



Monitoring soil water content and deficit using Electrical Resistivity Tomography (ERT) – A case study in the Cevennes area, France

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SUMMARY

At the hillslope scale, surface runoff depends to a great extent on the water deficit of the soils, i.e. the difference between the effective porosity and the water content, all through the soil profile. The aim of this study was to assess water deficits using Electrical Resistivity Tomography (ERT). ERT offers several advantages: (i) it is not intrusive, (ii) it exhibits spatial patterns at a decametric or hectometric scale, and (iii) it provides information on both water content and soil depth. The study was conducted on a Mediterranean shaley sandy soil, in the Cevennes (South of France). First, the factors (nature of the pores and organization of the material, water content, soil solution and temperature) on which electrical resistivities depend were analyzed in the laboratory. Soil samples were used to derive the 'm' and 'n' parameters of Archie's relation, changes in the soil solution over time, and the effect of temperature on resistivities. These results were then used to interpret ERT *in situ* at different times of the year. The water content and water deficit obtained from ERT were then compared with local measurements made using Time Domain Reflectometry (TDR) and the comparison was satisfactory. The results of this study show the potential of measuring soil water content and water deficit using electrical resistivities, and describe how uncertainties in temperature, porosity, and fluid conductivity impact ERT-obtained estimates of soil water content and deficit.

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Introduction

Flood processes can be related either to the excess surface infiltration or to the excess storage capacity of the soils. The latter case is typical in temperate climates (Beven and Kirkby, 1979; Ambroise et al., 1996), but is also found in Mediterranean areas (Cosandey, 1999; Bouvier et al. 2007) and is particularly true of the shaley Cevennes hillslopes. The soils are only a few tens of cm thick, but have very high hydraulic conductivity and can store up to dozens or even hundreds of millimetres of rain before flash floods occur during the most intense rainfalls. The water deficit of the soil, which corresponds to the difference between the porosity and the volumetric water content along the vertical profile, is thus a key element to predict hillslope runoff and river floods.

However, it is difficult to measure this deficit, in particular soil water content. Except for volumetric weighing, the measurement of water content is generally indirect, and requires interpretation and – to a varying extent –, calibration. Locally, electrical measurements by capacity or Time Domain Reflectometry (TDR) allow reasonably precise measurement of water content (Robinson et al.,

2003; Walker et al., 2004; Laurent et al., 2005); the disadvantages of these measurements are related to the limited volume of investigation, and to the fact that the installation of probes can modify the soil structure. In contrast, radar measurements of soil water content by satellite cover much bigger areas, but (i) their spatial or temporal resolution is not always compatible with the monitoring at the hillslope scale, (ii) their depth of investigation does not exceed a few centimetres (Pellarin et al., 2006), and (iii) this measurement is influenced by other factors such as the vegetation and microrelief (Wagner et al., 2007).

Geophysical methods (electromagnetism, ground penetrating radar, electrical resistivity) do not affect the soil structure, and the resulting measurement overlays a first level of spatial variability, at a decametric or even hectometric scale. Electrical Resistivity Tomography (ERT) has been widely used for hydrogeological investigation (Binley and Kemna, 2005); recently, it was also used to obtain detailed images of water content and of the structure near the surface or/and a representation of hydrological processes. Michot et al. (2003) showed that ERT was able to produce a good delimitation of soil horizon as well as to monitor soil water movement during irrigation of a corn crop. When combined to hydrologic classic measurements and tracers, ERT contributed to identify the flow pathways at the hillslope scale (Wenninger et al., 2008; Koch et al., 2009). Schwartz et al. (2008) used 2D-electrical resistivities

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imaging, 1D TDR-measurement of soil water content and physico-chemical properties of the soil to produce a distribution of 2D soil water content in a heterogeneous clay rich soil. However, electrical resistivities are known to be sensitive to other physical factors (temperature, soil solution (Friedman, 2005; Besson et al., 2008), soil mineralogy, particle-size distribution, pore connectivity, percentage of clay (Samouëlian et al., 2005)). Furthermore, numerical artefacts due to electrode arrays or mathematical inversion of the apparent resistivities can make interpretation complicated (Dahlin and Zhou, 2004). Because of this complexity, and because experimental studies that calibrate the relationship between soil water content and electrical resistivity are still rare, ERT measurement of soil structure or water content are difficult at a site where no previous works have been made e.g. the shaley hillslopes of the Cevennes.

The objective of this study was thus to make a regional evaluation of the performances of ERT measurements in producing both temporal and spatial patterns of the water deficit, by comparing them with local TDR measurements. Special attention was paid to the conditions required for a more systematic ERT survey in this region in the future. Moreover, in most studies that combine TDR and ERT, TDR field measurements are generally used for calibration of the relationship between electrical resistivities and water content. Here, we also dealt with the ability of ERT to derive soil water content measurements from laboratory-based calibration, inde-

pendently of the TDR measurements. This paper presents (i) the experimental site and the data collected *in situ*, (ii) the calibration of the soil water content in the laboratory, (iii) the results concerning ERT soil water content and water deficits, and their comparison with local TDR measurements, and (iv) the uncertainties related to the ERT measurement.

Study site

The experiments were performed on a small plot (about 400 m²) near Sumène, which is located in a shaley area in the southern Cevennes (Fig. 1). The coordinates of the centre of the plot are 43°59'08.15" N and 3°43'04.12" E, the altitude 235 m. The region is characterized by a Mediterranean climate, with precipitation occurring mostly in the autumn period. For the period 1981–2008, the mean annual rainfall recorded by the Météo-France raingauge located in Sumène was 1248 mm. The maximum mean monthly rainfall occurs in October (193 mm), and the minimum is in July (38 mm). In July, the mean of the minimum and maximum temperature is 15 °C and 31 °C (July), respectively, while in January, it is 0 °C and 10 °C, respectively (period 1999/2008). In this area, mean annual evapotranspiration can be compared to those of other sites having similar characteristics in the same area, between 1090 mm (Roujan, alt. 120 m, 60 km SW) and 1144 mm (La Grand-Combe, alt. 288 m, 35 km NE).

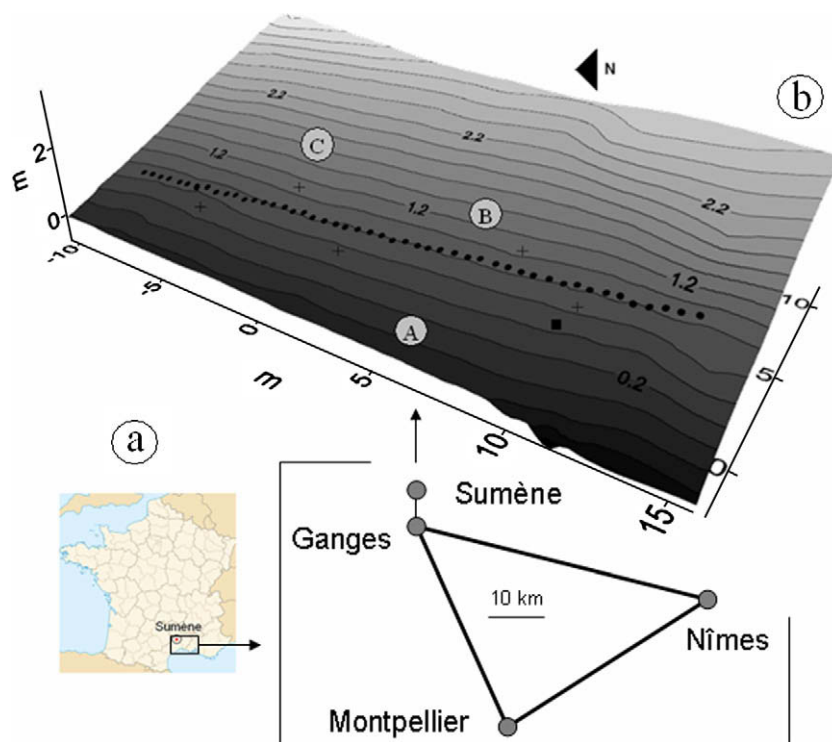


Fig. 1. (a) location of the study site and (b) sketch of the experimental field site. ● electrode line; ○ TDR tubes; ■ rain gauge and temperature probe at a depth of 80 cm, + five auger holes for granulometric measurement. Samples for the analysis of porosity were taken near A and B. Three temperature probes (–20, –40 and –65 cm) were located near C.

Table 1
Granulometric composition of the soil. Percent values are averaged values derived from five vertical profiles, at three depths. (1.08) standard deviation.

	Clay % < 2 μm	Silt 2 μm < % < 20 μm	Sand 0.02 mm < % < 0.2 mm	Coarse sand 0.2 mm < % < 2 mm	Coarse fragment % > 2 mm
Surface	2.5 (0.6)	13.8 (1.6)	15.7 (2.6)	21.6 (3.4)	46.4 (1.1)
–30 cm	1.8 (0.3)	12.6 (1.1)	17.4 (3.7)	22.2 (4.2)	46.0 (2.2)
–60 cm	2.8 (0.3)	14.9 (1.3)	15.1 (1.6)	20.6 (3.0)	46.6 (1.2)
Average	2.9 (0.4)	13.9 (1.6)	14.6 (1.7)	22.1 (3.4)	46.5 (0.9)

The soils in the plot are a little more than 1 m thick, and correspond to old cultivated terraces (vineyards); this type of farming was abandoned more than 20 years ago. The slope of the plot was regular (about 15°). The granulometric analysis was derived from five profiles, using laser granulometry for particle size under 2 mm (Table 1). No significant differences were found between samples or with increasing depth (from the surface to –60 cm). The coarse fragments represented 45% of the total sample weights. According to the FAO data base, the soil can be classified as Regosol. Measurements of porosity were performed with 100 cm³ cylinders (diameter 53 mm), which gave coherent soil samples along two vertical profiles from the surface to –80 cm. The samples were first saturated in water and then dried at 105 °C during 24 h; porosity was calculated from the differences between saturated weight and dried weight of the sample. The values decreased with the soil depth, from 0.52 to 0.37 for the first profile (average 0.42, standard deviation 0.06) and from 0.51 to 0.41 for the second profile (average 0.42, standard deviation 0.04).

Experimental device

More than 10 ERT measurements were made between February 2006 and December 2007 using a fixed line of 48 electrodes, regularly spaced at 50-cm intervals. This geometry was chosen to get a full characterization of the plot width, while sounding the first upper meter with a satisfactory sensitivity. The line of electrodes was perpendicular to the slope of the plot (Fig. 1), so that all the electrodes were located at the same altitude. The acquisitions were made with a Syscal Junior Switch 48 (IRIS Instruments), using a 510 quadripoles Wenner Schlumberger sequence. TDR measurements of water content were performed along three vertical profiles at different depths using the “TRIME-tube system” (Laurent et al., 2005) combining a commercial TDR moisture meter (TRIME-FM3, Imko GmbH, Ettlingen, Germany) with a cylindrical probe (T3 probe also from Imko). These profiles were located less than 5 m from the line of electrodes used for ERT (Fig. 1). The temperatures were measured at four depths ranging from the surface down to a depth of 80 cm: 20 cm, 40 cm, 65 cm and 80 cm (80 cm deep measurement was not available for all ERT).

Calibration of the resistivity/water content relationship in the laboratory

The relationship between the electrical resistivity and the soil water content was calibrated in the laboratory in order to set the parameters of Archie's law and to characterize the different factors involved in the calculation of soil water content. Archie's law is an empirical relationship which links electrical resistivity with the volume and the organization of the pores, water content, and pore water resistivity.

If Φ denotes the porosity of the material, ρ_w the electrical resistivity of the pore water, and S the water saturation of the soil (S equals the ratio of soil water content to soil porosity), Archie's law can be expressed by:

$$\rho = a\rho_w\Phi^{-m}S^{-n}$$

where a , m and n are empirical coefficients (Telford et al., 1990). The m parameter, known as the coefficient of cementation, can range from 1.2 to 4. For shaley soils whose porosity ranges between 0.41 and 0.48, the literature gives $m = 1.52$ (Friedman, 2005). The n parameter, known as the coefficient of saturation, is less variable, and usually takes the value $n = 2$, when the fluid concerned is water (Ward, 1990). For unconsolidated granular material, the constant a is set equal to 1 (Frohlich and Parke, 1989). Archie's law is considered to be valid for medium to coarse-grained soils and rocks, and

because sand and coarse fragments represented over 80% of the mineral soil, we assumed that Archie's law would be valid for the soil in Sumene plot.

The electrical resistivities of the soil samples were measured in 8-electrode cylindrical boxes (diameter = 90 mm, height = 40 mm, Kalinsky et al., 1993). The cylinder and bottom wall were made of PVC. The bottom wall was perforated to allow excess water to drain off. The electrodes were stainless steel rods 4 mm in diameter. The sharp end of the electrodes was inserted to a depth of 5 mm in the sample. The geometric factor for disaggregation of the apparent resistivities was determined using water samples with known resistivity. Resistivity measurements were performed with the same Junior Syscal Switch 48 instrument used for the field experiment.

Temperature control

As the resistivities depend on the temperature, a temperature sensor was also placed in each soil sample, and the temperature was taken into account in all the measurements. The resistivities were corrected to a standard temperature of 25 °C, using Keller and Frischknecht (1966):

$$\rho_T = \rho_{25} / (1 + \alpha(T - 25))$$

where ρ_T denotes electrical resistivity at temperature T , ρ_{25} electrical resistivity at $T = 25$ °C; α is an empirical coefficient that often equals 0.025 °C⁻¹. In these conditions, a deviation of 1 °C in the temperature leads to a deviation of 2.5% in electrical resistivity. The influence of temperature is thus very important for the interpretation of measurements of electrical resistivity in terms of soil water content.

Keller and Frischknecht's coefficient α was experimentally checked in the laboratory. One soil sample with known water content was cooled from 26 °C to 6 °C, and electrical resistivity was measured for each 4 °C increment in temperature. The value $\alpha = 0.025$ °C⁻¹ was found to fit the experimental curve. Thus, in the following, the resistivities have always been corrected to a standard temperature of 25 °C using $\alpha = 0.025$ °C⁻¹.

Measurement of pore water resistivity

The aim of the measurement was to assess the evolution of the pore water resistivity in time. A first goal was to know how pore water mineralization would increase with time, from low mineralized water (like rain water). A second goal was to find initial water mineralization conditions that would not change with time, making more reliable the interpretation of relationship between water content and electrical resistivity of the soil samples during the experiment. The pore water resistivity experiments were performed in 50-cc test tubes. The extraction of pore water was destructive. Each measurement of pore water resistivity thus corresponded to one sample. Soil samples were first saturated with solutions with different degrees of mineralization; after increasing periods of contact with the soil, pore water was extracted by centrifugation, and its resistivity was measured (Fig. 2). Initially low mineralized water tended towards a threshold value of 22 Ω m resistivity, reached after less than 100 h. When initial resistivities did not exceed this threshold, the evolution in time was very limited. Consequently, water mineralization lower than 22 Ω m was always used for the calibration of Archie's law, so that no variation in mineralization would occur during the calibration procedure.

Porosity

For each soil sample, porosity was measured at the end of the experiment by weighing samples after drying and saturation. The

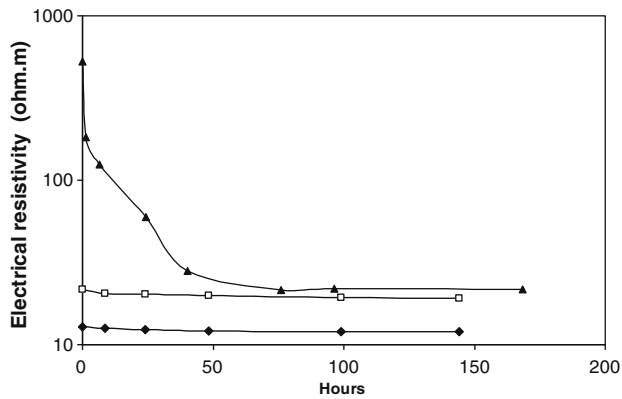


Fig. 2. Evolution in time of pore water resistivity. Soil samples have been saturated with different solutions: $\blacktriangle$ osmosed water (rain like) initial resistivity 526 Ω m, $\square$ initial resistivity 21 Ω m, $\blacklozenge$ initial resistivity 13 Ω m.

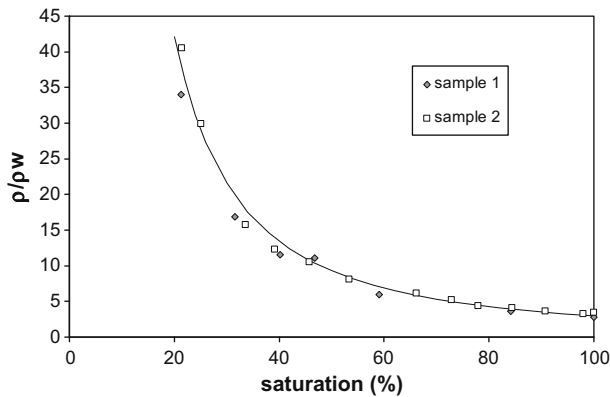


Fig. 3. Relationship between electrical resistivity and water saturation. The first soil sample was wetted with a 17 Ω m solution; the second soil sample with a 25 Ω m solution. Archie's law: $\rho/\rho_w = 2.96 \cdot S^{-1.65}$ ($2.96 = 0.42^{-1.25}$). R^2 to sample 1 and to sample were higher than 0.99 for both samples.

measured porosity, 0.41 and 0.42, was consistent with *in situ* measurements.

Archie's law was calibrated measuring the electrical resistivities of the soil samples with increasing water content by progressive wetting (Fig. 3). The initial water resistivity ($<25 \Omega$ m) was known and assumed to be invariant during the process. The water content of the sample at a given time was measured by weighing. However, it was not possible to obtain reliable measurements of electrical resistivities for soil saturations of less than 20% because under this threshold, the contact resistance between the soil and electrodes was too high. The two samples gave very similar results, which allowed us to derive $m = 1.25$ and $n = 1.65$ for the calibration of Archie's law. Although other parameters could also fit (for example

$m = n = 1.5$) for these samples, $n = 1.65$ was assumed to have more physical sense, because the calibration of Archie's law with additional samples of disturbed soils matched this value better. The values of m and n are a little less than the references found in the literature for shaley soils: $m = 1.52$ and $n = 2$.

ERT results

The Res2Dinv software package was used to calculate ERT resistivities. The main options used in this study were: an isotropic smoothness constraint, a Gauss-Newton regularization method with fixed parameters, a flatness ratio was 0.5 because we wanted to find horizontally elongated structure, and the model refinement, cells width being half the unit spacing (Loke, 1999), i.e. 25 cm and cell height ranging from 12.5 to 22 cm. All ERT inversions show a RMS error smaller than 1%, indicating reliable resistivity models. The slope of the plot was considered to have no influence on the ERT measurements.

The temperatures were taken into account to correct the electrical resistivity to a 25 $^{\circ}\text{C}$ standard temperature. In order to calculate the soil temperatures deeper than 80 cm, or for missing data at depth 80 cm, a model derived from Musy and Soutter (1991) was used:

$$T(z, t) = T_{\text{mean}}(\text{air}) + A/2e^{-(z/d)} \sin(wt + \phi - z/4d)$$

where $T(z, t)$ denotes the temperature at day t and at depth z , $T_{\text{mean}}(\text{air})$ the yearly mean value of the air temperature, A the yearly amplitude of the air temperature, z the depth of the soil, d the depth parameter of the model, $(\phi - z/4d)$ the phase lag, w the frequency of the periodic variation P ($w = 2\pi/P$, where the period $P = 365$ days). The difference with Musy and Soutter's original model stands in $z/4d$ instead of z/d . From 10 years of recorded air temperatures, we obtained $T_{\text{mean}}(\text{air}) = 14 \text{ }^{\circ}\text{C}$, $A = 26 \text{ }^{\circ}\text{C}$, $\phi = -w \cdot t_0$ with $t_0 = 111$ days as the medium temperature occurs on April 21st. The model was applied successfully at the depths 20, 40 and 65 cm, using $d = 1.33$ m. At a depth of 65 cm for example, the deviation in the temperature did not exceed 2 $^{\circ}\text{C}$ (Fig. 4); as the temperature variations became smaller and smoother with depth, we could assume that the difference between the calculated and real temperature was less than 2 $^{\circ}\text{C}$ deeper than 65 cm.

Each ERT exhibited the same structural organization of the resistivities, which were higher in both the upper and the deepest horizons than in the intermediate horizons (Fig. 5 displays the six first ERT). In the upper horizons, the high resistivities can be explained by evaporation due to the vegetation and root area; in the deepest horizons, the high resistivities are due to the presence of bedrock.

Assessment of the water content and water deficit from ERT measurements

Both water content and water deficit could then be derived from the *in situ* electrical resistivities corrected to 25 $^{\circ}\text{C}$, using



Fig. 4. Comparison of measured and calculated temperatures at 65 cm depth ($R^2 = 0.96$).

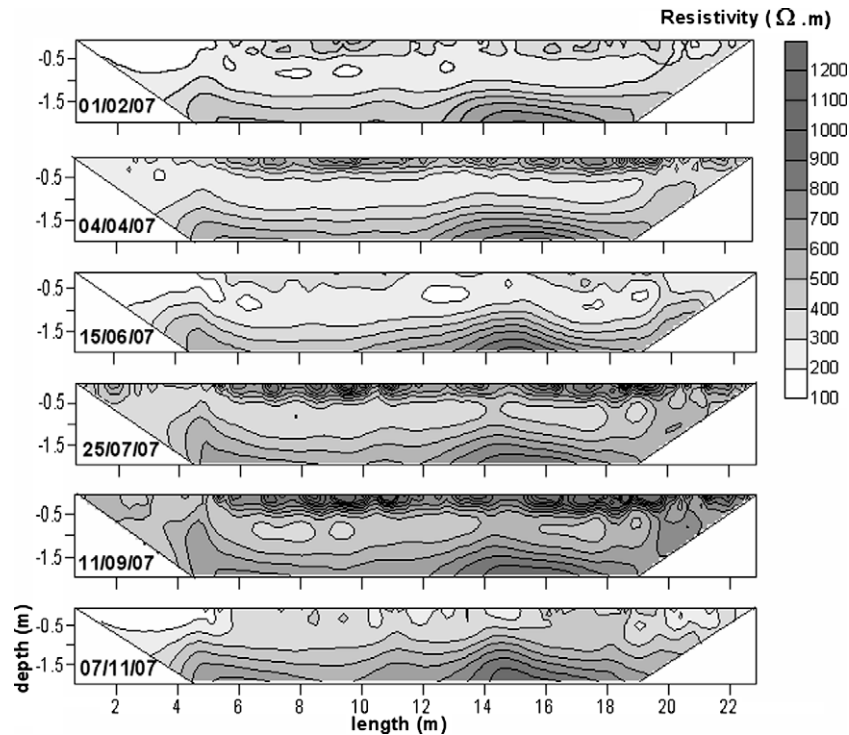


Fig. 5. ERT resistivities in space and over time, during year 2007. Resistivities have been corrected to a standard 25 °C temperature.

Archie's law calibrated in the laboratory. The water deficit of the soil was considered to be the difference between the porosity and the water content throughout the vertical profile:

$$\text{deficit} = \delta = \int_{-\infty}^0 (\Phi(u) - \theta(u)) \cdot du,$$

where Φ denotes porosity and θ water content.

The following assumptions were made:

- The laboratory calibration was applicable to the whole plot at all soil depths. Archie's law was thus applied using $m = 1.25$ and $n = 1.65$, for all the soil depths down to 1.20 m, which was taken to be the average depth of the bedrock; this value was estimated from some ten auger holes in the plot. The bedrock was assumed to have negligible porosity, probably no more than 1% or 2% (based on observations at neighbouring sites). Thus, the water storage capacity was expected to be very low below 1.20 m.
- Because the soils profiles were found to be relatively homogeneous, a constant porosity was used for the whole ERT.

This porosity was set from the averaged porosities measured along the two profiles surveyed, (0.42).

- The resistivity of soil solution was assumed to be close to 22 Ω m, because the ERT measurements were always performed more than 100 h after the previous rainfall (which is more than the time needed for pore water conductivity to stabilize in the laboratory). This value was also compared with the water resistivity in springs located near the plot, which yielded about 20 Ω m most of the year.

Based on these assumptions, the soil water content was calculated for each cell of ERT, 25 cm in width and 12.5–22 cm in height (using the optional model refinement of RES2Dinv). The calculations were made up to a depth of 1.20 m.

The deficits were then calculated along the 2D ERT profile by assuming that the inverted electrical resistivity values represented a suite of vertical soil columns, each with a width of 0.25 m and a height of 1.2 m, (Table 2), i.e. 70 columns along the line of electrodes. For each ERT we give the minimum and maximum temperatures (at the cell scale), the minimum and maximum deficits (at

Table 2

Soil water deficits calculated by ERT; $T_{\min}$, $T_{\max}$: minimum and maximum observed or calculated temperature values for a given vertical profile; δ_{mean} : mean value of the water deficit calculated from the 70 vertical profiles for a given date; $\delta_{\min}$, $\delta_{\max}$: minimum and maximum deficit value among the 70 vertical profiles, for a given date; σ : standard deviation of the 70 vertical profiles, for a given date.

Date	$T_{\min}$ (°C)	$T_{\max}$ (°C)	δ_{mean} (mm)	$\delta_{\min}$ (mm)	$\delta_{\max}$ (mm)	σ (mm)
20/07/06	20	25	365	325	402	17
06/10/06	16	19	290	240	340	21
01/02/07	2	8	273	211	308	20
04/04/07	9.5	12	304	250	334	16
04/05/07	13	14	280	200	308	27
15/06/07	18	19.5	270	194	318	26
25/07/07	20	21	340	310	369	12
11/09/07	19	21	365	340	389	12
07/11/07	9	13	312	222	348	27
05/12/07	7	11	279	226	308	18

the column scale), the average deficit in the whole ERT section, and the standard deviation of the deficits for all the columns of soils for a given ERT.

The averaged water deficits ranged from 270 mm in the wettest conditions up to 365 mm in the driest conditions, which means about 100 mm in absolute deviation. Considering the minimum and maximum deficits at the soil column scale, the extreme values were 402 and 194 mm, which is twice the mean absolute deviation.

Comparison TDR and ERT water deficit

The soil water content values were also derived from TDR measurements (Fig. 6) using the standard calibration given by IMKO for mineral soils. The averaged values of the three vertical profiles are found to be relatively stable between September and April, decreasing then during the hottest months (June to August), and finally quickly increasing when the first rains occur in autumn. TDR measurements are available at the same dates than ERT measurements (see dates in Fig. 6). Both TDR and ERT water content have been compared as follows: TDR water content were averaged from the three vertical profiles, while ERT water content values were averaged from 70 “soil columns” of width 25 cm; these values were compared at three depths: 0–20 cm, 20–40 cm, 40–70 cm. The comparison shows that ERT and TDR water content values globally exhibit the same temporal pattern, but differences up to of $0.05 \text{ m}^3/\text{m}^3$ in absolute can also be seen (Fig. 7). Estimates of the TDR-obtained water content values are lower than the ERT-obtained values in dry soils but higher than the ERT-obtained values in wet soils. This could be due to the fact that Archie's calibration could not be developed for low ERT water content, as said above. Furthermore, TDR measurements could be biased by the standard calibration given by Imko, as pointed out by Laurent et al. (2005). Additional measurements of soil water content were performed more recently, using different buried probes such as Delta ThetaProbe ML2 Deltacod or E+ soil MCT probe: the comparison of these measurements with the TRIME measurement clearly shows that TRIME measurements are underestimated for dry soils, and overestimated for wet soils. Thus, the bias in water content could be mainly due to the TDR TRIME measurement.

The water deficits were derived from the averaged values of the soil water content measurement in the three TDR tubes, at the three depths shown in Fig. 6. As TDR measurements were made only from a depth of 0 to 70 cm, the water deficits calculated by

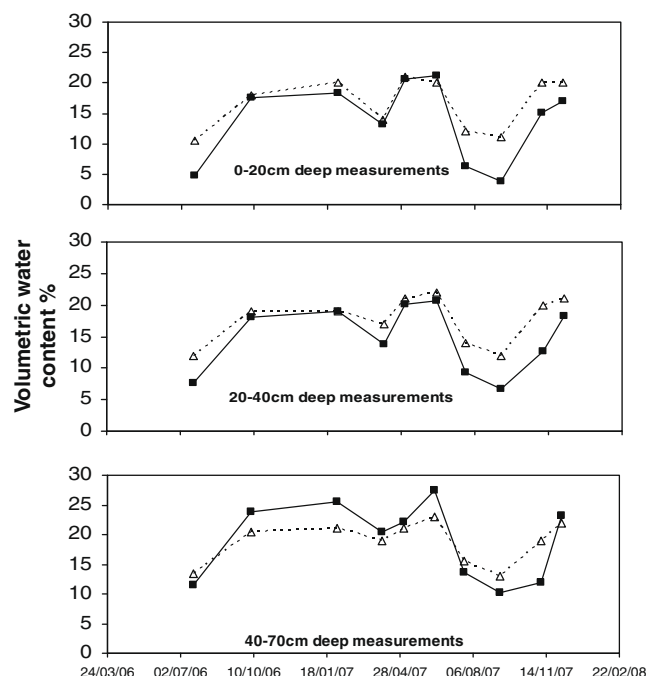


Fig. 7. Comparison of TDR water content (■ line) and ERT water content (△ dotted line) at three depths: 0–20 cm, 20–40 cm, 40–70 cm.

the two methods (TDR and ERT) were integrated between 0 and 70 cm, using the same porosity (0.42) on the vertical profile. This TDR averaged water deficit was compared to the ERT averaged water deficit relative at the depth 70 cm (while Table 2 stands for depth 120 cm). A good agreement was finally found between the two sets of data, i.e. the maximum difference did not exceed 15% (Fig. 8). However, the ERT water deficits exhibited smoother seasonal variations than the TDR water deficits: ERT values were about 10% lower than TDR values in wet conditions, and about 10% higher in dry conditions. ERT extreme values ranged from 140 to 200 mm, while TDR extreme values ranged from 120 to 230, i.e. nearly twice the ERT range. This could be due to physical measurements, but also to sampling effects, because TDR was averaged from three values while ERT was averaged from 70 values. This point requires further investigation, but nevertheless, the bias is small enough so that the use of ERT opens interesting perspec-

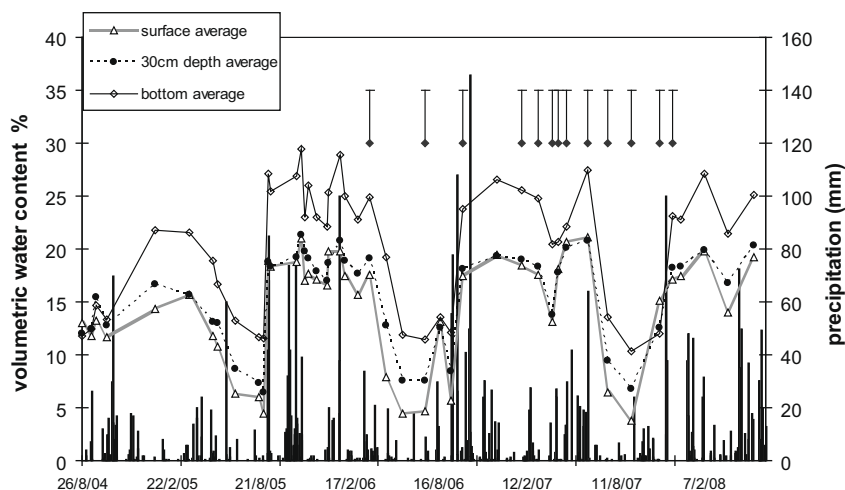


Fig. 6. Seasonal variations of the TDR soil water content since 2004. Averaged values from the three soil profiles surveyed. Arrows indicate the dates when ERT were measured.

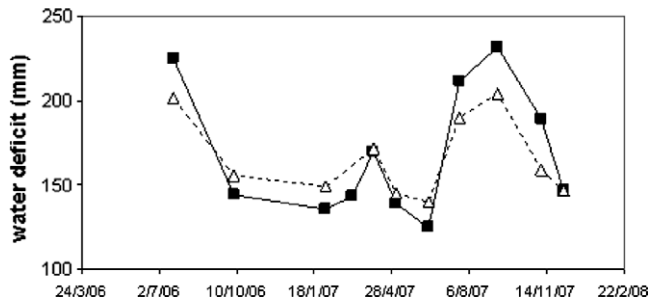


Fig. 8. Comparison of TDR water deficit (■ line) and ERT water deficit (△ dotted line) calculated for the top 70 cm of soil.

tives for the assessment of the soil water content and the water deficit.

Uncertainties in ERT soil water content and water deficit estimates

Several uncertainties must be taken into account for the interpretation of ERT in terms of water content or water deficit. These uncertainties mainly concern temperature, water solution resistivity, porosity, and the algorithm for the inversion of the apparent resistivities:

- a bias in the temperature has a direct effect on the assessment of resistivity at the standard temperature, which also affects both water content and water deficit:

$$\frac{\Delta \rho_{25}}{\rho_{25}} = \frac{\alpha T}{1 + \alpha(T - 25)} \cdot \frac{\Delta T}{T}$$

$$\frac{\Delta S}{S} = \frac{\Delta \theta}{\theta} = -\frac{1}{n} \cdot \frac{\Delta \rho_{25}}{\rho_{25}}$$

$$\frac{\Delta \delta}{\delta} = \frac{S}{1 - S} \cdot \frac{\Delta S}{S}$$

Thus, a 10% positive bias in temperature leads to a +4% bias in ρ_{25} , if $T = 10^\circ\text{C}$, and a 6.25% bias in ρ_{25} , if $T = 25^\circ\text{C}$. This bias will be –2.4% in both water content and water saturation, if $T = 10^\circ\text{C}$, and –3.7% if $T = 25^\circ\text{C}$.

Finally, a 10% positive bias in the temperature will affect the water deficit, depending on the temperature and the water saturation. The bias in the water deficit will be 0.9% if $T = 10^\circ\text{C}$ and $S = 0.2$, 9.6% if $T = 10^\circ\text{C}$ and $S = 0.8$. The highest bias occurs when the soils are very wet. It should be noted that in this case, because the water deficit is low, the absolute deviation is also low.

- A bias in pore water resistivity has a direct effect on the assessment of water content or water saturation of the soil, and also of the water deficit of the soil:

$$\frac{\Delta S}{S} = \frac{\Delta \theta}{\theta} = \frac{1}{n} \cdot \frac{\Delta \rho_w}{\rho_w}$$

$$\frac{\Delta \delta}{\delta} = \frac{S}{1 - S} \cdot \frac{\Delta S}{S}$$

A 10% positive bias in pore water resistivity thus leads to a +5.9% bias in both water content and water saturation in the soil. Concerning water content, the bias depends on soil water saturation: the bias equals +1.5% if $S = 0.2$, +23.6% if $S = 0.8$.

- A bias in porosity affects both water content and soil water saturation, but the effects are not the same, since soil water saturation depends on both water content and porosity. If $\theta = \left(\frac{\rho}{\rho_w} \cdot \Phi^{n-m}\right)^{1/n}$, a bias in porosity leads to a bias in water content:

$$\frac{\Delta \theta}{\theta} = \frac{n - m}{n} \cdot \frac{\Delta \Phi}{\Phi}$$

the bias in soil water saturation is given by:

$$\frac{\Delta S}{S} = \frac{\Delta \theta}{\theta} - \frac{\Delta \Phi}{\Phi}$$

and finally, the bias in the water deficit is given by:

$$\frac{\Delta \delta}{\delta} = \frac{1}{1 - S} \cdot \frac{\Delta \Phi}{\Phi} - \frac{S}{1 - S} \cdot \frac{\Delta \theta}{\theta}$$

Thus, a 10% positive bias in porosity leads to a +2.35% bias in water content, and to a 7.65% bias in soil water saturation. The corresponding bias in the water deficit is +11.4% when $S = 0.2$, and 40.6% when $S = 0.8$. A reliable assessment of porosity is thus a key factor for the assessment of the water deficit.

- The inversion algorithm leads to uncertainties that increase with depth (Marescot et al., 2003). These uncertainties are difficult to quantify, and sensitivity analysis should be performed using appropriate simulations, introducing modifications in the apparent resistivities of the upper cells of the ERT and measuring its impact on the deeper cells.

To summarize, it should be noted that the errors in the estimation of the parameters involved in Archie's relationship do not much influence the estimation of soil water content and water deficits. In most cases, a 10% bias in one of the variables in Archie's law does not generate errors greater than 5% in soil water content, soil water saturation and water deficit. The most unfavourable cases concern the estimation of the water deficit when the soil is very wet (but in this case, absolute errors are low) or when the uncertainty in the estimation of porosity is high.

Conclusions

The aim of the Electrical Resistivity Tomography performed on shaley soils in the Cevennes (southern France) was to assess the feasibility of such measures for a more systematic exploration of water content and water deficit in superficial soils.

The calibration of Archie's law was derived in the laboratory from two samples of undisturbed soil, and the set of parameters $m = 1.25$, $n = 1.65$ proved to be satisfactory. These values are little lower than the values found in the literature for this type of soil. The work in the laboratory also helped characterize changes in the resistivity of the water in the soil, which tends to a limit value of about $22 \Omega \text{ m}$, which was reached after little less than 100 h.

In situ soil water content and water deficit were calculated on the basis of Archie's law, which was previously calibrated in the laboratory. The comparison of the ERT results with the values measured by TDR along three vertical profiles led to differences of less than 15% on average. This result is relatively conclusive and offers interesting prospects. Measurements of electrical resistivities indeed have several advantages over measurements by TDR: non-intrusive measurement, integration of a first level of spatial variability at decametric or hectometric scale, the opportunity to probe greater depths and to characterize the structure of the whole soil profile.

However, the interpretation of electrical resistivity measurements remains complex, given the different factors on which they depend: porosity, temperature, pore water resistivity, water content. The more systematic use of ERT measurements depends on the control that can be exerted on these factors.

For example, the interpretation of electrical resistivity measurements in terms of water content, and especially in terms of water deficit, is quite sensitive to the porosity of the different horizons of

the vertical soil profile. The porosity measurements we made so far are relatively homogeneous, with values ranging from 0.5 for the humus horizon, and 0.45–0.4 for the underlying horizons. These values could be used for an initial approximation to define a typical profile of shaley soils in the Cevennes.

The other factors appear to have less influence on water content assessment, and they could be considered to be more reliable. Temperatures at different soil depths can be derived from the air temperatures near the surface. Pore water resistivity tends rapidly to a value of 22 Ω m, which could be considered as a first regional reference for this soil unit. The parameters m and n of Archie's law that were derived in the laboratory from the soil samples were stationary, and close to estimates reported in the literature.

This work thus opens interesting prospects for the use and exploitation of electrical resistivity measurements for the assessment of soil water content and water deficit, and for a more systematic exploration of this type of soil, which is widely distributed in the Cevennes. However, further testing is required to confirm the assumed homogeneity of the characteristics of this type of soil. The study should also be extended to other types of soils lying above magmatic or sedimentary materials, which are also frequent in the area.

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