas exchange will require half-hourly to hourly data input.

Where to start

Buchan (2001), Deryng et al. (2014), IPCC (2014)

3.5.2 Special cases, emerging issues, and challenges

Since most soils support plant growth naturally, it is important to understand the impact of plant canopies, which can insulate and alter temperature gradients (important in modelling) between the soil surface and the atmosphere. This suggests that both canopy and soil surface temperature measurements are important to consider.

3.5.3 References

Theory, significance, and large datasets

Bond-Lamberty & Thomson (2010), Chapin et al. (1979), Dai et al. (2004), Davidson & Janssens (2006), Hay & Wilson (1982), Jury & Horton (2004), Parlange et al. (1998)

More on methods and existing protocols

ASTM (2017), McInnes (2002)

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3.6 Soil wettability or water repellency

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Measurement unit: s (seconds); Measurement scale: plot; Equipment costs: €; Running costs: -; Installation effort: -; Maintenance effort: -; Knowledge need: low; Measurement mode: manual

Soil water repellency (SWR) is the inability of water to wet or infiltrate a soil. SWR affects water infiltration, soil water retention, and soil water content and thus plant and microbial available water (Doerr et al., 2000; Hallett, 2008; Dekker et al., 2009). Multiple factors are thought to contribute to the development of SWR and include litter breakdown products, such as waxes, root exudates, fungi, and bacteria, and litter structures such as a rugose surface. SWR is most evident when the soil dries below a critical water content, leading to movement and reorganisation of organic solutes at the soil/water interface, often creating a water repellent coating (Wallis & Horne, 1992; Ritsema & Dekker, 1996). SWR has been used in studies characterising land-use change (Lebron et al., 2012), effect of fire (DeBano, 2000b), and in transect studies looking at the patterns and finger flow caused by SWR (Robinson et al., 2010). The distribution of SWR in soils has high spatial variability (Lozano et al., 2013) with isolated patches of SWR and soil water content. Such variability can be critical because spatial isolation may be the key in structuring soil microbial communities (Treves et al., 2003).

Although most noticeable in dry soils, the impacts of SWR are persistent in wet soils, for example reduced infiltration rates and water-holding capacity can occur due to air entrapment (van Dam et al., 1990). SWR has been reported under all climatic and geographic conditions as well as in all soil types: SWR is regularly observed in most land-use and management types, and is often independent of organic matter content and soil texture (Doerr et al., 2006; Dekker et al., 2009). Climate change may increase the prevalence of SWR through increased frequency of extreme events such as drought and storms, or the increased occurrence of fire (Goebel et al., 2011). Given that SWR can induce a shift of soil moisture to an alternative stable state, altering infiltration processes from piston flow and uniform wetting to bypass flow and heterogeneous wetting, it can have a substantial impact on microbial activity, biogeochemical cycling, and gas fluxes.

3.6.1 What and how to measure?

Soil water repellency measurement

Soil water content and temperature are the main factors to control when measuring SWR. Although SWR can be measured in the field, the best, most reproducible results are measured in the laboratory under controlled conditions on air-dry samples. Ideally, an undisturbed soil core from the top 2 cm is collected (this may be a core extracted for bulk density or other measurements). SWR is then measured under a laboratory temperature of 20–25 °C using the water drop penetration time (WDPT, see Figure 3.6.1) test (Van't Woudt, 1959), for example. In the WDPT test, water drops are applied to the soil and the length of time it takes for the water drops to penetrate the soil is measured. A longer duration indicates stronger water repellency. The WDPT is best characterised as measuring the persistence or stability of soil water repellency (Letey et al., 2000). The WDPT test groups soils into six classes according to the time taken for water penetration (see below).

The WDPT test separates soils which are classified as being water repellent from those which are not. Since water penetrates the soil if the contact angle (θ) is less than 90 °, WDPT is a measure of the time required for θ to change from its original value, which was greater than 90 °, to a value approaching 90 °. Therefore, it is a measurement of the stability of the repellency, or persistence. In repellent soils, it takes a considerable time for the drop to penetrate the soil. For this reason, work previously published may have a maximum time of observation, which effectively truncates the data and varies according to different authors. A minimum time of at least 6 h is preferred (see below). In order to reduce observation time the process can be filmed using a microcamera with a timer connected to a computer, which when replayed provides the exact time at which the drop disappears from the soil surface and can easily be reviewed.

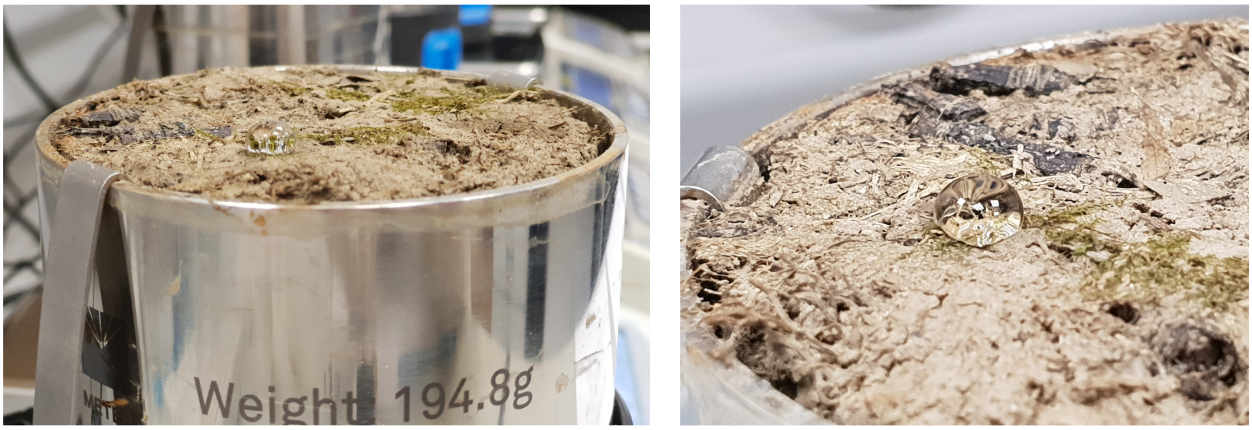

Figure 3.6.1 Measurements of water drop penetration time: A drop of 100 µl deionised water on a vegetation-free soil surface of a dried soil core (left) and a close-up of the drop (right). Photo: Francis Parry (Centre for Ecology & Hydrology).

Example method

- Take the undisturbed, air-dried soil core (top 2 cm) and remove the loose litter or green vegetation from the top of the soil.
- Place the soil-surface sample under a USB microcamera on a stand.
- Pipette at least 6 distinct 100 µl drops of deionised water on to the soil surface, avoiding any vegetation or root matter.
- When all drops have been absorbed into the soil surface ensure all files are saved.
- Record the time at which each drop disappeared from the surface of the soil.
- The median of the six drops constitutes the penetration time for that sample.

Water penetration time for the soil is estimated as the median rather than the average since the median eliminates the bias that one or two drops with very long penetration times can introduce. The use of digital recording for the collection of data eliminates the need to set a maximum time of observation that can effectively truncate the data, in this way we collect the real time for the water to infiltrate even for those soils which are highly water repellent and may take 5 or 6 hours.

Interpretation

Penetration times are used to classify the soil according to different repellency levels. Different authors use different definitions for the various water repellency classes, however most scientists agree that when penetration time is higher than 1 hour, the soil is severely water repellent, and below 5 seconds the soil is not water repellent. It is preferable to report the WDPT in seconds, but there are also WDPT repellency classes. Class 0: non-repellent (infiltration within 5 s); class 1: slightly water repellent (5–60 s); class 2: strongly water repellent (60–600 s); class 3: severely water repellent (600–3600 s); class 4: extremely water repellent (1–3 h); class 5: extremely water repellent (3–6 h); and class 6 (> 6 h) (Dekker et al., 2009).

Where to start

Doerr et al. (2000), Dekker et al. (2009), Letey et al. (2000)

3.6.2 Special cases, emerging issues, and challenges

While initially considered an issue in fire-prone Mediterranean climates, it is becoming clear that SWR is a much broader issue. Understanding the distribution, ecological significance, and mechanisms generating SWR remain a challenge.

Other methods of soil water repellency measurement

The molarity of an ethanol droplet (MED) test (also known as the percentage ethanol or critical surface tension test) is also used to measure SWR (Letey et al., 2000). In the MED test, drops with an increasing concentration of ethanol are applied to the soil in order to indirectly measure the apparent surface tension. This effectively determines how strongly the water is repelled, and this property is best reported as the strength or severity of SWR as distinct from the WDPT persistence (Letey et al., 2000).

3.6.3 References

Theory, significance, and large datasets

DeBano (2000a, 2000b), Doerr et al. (2000), Hallett (2008), Ritsema & Dekker (1994)

More on methods and existing protocols

Hallin et al. (2013)

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3.7 Sap flux

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Measurement unit: kg H_2O m^{-2} sapwoods^{-1}; Measurement scale: plot; Equipment costs: €–€€; Running costs: €–€€; Installation effort: medium; Maintenance effort: medium to high; Knowledge need: high; Measurement mode: data logger

Transpiration is the largest component of the terrestrial hydrologic cycle (Jasechko et al., 2013; Schlesinger & Jasechko, 2014) and forms a critical intersection of the water and carbon cycles. Transpiration and the associated vegetation water fluxes (i.e. root-water uptake, recharge and discharge of biomass water storage, sap flux) are responsible for the rewetting of the atmospheric boundary layer and have been shown to drive cloud formation and downstream precipitation (Hesslerova & Pokorny, 2010; de Arellano et al., 2014). The energy associated with transpiration and evaporation, latent heat flux (LE), is a fundamental component of the Earth's surface energy balance. Over densely vegetated areas, transpiration is the principal component of LE. Properly constraining estimates of transpiration, and thereby LE, is essential for solving the terrestrial surface energy budget and partitioning incoming solar energy into the constituent sensible, latent, and ground heat fluxes. Thermal cooling of the canopy associated with transpiration feeds back into the radiation budget through the Stefan-Boltzmann law and further alters surface-energy flux partitioning and Earth's skin temperature (see Matheny et al., 2017b and references therein). Global climate change may be characterised by changes to precipitation and soil moisture dynamics and increased temperature and evaporative water demands – all affect and are affected by transpiration. Furthermore, changes to CO_2 concentrations that drive climate change also change the water-use efficiency of plants, i.e. how much water is lost per unit of photosynthetic carbon uptake (Keenan et al., 2013). Measurements of transpiration at the ecosystem scale, and more particularly at the species level, are needed to understand how the ecosystem responds and feeds back to climate change.

While LE can be directly measured in the atmosphere above the canopy using the eddy covariance (EC) approach, these measurements are characteristic of a large measurement footprint (~1–10 km^2) and include all evapotranspiration. EC measurements of LE are not specific to transpiration, nor can they distinguish plant species-specific contributions (Baldocchi, 2005). Sap flux (kg H_2O m^{-2} sapwood area^{-1}) is the most common individual-scale measurement of transpiration from woody plants, and can be scalable to the plot scale if measurements are sufficiently replicated. Sap flux measurements are widely accepted as indicators of the magnitude and patterns of transpiration (Poyatos et al., 2016). Such measurements are considered quantitative on a daily time step, but hourly inferences require correction for a time lag to account for the influence of the hydraulic capacitance of the vegetation (Granier & Loustau, 1994; Schäfer et al., 2000; Matheny et al., 2017a).

3.7.1 What and how to measure?

By definition, the term "flux" refers to a flow across an area (kg s^{-1} m^{-2}), and in this case is specifically defined as the flow of sap (kg H_2O s^{-1}) per unit area of sapwood or active xylem (m^2). This measure of sap flux is frequently referred to in the literature as sap flux density (Granier, 1987), while sap flow is typically presented as a rate (i.e. mm H_2O s^{-1}). Sap flux has historically been studied using dyes, radioisotopes, and salts as tracers but it is now most frequently measured using a thermal technique (Marshall, 1958).

Thermometric techniques allow sap flux to be monitored continuously *in situ* without the need for destructive sampling. The most common thermal monitoring techniques are the trunk segment or tissue heat balance (Čermák et al., 1973; Kučera et al., 1977), the stem heat balance for very small stems (Sakuratani, 1981), heat field deformation (Nadezhdina et al., 1998), heat pulse velocity (Swanson & Whitfield, 1981; Dragoni et al., 2009), and thermal dissipation (Granier, 1987). While each method has its own set of strengths and limitations (see Wullschleger et al., 1998; Čermák et al., 2004; Steppe et al., 2010; and references therein), sap flux measurements are considered to be reliable estimates of vegetation water flux and transpiration (Poyatos et al., 2016). These methods differ fundamentally in the manner in which heat is applied and the timing of heat application, yet all are similar in that the rate of heat transfer is mathematically related to the velocity of water movement through the conductive woody tissue (Smith & Allen, 1996). In field experiments, it is recommended to measure at least five replicate trees per plot; if the site is heterogeneous (soil depth, slopes, multiple species, etc.) it is desirable to increase the number of measured trees.

Gold standard

The trunk segment or tissue heat balance (THB) method requires no calibration and is frequently used as a standard against which other thermal tracer methods are evaluated and/or calibrated (Lundblad et al., 2001; Renninger & Schäfer, 2012). Commercially available THB sensors consist of three or more uniformly, continuously heated, plate electrodes and one unheated "reference" electrode located vertically below the central heated electrode (Figure 3.7.1 left). Temperature sensors located in the centre of each plate electrode measure temperature differences between the plates. The THB method calculates the temperature balance of the well-defined heated trunk volume in a manner that accounts for conductive transfer through the woody tissue and convective transfer due to water flow (Čermák & Nadezhdina, 2011). THB is suitable for large trees and can also be applied to smaller diameter stems (0.6–2 cm) using modified sensors (e.g. environmental measuring systems (EMS) "baby" sensors, Figure 3.7.1 right). Measurement inaccuracies have been reported during low flow conditions and in cases where the sensors are installed into stems with narrow sapwood (Čermák et al., 2004). THB sensors require more power input (0.6–1 W) than other methodologies.

Figure 3.7.1 THB sap flux sensor (left) and "baby" THB sensor manufactured by EMS (right).

Bronze standard

The heat field deformation method (HFD) (Nadezhdina et al., 2008) is a constant-heat technique that consists of one heated needle and three measurement needles (1–1.5 mm diameter), each containing several thermocouples. Generally, measurement needles are placed above, below, and to the side of the heater (Nadezhdina et al., 2012; Figure 3.7.2 left), although alternative configurations have been used (Čermák et al., 2004). Sap flow is calculated based on the ratio of temperature differences between the vertical and lateral thermocouple pairs. Commercially available HFD sensors contain multiple (up to ten+) thermocouples along their length and are ideal for monitoring sap flow at multiple radial depths (Nadezhdina et al., 2012). HFD sensors produced by ICT International require 0.06–0.08 W of electrical power per cm of sensor length (Nadezhdina et al., 2012).

Figure 3.7.2 HFD sensor HFD8-100 manufactured by ICT international (left) and HPV sensor manufactured by Edaphic Scientific (right).

The HFD method is capable of monitoring reverse flows and low flows without the need to assume a no-flow condition at night (Nadezhdina et al., 2012). In an intercomparison of sap flux measurement techniques, Steppe et al. (2010) demonstrated that the HFD method tends to underestimate sap flux and is very sensitive to sapwood depth. Results from Steppe et al. (2010) show the heat pulse velocity method (HPV) to be slightly more accurate than HFD, but only after an empirical correction for wounding effects (Swanson & Whitfield, 1981) is applied. Commercially available HPV sensors are typically low power and consist of a heater bracketed vertically above and below the temperature sensors (Burgess et al., 2001; Figure 3.7.2 right). Sap flux is calculated proportionally to the velocity of a heat pulse between the two temperature sensors (Marshall, 1958). Similarly to HFD, accuracy of HPV techniques can suffer during low flow conditions and they are prone to higher noise than HFD or thermal dissipation techniques (Steppe et al., 2010).

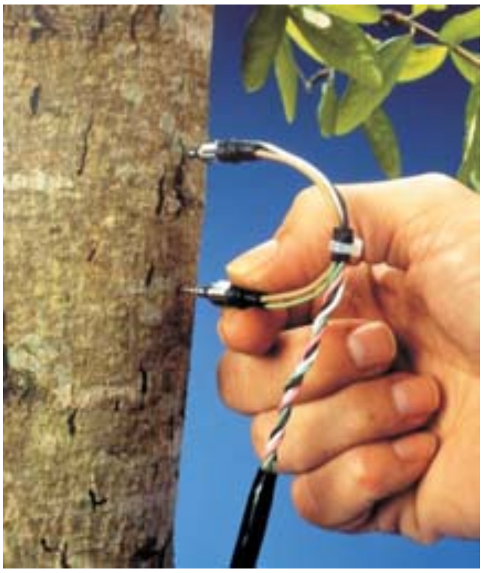

Figure 3.7.3 A thermal dissipation sensor as manufactured by Dynamax.

The thermal dissipation (TD) method (Granier, 1987) is the most broadly used methodology for sap flux measurements (Poyatos et al., 2016). TD sensors consist of one unheated reference needle and one heated needle, each containing a thermocouple (Figure 3.7.3). The temperature difference between the upper heated needle and the lower unheated needle is empirically related to sap flux density (Granier, 1985). TD sensors are frequently manufactured in-house at very low cost and require a relatively low power use of 0.2 W, making them ideal for studies with high measurement replication. TD sensors are known to underestimate sap flux (Steppe et al., 2010; Renninger & Schäfer, 2012), but post-processing corrections for needle penetration into non-conductive wood (Clearwater et al., 1999) and nocturnal flow (Oishi et al., 2008) can improve accuracy. They are sensitive to radial and circumferential variation in sap flow rate within the stem. With careful data processing (available open-source through Oishi et al., 2016), this method can yield quantitative measurements without calibration in the lab prior to deployment. Measurements from TD sensors have been shown to align with data collected through porometery, whole-tree gas exchange, gravimetric water loss, above and below canopy EC measurements, and estimates from the Penman-Monteith equation (Renninger & Schäfer, 2012).

How to install and measure

Considerations of the methodology frequently include the purpose of the campaign (i.e. quantitative individual performance or upscaling), electrical power supply and consumption, maintenance capabilities, and the number of potential replicates. All sap flux measurement techniques are sensitive to installation location. Measurement points should be carefully selected and should not be in areas near knots or defunct branching points, as flow around these regions is generally altered. Conductive sapwood depth is known to vary circumferentially around the tree, and sap flow velocity is known to vary radially (Phillips et al., 1996). Typically, if only one measurement point is being made on an individual, sensors are installed on the north face (in the Northern Hemisphere; south face in Southern Hemisphere) of the stem to avoid direct heating by solar radiation. Typically sensors are covered by reflective insulation after installation as additional protection against external solar heating. In the case of multiple instrumentation points on a single tree, tertiary sensors can be located circumferentially or radially to account for variability in flux, or along branches to monitor lag time in flow rates. It is also generally recommended for all techniques, with the exception of heat balance methods, that measurements are made at multiple sapwood depths to account for radial variations in flow. Many commercially available sensors come with proprietary data logging systems (e.g. ICT international, EMS), while others (e.g. Dynamax, Edaphic Scientific) are compatible with common data loggers such as Campbell Scientific or HOBO. Proprietary loggers typically are installed adjacent to the sensor on the tree and are paired such that one logger records data from only one sensor. Non-proprietary loggers can be used to record data from multiple sensors simultaneously and are generally located at an easily accessible point central to the measurement plot. Data is usually recorded continuously at intervals of 15 minutes or less with 5 minutes being fairly typical. Depending on the species phenology measurements can be made year round or during, including immediately before and after, the growing season. The length of the measurements campaign is inherently dependent on the research question being explored, most typically campaigns last on the order of a few months to several years.

Where to start

Wullschleger et al. (1998) provide a good overview of water flux measurements in trees, while Steppe et al. (2010) provide comparisons of several of the leading methods. Guides for processing data are provided by Oishi et al. (2008, 2016). Matheny et al. (2014) provide guidance for collecting large numbers of replicates for scaling tree-level measurements to the plot scale.

3.7.2 Special cases, emerging issues, and challenges

In the case of HPV, and long-term measurement campaigns with other methods, the effects of wounding should be accounted for either numerically or through periodic sensor relocation during maintenance. Maintenance of sap flux instrumentation in the field is moderate to difficult: custom-made sensors require marginally more maintenance than commercial ones. Tropical environments with high wood-boring insect and rodent activity typically require maintenance that is more intensive, with a recommended minimum maintenance period of less than six months. Data collection for sap flux sensors is typically automated through a combination of data loggers and multiplexers, in the case of numerous replicates, with most measurements occurring on a 1–10 minute basis. Collected data are typically processed from the raw signal (usually mV) to sap flux through commercial or open-source software packages in MATLAB or R (Oishi et al., 2016). Upscaling sap flux measurements to the canopy or forest level can be accomplished using biometric and allometric relationships but this requires a large sample and representation of all significant species in the plot (Wullschleger et al., 1998; Matheny et al., 2014).

3.7.3 References

Theory, significance, and large datasets

Granier (1987), Poyatos et al. (2016), Steppe et al. (2010)

More on methods and existing protocols

Clearwater et al. (1999), Oishi et al. (2016), Phillips et al. (1996)

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3.8 Ecosystem water stress

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Measurable unit: ratio (unitless); Measurement scale: site; Equipment costs: €€; Running costs: €; Installation effort: medium; Maintenance effort: medium; Knowledge need: medium to high; Measurement mode: data logger

Water stress (WS) in terrestrial ecosystems develops when soil water is depleted below a threshold. The depletion of water, or water deficit, implies decreases in the soil water potential that propagate through the soil–plant–atmosphere continuum, typically reducing activity of plants and of soil organisms. WS thus has a strong impact on ecosystem functioning and, in the longer-term, also on ecosystem structure. A few indices to quantify WS have been proposed aiming to indicate how vegetation activity is constrained by water depletion (Myers, 1988; Granier et al., 1999; Vicca et al., 2012). WS indices incorporate information on precipitation characteristics, which make them useful in analysing climate treatment effects where the timing and intensity of rain differs and for between-year comparisons in observational studies. WS indices also incorporate information on soil properties (soil water holding capacity) and rooting depth (vegetation exploration of the soil), which facilitates across site comparisons, i.e. along climatic gradients or to compare land-use change and cover.

3.8.1 What and how to measure?

Indices of plant WS can be obtained from direct measures of plant water potential (i.e. Myers, 1988; see protocol 5.9 Psychrometry for Water Potential measurements). Alternatively, whole ecosystem WS can be estimated from measurements of soil water content (SWC). Granier et al. (1999), worked with the relative extractable water (REW) defined as the ratio of the actual content of extractable water to the maximum amount that can be extracted over the entire soil profile reached by roots. They considered that WS starts when the actual water content decreases below 40% of the maximum water content, i.e. REW < 0.4. This threshold was defined empirically after the value above which the transpiration to potential evapotranspiration ratio keeps constant, and below which the ratio decreases linearly as SWC decreases, indicating stomatal closure to regulated transpiration (Granier et al., 1999). Hence, at RES < 0.4, changes in SWC affect plant functioning, while above this threshold, SWC dynamics have little influence on plant activity (Granier et al., 1999) unless the soil is water saturated.

Long-term monitoring of SWC across the soil profile using TDR sensors (see section 3.1 above) may be limited by economic constraints for the instrumentation in several plots, or by the need to preserve the integrity of the experimental plots. A less invasive alternative is a plant–soil water budget model. The modelling of the SWC requires daily meteorological data and the description of the soil and vegetation properties at least at site level (Granier et al., 2007; Longepierre et al., 2014). Modelling allows the estimation of water balance both in reference plots under the natural water regime and in plots with modified precipitation. Modelling the water balance in climate-change experiments, especially under precipitation-manipulation treatments, is highly advised because it completes the description of the variability in ecosystem water relationships. Modelling results help with the interpretation of treatment effects, between-year variability, and facilitate the integration of the experiments into wider across-sites syntheses.

A mid-term option is recording SWC to root depth at site level outside the manipulation plots. This reduces instrumentation costs and perturbation within experimental plots, provides field data for across-site comparisons and, at the same time, can improve the estimation of the SWC within the treated plots through fine-tuning of the parameters during modelling. If experimental manipulations or equipment change the SWC over time inside the experimental plots differently than outside the plots, this method cannot be used. This is the case in, for example, long-term experiments altering rain.

Gold standard

Measuring SWC. Requires the measurement of SWC at all soil layers reached by the roots and the characterisation of the size of the soil water reservoir at each layer down to maximum root depth (we consider a layer the whole horizon if it is fully reached by roots or only the fraction of the horizon down to maximum root depth (see Vicca et al., 2012). The amount of maximum extractable water for plants for a soil layer (EW_{max}) is defined as:

$$EW_{max} = (SWC_{fc} - SWC_{wp}) * \text{thickness of soil layer}$$

where SWC_{fc} is the soil water content at field capacity for the layer and SWC_{wp} is the soil water content at wilting point. The water that can be extracted from a soil layer at a specific soil water content (SWC) is

$$EW = (SWC - SWC_{wp}) * \text{thickness of soil layer}$$

The maximum total amount of extractable water (TEW_{max}) and the actual total amount of extractable water (TEW) integrate the entire rooting zone and are thus calculated as the sum of all the n soil layers.

$$TEW_{max} = EW_{1max} + \ldots EW_{nmax}$$

$$TEW = EW_1 + EW_2 + \ldots + EW_n$$

The relative extractable water (REW) is defined as:

$$REW = TEW / TEW_{max}$$

Two WS indices can be estimated: i) the duration of the WS, as the number of days when REW < 0.4, with 0.4 the threshold value for the development of stress, and ii) the intensity of the WS (IWS) calculated as

$$IWS = \text{sum}(\text{maximum}[0; (0.4 - REW) / 0.4])$$

Other methods

Modelling soil water content. Water balance models require continuous meteorological data and parameters describing the soil and the vegetation. A simple model uses the daily precipitation and the estimated potential evapotranspiration obtained from meteorological measurements at site level, or within the plots in case the precipitation or temperature are altered by the manipulation. Soil data to estimate maximum SWC and drainage is required for soil layers down to the rooting depth. Required vegetation description includes maximum leaf area index (LAI) and its phenology if LAI is not constant (e.g. in deciduous species), maximum rooting depth, root density across soil layers and canopy coverage.

Where to start

Granier et al. (1999) defined the method and justified the threshold for relative extractable water and Vicca et al. (2012) suggested using it to compare experiments. Good examples on its application in combination with water balance models are in Granier et al. (2007) and Longepierre et al. (2014). Myers (1988) defined an index of plant water stress from the cumulative integral obtained by frequent measurements of predawn plant water potential.

Installation, field operation, maintenance, interpretation

An initial identification and description of the soil horizons are needed to decide the number and depth of soil moisture probes. See protocols 1.2 Soil type and physical characteristics and 3.1 Soil moisture. For practical reasons of instrumentation, some authors restrict the measurements and calculations to the profile depth were 80% of fine roots occur (Gebauer et al. 2012).

Lateral water flow at deep soil layers may occur especially at the interface between soil and bedrock. Lateral flow can be unravelled by the modelling, for instance as described in Longepierre et al. (2014).

For modelling, see also protocols 4.5 Aboveground plant phenology for leaf phenology and 1.2.1 Soil types and horizons, 4.16 Functional traits and Pérez-Harguindeguy et al. (2013) for rooting depth distribution and LAI (Leaf Area Index).

3.8.2 Special cases, emerging issues, and challenges

Punctual measurements of pre-dawn plant water potential are a good complement. Additionally, provided minimum water potential for extracting water by plant species is available, plant water potentials can be modelled from the modelled soil water (e.g. Longepierre et al. 2014).

Determination of maximum root depth may be challenging in deep-rooted ecosystems and when roots enter rocks.

3.8.3 References

Theory, significance, and large datasets

Allen et al. (1998), Breda et al. (2006), Granier et al. (2000), Piedallu et al. (2011)

More on methods and existing protocols

Gebauer et al. (2012), Longepierre et al. (2014), Vicca et al. (2012)

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