Field, laboratory and estimated soil-water content limits

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Abstract

For the purpose of irrigation scheduling, estimates of soil-water content limits are determined using field or laboratory measurements or empirically-based regression equations. In this study the field method involved measuring simultaneously the soil-water content (using a frequency domain reflectometer with the PR1 profile probe that relies on changes in the dielectric constant of soil), and soil-water potential (using Watermark granular matrix sensors and tensiometers) at three depths (100, 300 and 600 mm) from a 1 m^2 bare plot. A retentivity relationship was developed from these measurements and the drained upper limit was estimated to be 0.355 $m^3 \cdot m^{-3}$ when the drainage from the pre-wetted surface was negligibly small. The lower limit, corresponding to -1 500 kPa, was estimated to be 0.316 $m^3 \cdot m^{-3}$. In the laboratory, soil-water content and soil matric potential were measured on undisturbed soil samples taken from the edge of the bare plot. The undisturbed soil samples were saturated and exposed to different matric potentials between -1 and -1 500 kPa. A retentivity relationship was then developed from these measurements. The laboratory method realized a drained upper limit value of 0.390 $m^3 \cdot m^{-3}$ at -33 kPa and a lower limit value of 0.312 $m^3 \cdot m^{-3}$ at -1 500 kPa. A regression equation, which uses the soil bulk density and the clay (<0.002 mm) and silt (0.002 to 0.05 mm) percentage to estimate the soil-water content at a given soil-water potential, realised a drained upper limit value of 0.295 $m^3 \cdot m^{-3}$ at -33 kPa and a lower limit value 0.210 $m^3 \cdot m^{-3}$ at -1 500 kPa. Comparisons were made between field, laboratory and regression equation methods of estimating the upper and lower soil-water content limits. The field-measured soil-water content was statistically different from the laboratory-estimated and regression equation estimates of soil-water content. This was shown from a paired *t*-test, where the probability levels for the laboratory and regression equation methods were 0.011 and 0.0005 respectively at the 95 % level of significance. Field method soil-water content comparisons with the laboratory method resulted in a linear regression coefficient of determination of 0.975 with a root mean square error (RMSE) of 0.064 $m^3 \cdot m^{-3}$. By contrast, field method comparisons with the regression equation method showed a coefficient of determination of 0.995 with an RMSE of 0.035 $m^3 \cdot m^{-3}$. The frequency domain reflectometry method used for monitoring soil-water content has been shown to be useful in this case of relatively homogenous soils supporting perennial crops.

Keywords: soil-water content limits, Watermark granular matrix sensor, tensiometer measurements, PR1 soil-water profile probe dielectric method

Introduction

The field-measurement of soil-water content is of paramount importance in irrigation science and irrigation management. It affects irrigation system design, irrigation system management, crop selection and crop management. There are, however, very few practical field methods for measuring or estimating soil-water content. Accurate measurement of the lower limit and the drained upper limit is required to estimate the available water reserve of a soil and these limits are critical inputs required by soil-water balance models (Ritchie, 1981). Both the lower limit and drained upper limits can be measured in the field or laboratory or they can be estimated using empirical equations based on easily measured soil properties such as soil texture, soil bulk density and soil organic matter content. The field-measured lower limit is taken as the soil-water content at which plants were practically dead or dormant as a result of the soil-water deficit (Ratliff et al., 1983). The lower limit could also be measured using *in situ* soil psychrometers (Savage et al., 1996). The drained upper limit is taken as the soil-water content at which drainage from a pre-wetted soil practically ceases or when the soil-water content decrease is about 0.001 to 0.002 $m^3 \cdot m^{-3} \cdot d^{-1}$ (Ratliff et al., 1983).

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In the laboratory, the most common procedure for estimating the drained upper limit and lower limit is to extract water from a disturbed or undisturbed soil sample using the soil-water extraction apparatus (Richards and Weaver, 1943). The lower limit is estimated using the apparatus at a soil matric potential of -1 500 kPa (Richards and Weaver, 1943). The water content at a matric potential of -33 kPa is used as an estimate of the drained upper limit for moderately coarse and fine-textured soils, whereas -10 kPa is used for coarse-textured soils (Colman, 1947; Jamison and Kroth, 1958).

Field or laboratory measurement of the relationship between soil-water potential and soil-water content is expensive, difficult, and often impractical (Saxton et al., 1986). Thus, for many purposes, general estimation is often based on more readily available information such as soil texture, soil bulk density and soil organic matter, thereby reducing the time and cost of laboratory and field measurements. Many researchers (Brooks and Corey, 1964; Gupta and Larsen, 1979; Mottram et al., 1981; Rawls and Brakensiek, 1982; Cosby et al., 1984; Schulze et al., 1985; Hutson, 1986; Saxton et al., 1986; Ritchie et al., 1999) have developed regression equations to estimate the soil-water potential and water content relationships from soil texture, soil bulk density and soil organic matter. Mottram et al. (1981), for example, developed regression equations for the top- and sub-soil of 31 soil types at Mkuzi (KwaZulu-Natal, South Africa) based on the soil texture clay (< 0.002 mm) and silt (0.002 to 0.05 mm), organic matter and bulk density. The lower limit was estimated at a matric potential of -1 500 kPa and the upper limit of plant available water was defined for a matric potential of -5 kPa as opposed to the normally accepted value of -10 or -33 kPa. This choice supports the findings of MacLean and Yager (1972) in Zambia.

Ratliff et al. (1983) compared field and laboratory measurements of the lower limit and the drained upper limit and found that laboratory estimates of the drained upper limit obtained at a water content corresponding to -33 kPa were significantly less than field-measured drained upper limits for sands, sandy loams, and sandy clay loams and significantly greater than field estimates for silt loams, silty clay loams, and silty clays. Laboratory estimates of the lower limit corresponding to -1 500 kPa were significantly less than field lower limit measurements for sands, silt loams and sandy clay loams and significantly more than field observations for loams, silty clays, and clays. Ratliff et al. (1983) also suggested that, if accuracy is necessary in soil-water balance calculations, laboratory-estimated soil-water limits should be used with caution and also that field-measured limits are preferred.

Salter and Haworth (1961) also found that the direct method in the field involving soil sampling after irrigation and drainage had almost ceased, gave more accurate and consistent measurements using the soil-water extraction laboratory method. From their results they concluded that for rough estimation of soil-water content limits, the laboratory method using undisturbed soil cores yields satisfactory results, but for more critical work, the use of the direct measurement (field method) is essential using soil-water content or soil-water potential sensors (Lukangu et al., 1999; Gebregiorgis and Savage, 2006; Lukanu and Savage, 2006). An instrument and/or field method for estimating soil-water content or soil-water potential that is practical and sufficiently accurate will be of great benefit.

In this study, three methods for determining the lower and drained upper soil-water content limits were tested and the measurements compared. Field, laboratory and estimated values of soil-water potential and soil-water content values were measured to determine the soil-water content limits.

Materials and methods

In the field, soil-water content was measured using a PR1 profile probe (Delta-T Devices, Cambridge, UK) which provides measurements at 100, 200, 300, 400, 600 and 1 000 mm depths and soil-water potential was measured using tensiometers (Cassel and Klute, 1986) and Watermark granular matrix sensors (Irrometer, Riverside, USA).

The electronics embedded in the PR1 profile probe (dielectric method that uses frequency domain reflectometry), generates a 100 MHz signal. If the dielectric properties of the soil are different from the probe electronics, the reflected signal combines with the generated signal to form a standing wave with an amplitude that is a measure of the soil-water content. The PR1 profile probe contact between soil and sensor is crucial for accurate soil-water measurements and particularly so for soil-water content measurements (Gebregiorgis and Savage, 2006). Currently, there is no measure of whether there is adequate soil contact in any of the soil-water dielectric methods used.

The Watermark sensor is designed to offer the advantage of the tensiometric and gypsum block approaches (Armstrong et al., 1987). The Watermark sensor electrodes are embedded in a non-dissolvable matrix material, and the sensor incorporates an internal gypsum buffer to minimise the effect of salts experienced in irrigated landscapes. The matrix material is held in place by a porous membrane. Watermark sensors are sensitive to salinity and soil temperature and the matrix characteristics of the sensor changes with time (Jovanovic and Annandale, 1997). In the laboratory, the soil-water content and corresponding soil-water potential value were measured at the same time to develop a retentivity relationship. A known pressure was applied using a tension table creating a matric potential between -1 and -10 kPa, pressure pot (-50 and -100 kPa) and pressure chamber at -1 500 kPa and the soil-water content was measured after equilibrium was reached. The empirical equations developed by Hutson (1986) were used to estimate the soil-water content limits. These equations require the clay, silt, fine sand percentages and soil bulk density as an input to estimate the soil-water content at the corresponding matric potential.

For each method, the soil-water content and soil-water potential were related using the retentivity relationship developed by Gardner et al. (1970). In this relationship, the soil-water potential was treated as the independent variable and the soil-water content the dependent variable. The retentivity relationship is expressed by:

$$\theta = \left(-\psi/a\right)^{-1/b} \tag{1}$$

where:

θ is the volumetric soil-water content ($m^3 \cdot m^{-3}$) and

Ψ is the soil-water potential (kPa) [where the negative sign in front of Ψ ensures that the exponent may be calculated].

The a and b empirical constants, which can be developed from the regression line of lnθ vs. ln ($-\Psi$), are given by:

$$a = exp\ (a_r\ b) \tag{2}$$

$$b = -1/b_r \tag{3}$$

where:

a_r and b_r are the intercept and slope values of the regression line respectively for the ln θ *vs* ln ($-\Psi$) graph fitted by a straight line.

Field measurements

In the field, inside a 1 m^2 bare plot (Photo 1), six tensiometers and six Watermark sensors were installed at 100, 300 and 600 mm soil depths in two replications with all sensors equidistant from a single PR1 profile probe. The depths were chosen to represent the root zones within the cultivated soil and immediately below the depth of cultivation. The tensiometers and the Watermark sensors were installed around the PR1 profile probe within a radius of 200 mm and 150 mm apart from each other.

The access tubes of the PR1 profile probe were installed using gouge and spiral augers, taking care to not disturb the soil profile. First, the gouge auger (22 mm diameter) was pushed into the soil to the depth of the blade. The auger was then fully rotated to excavate the soil, and withdrawn while continuing to rotate. When the desired depth was reached, the hole was shaped with the spiral auger (25 mm diameter). The access tube was then pushed into the slightly narrower hole to ensure a tight fit that prevented the creation of air gaps between the access tube and the soil.

The Watermark sensors were soaked in water for two days prior to their installation. In the field, 25 mm diameter holes were made using the gouge and spiral augers, to the desired depth, and filled with a slurry made from the excavated soil. These procedures ensured a snug fit between sensor and soil.

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Photo 1

A PR1 profile probe is shown at the centre of a 1 m^2 bare plot. Tensiometers and Watermark sensors surround the PR1 profile probe. A Pronamic rain-gauge is shown at the bottom right

TABLE 1
Regression equations (Eqs. (4) to (10)) developed by Hutson (1986) to estimate the soil-water content at matric potentials of -1, -3, -10, -30, -100, -500 and -1 500 kPa respectively

$\theta_{-1} = 0.686 + 0.000794\,(Cl + Si) - 0.229\rho_b$	4
$\theta_{-3} = 0.349 + 0.00211\,(Cl + Si) - 0.096\rho_b$	5
$\theta_{-10} = 0.112 + 0.00380\,(Cl + Si)$	6
$\theta_{-30} = 0.065 + 0.00396\,(Cl + Si)$	7
$\theta_{-100} = 0.038 + 0.00372\,(Cl + Si)$	8
$\theta_{-500} = 0.0185 + 0.00366\,(Cl + Si)$	9
$\theta_{-1500} = 0.0187 + 0.00337\,(Cl + Si)$	10

θ is the volumetric water content in $m^3{\cdot}m^{-3}$, (Cl + Si) is the sum of clay and silt content of the soil in percentage and ρ_b is the bulk density of the soil in $Mg{\cdot}m^{-3}$.

The sensors were then pushed with the PVC pipe, which was fitted tightly over the sensor collar until it reached the required depth. The holes were then carefully backfilled and trampled down to prevent air pockets, which could allow water to be channelled to the sensor depth.

The ceramic tips of the tensiometers were saturated for 24 h in water before installation. In the field, 22 mm diameter holes were made using the gouge auger and a slurry was poured into the holes. The slurry was made from the excavated soil and ensured a tight fit between sensor and soil. The tensiometers were pushed carefully so that they reached the desired depth. The holes were then backfilled and trampled to prevent channelling of water to the sensor depth.

After all the sensors had been installed, the plot was flooded and covered for two days with black plastic to prevent evaporation and to allow a redistribution of water throughout the soil profile. The soil-water potential Watermark and tensiometer sensors were connected to a CR23X data logger (Campbell Scientific, Logan, USA) and the PR1 profile probe connected to a CR10X data logger programmed to measure the soil-water content every 3 h. The soil-water content vs. soil-water potential relationship was determined from the simultaneous measurement of the PR1 profile probe soil-water content and the soil-water potential sensors (tensiometer and Watermark) while the soil dried. The drained upper limit value was determined after two days when the rate of change in the soil-water content was negligible (Ratliff et al., 1983). The lower limit was also calculated, using the retentivity relationship, as corresponding to a soil-water potential of -1 500 kPa, which corresponds closely to the field lower limit of soil-water availability (Savage et al., 1996).

Laboratory measurements

To determine the retentivity relationship, six undisturbed soil samples were taken 800 mm away from the position of the sensors, using a core sampler, at the 100, 200, 300, 400, 600, and 1 000 mm soil depths in three replications at the edge of the 1 m^2 bare plot prior to the installation of the sensors. This choice of distance between sensor position and sampling site ensured that the disturbance created by sensor installation did not impose on the sphere of influence of the installed sensors. The sphere of influence of the sensors is less than 100 mm and 100 mm for the profile probe (Delta-T Devices, 2001). The tensiometer (Cassel and Klute, 1986) and Watermark (Armstrong et al., 1987) sensors equilibrate with the soil-water potential of the neighbouring soil.

After the core samples were taken, they were trimmed carefully to the edge of the sleeve and saturated in a water bath by capillary action. After the cores were totally saturated, they were weighed while water was dripping to obtain the saturation weight and transferred to a tension table where a hanging water column was used to create a matric potential between -1 and -10 kPa, a pressure pot was used for matric potentials between -50 and -100 kPa and a pressure chamber for a matric potential at -1 500 kPa. In each method, the pressure was changed after the cores attained equilibrium and weighed before subjecting them to the next matric potential. The time to equilibrate varied from 2 d at the higher tension (-1 to -10 kPa) to 10 d at the lower tension (-1 500 kPa). Finally, the cores were oven-dried at 105°C for 4 d and reweighed to determine the water content on dry mass basis. Soil bulk density was also determined to convert the gravimetric soil-water content ($kg{\cdot}kg^{-1}$) to volumetric water content ($m^3{\cdot}m^{-3}$).

Estimated values of soil-water content limits

The regression equations developed by Hutson (1986) were used to estimate the soil-water content at -1, -3, -10, -30, -100, -500 and -1 500 kPa (Eqs. (4) to (10) of Table 1).

These equations were developed based on 409 South African soil samples. To estimate the soil-water content at the corresponding matric potential the percentage of clay, silt, fine sand, and soil bulk density in $Mg{\cdot}m^{-3}$ was determined. These equations use the particle size classification of the South African Soil Classification System (Soil Classification Working Group, 1991). According to this classification, the average values between 100 and 300 mm soil depths are clay 40%; silt 17%; fine sand 43% and soil bulk density 1.354 $Mg{\cdot}m^{-3}$.

The drained upper limit was calculated using the retentivity relationship:

$$\theta = (-\Psi / 138 \times 10^{-6}\ \text{kPa})^{-0.0987} \qquad (11)$$

The constants used for the retentivity relationship are a = 138 x 10^{-6} kPa and b = -10.13 (Eq. (1)).

A matric potential of -33 kPa was used in Eq. (11) (Colman, 1947; Jamison and Kroth, 1958) and the lower limit was calculated using the regression equation at a soil matric potential of -1 500 kPa (Richards and Weaver, 1943). The plant-available

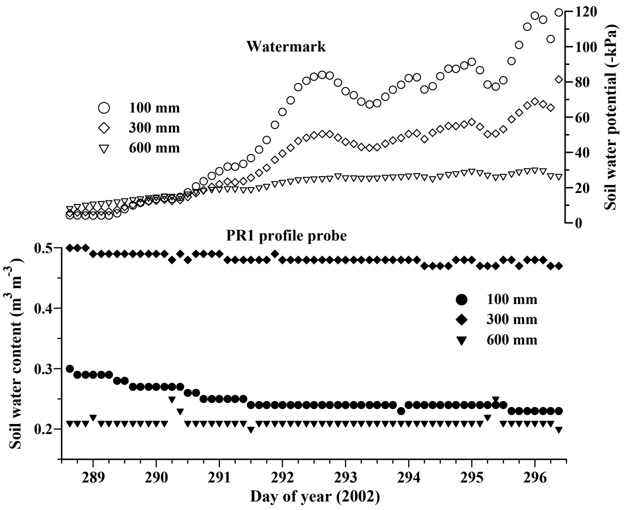

Figure 1
Field-measured soil-water content (m³·m⁻³) using a PR1 profile probe (lowest set of three curves) and soil-water potential (kPa) measured using Watermark sensors at three depths (upper set of three curves)

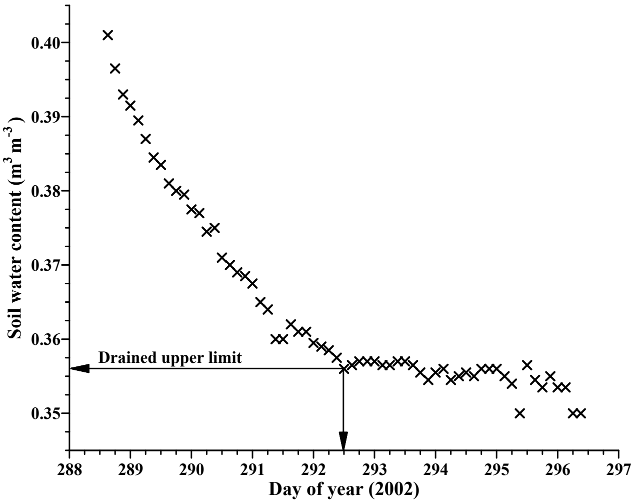

Figure 2
PR1 profile probe soil-water content vs. day of year after flooding the plot to determine the drained upper limit of the soil between the 100 and 300 mm soil depths

water (PAW) was then calculated from the difference between the drained upper and lower limits.

Results and discussion

Field measurements

The field-measured PR1 profile probe soil-water content and Watermark soil-water potential at 100, 300 and 600 mm soil depths for the 1 m^2 plot are shown in Fig. 1. The temporal soil-water content variation was between 0.30 and 0.23 m^3·m^{-3} at the first 100 mm soil depth with a corresponding soil-water potential of -4 to -119 kPa. At this depth, soil-water content generally decreased due to evaporation and redistribution. Even though there were cover crops (oats, rye and rye grass) around the plot, there were not many plant roots to extract water from this depth. At the 300 mm depth, soil-water content varied between 0.50 and 0.47 m^3·m^{-3}, corresponding to a soil-water potential between -5 and -81 kPa. This small change in soil-water content could be due to the high clay content of the soil but it could also be due to the fact that there were no plant roots to extract the soil water. In clay soils, since the pore-size distribution is more uniform, more of the water is adsorbed, so that increasing the matric potential causes a more gradual decrease in soil-water content (Hillel, 1971). The soil at the 600 mm soil depth has a low soil-water content compared to that at the shallower depths since the applied water did not reach this depth. The soil-water content was almost constant at around 0.21 m^3·m^{-3} at the soil potential of -8 to -30 kPa.

The field-measured soil-water content and soil-water potential values were compared with the results obtained by Schmidt and Schulze (1989) for the Cedara catchments. They calculated the PAW in the laboratory from the difference of the soil-water content at -33 and -1 500 kPa matric potentials. They obtained different ranges of soil-water content at -33 and -1 500 kPa. The lowest soil-water content varied between 0.26 and 0.23 m^3·m^{-3} at -33 and -1 500 kPa matric potentials respectively. The largest soil-water content ranged between 0.43 and 0.24 m^3·m^{-3} at -33 and -1 500 kPa matric potentials respectively. Considering that these measurements were made in the laboratory at a wider range of soil-water potential (-33 to -1 500 kPa), the field-measured soil-water content values obtained using the PR1 profile probe and soil-water potential measurements obtained using the Watermark sensors were reasonable when compared with the results of Schmidt and Schulze (1989).

TABLE 2
Retentivity function calculated soil-water content at certain points of soil-water potential

Soil-water potential (-kPa)	Field-measured θ (m³·m⁻³)
1	0.415
5	0.391
10	0.381
33	0.365
50	0.359
100	0.350
500	0.329
1 000	0.321
1 500	0.316

The values of the PR1 profile probe soil-water content and Watermark soil-water potential were averaged for the 100 and 300mm soil depths to determine the drained upper limit and extrapolated to determine the lower limit since the soil-water potential did not reach -1 500 kPa in the field. The drained upper limit was 0.355 m^3·m^{-3} for the 300 mm soil depth. This value was taken as the soil-water content when the decrease in soil-water content at this depth was negligible (Fig. 2). The lower limit was inferred as corresponding to a matric potential of -1 500 kPa (Table 2) using the retentivity relationship:

$$\theta = (-\Psi / 54 \times 10^{-12}\ \text{kPa})^{-0.0372} \quad (12)$$

The constants for the retentivity relationship a = 54 x 10^{-12} kPa and b = -26.88 (Eq. (1)) were calculated from the graph of ln θ *vs.* ln (-Ψ).

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TABLE 3
Comparison of field-measured (*X*) soil-water content ($m^3 \cdot m^{-3}$) against the laboratory (Y_1) and estimated (Y_2) soil-water content ($m^3 \cdot m^{-3}$)

Statistical parameters	Laboratory-measured (Y_1)	Estimated values (Y_2)
N	9	9
r^2	0.975	0.995
RMSE	0.064	0.035
P (T≤t) two-tail (95%)	0.011	0.0005
Slope	1.563	2.118
Intercept	-0.179	-0.472
Standard error of Y on X	0.009	0.006
SE slope	0.094	0.059
Slope confidence limit 99%	1.234, 1.892	1.912, 2.324
Slope confidence limit 95%	1.341, 1.785	1.978, 2.257
SE intercept	0.060	0.037
Intercept confidence limit 99%	-0.2979, -0.061	-0.603, -0.341
Intercept confidence limit 95%	-0.2595, -0.099	-0.561, -0.384

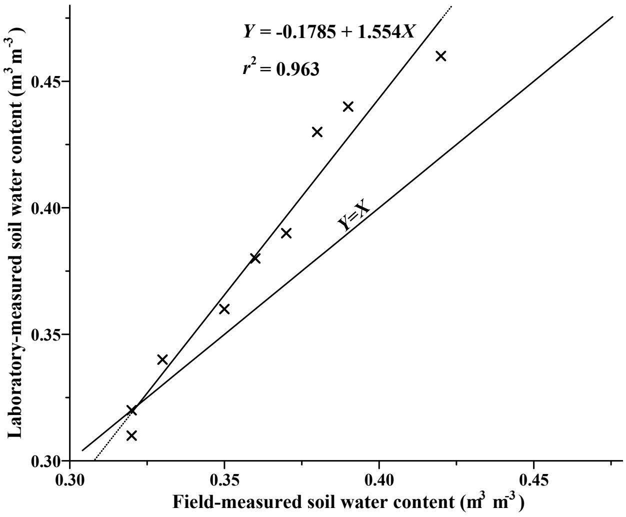

Figure 3
Laboratory-measured soil-water content ($m^3 \cdot m^{-3}$) vs. field-measured soil-water content ($m^3 \cdot m^{-3}$) at the same soil-water potential

The PAW was then equal to 0.039 $m^3 \cdot m^{-3}$ or 3.9%. Schmidt and Schulze (1989) calculated the PAW for the Cedara catchments to be between 2.67 and 19.8%.

The drained upper limit (0.355 $m^3 \cdot m^{-3}$) determined in the field (Fig. 2), when the soil-water content decrease was negligible, was close enough to the drained upper limit (0.365 $m^3 \cdot m^{-3}$) estimated using the retentivity relationship at a soil-water potential of -33 kPa. This result agreed with the estimation of the drained upper limit at a soil-water potential of -33 kPa, which was proposed by Colman (1947) and Jamison and Kroth (1958). Other workers have also proposed different drained upper matric potentials with satisfactory results. For example, Hanks et al. (1954) used -20 kPa, Haise et al. (1955) used -10 kPa and Russel and Balcerek (1944), MacLean and Yager (1972) and Mottram et al. (1981) used -5 kPa to estimate the drained upper limit values. These variations in matric potential depend on soil texture. For example, sandy soils reach the drained upper limit at -6 kPa, loamy sand at -10 kPa, silt loams at -30 kPa and clay soils at -60 kPa (Water Resource Publications, 1964).

Laboratory measurements

The average soil-water content (θ) and soil-water potentials (Ψ) at the rooting depth (100 to 300 mm) were used to estimate the drained upper limit and lower limit values of the soil. From the laboratory measurements, the drained upper limit was 0.39 $m^3 \cdot m^{-3}$ at -33 kPa and the lower limit was 0.31 $m^3 \cdot m^{-3}$ at -1 500 kPa. The PAW was then calculated to be 0.08 $m^3 \cdot m^{-3}$ (or 8%).

The statistical analysis (Table 3) showed that the laboratory measurement of soil-water content was statistically different from the corresponding field-measured soil-water content at a given soil-water potential. From the result of the paired *t*-test it was found that the probability level (P = 0.011) was lower than the critical alpha value (α = 0.05), which indicated that there were significant differences between the two means at the 95% level of significance. The slope and intercept of the regression line were also statistically different from one and zero respectively (Table 3), which demonstrated that the soil-water content measurement in the laboratory was not a perfect estimation of the field measurement. The laboratory-measured soil-water content showed a bias (Fig. 3) with a systematic error of 94.6%.

When the laboratory-estimated drained upper limit values were compared with the field-measured drained upper limit values, the laboratory measurement over-estimated the drained upper limit value by 0.045 $m^3 \cdot m^{-3}$. This result agreed with the conclusion made by Ratliff et al. (1983) that the laboratory estimates of the drained upper limit value obtained at -33 kPa water contents were significantly more than the field-measured drained upper limit. However, the laboratory estimate of the lower limit (0.312 $m^3 \cdot m^{-3}$) obtained at -1 500 kPa matric water potential was almost equal to the field-measured lower limit (0.316 $m^3 \cdot m^{-3}$). This result agreed with the experimental result of Savage et al. (1996). They found that the choice of the -1 500 kPa soil-water potential was appropriate and corresponded closely to the field lower limit of soil-water availability.

The retentivity relationship for the laboratory measurement (Eq. (13)) was developed to estimate the soil-water content for a given soil-water potential:

$$\theta = (-\Psi / 3.94 \times 10^{-6}\ \text{kPa})^{-0.0588} \qquad (13)$$

The constants for the retentivity relationship $a = 3.94 \times 10^{-6}$ kPa and $b = -17.01$ were calculated from the slope and intercept of the graph ln θ *vs* ln (-Ψ) (Fig. 4) using Eqs. (2) and (3).

Estimated values of soil-water content limits

The soil-water content was estimated for the respective soil-water potentials based on the Hutson (1986) regression equations and a comparison was made between the soil-water content limits of field-measured and laboratory-measured values. Using the regression equations, the drained upper limit was 0.295 $m^3 \cdot m^{-3}$ at -33 kPa and the lower limit was 0.210 $m^3 \cdot m^{-3}$ at -1 500 kPa. The PAW was therefore 0.085 $m^3 \cdot m^{-3}$. The drained upper limit was under-estimated by 0.06 $m^3 \cdot m^{-3}$ and 0.095 $m^3 \cdot m^{-3}$ from the field-measured and laboratory-measured values respectively. The lower limit was also under-estimated by 0.11 $m^3 \cdot m^{-3}$ and 0.099 $m^3 \cdot m^{-3}$ from the field and laboratory measurements respectively.

From the statistical analysis (Table 3) the estimated soil-water content measurements were statistically different (P<α) and biased (systematic error = 97 %) from the corresponding

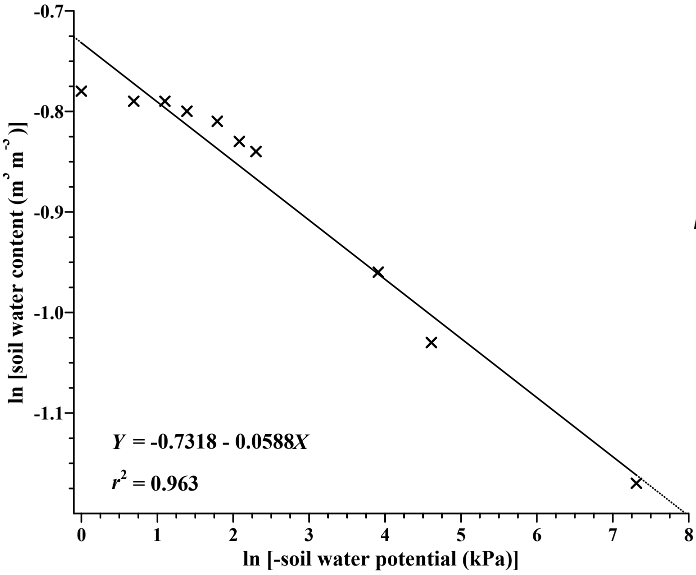

Figure 4
Field-measured soil-water content (ln θ) vs. field-measured soil-water potential (ln (-Ψ))

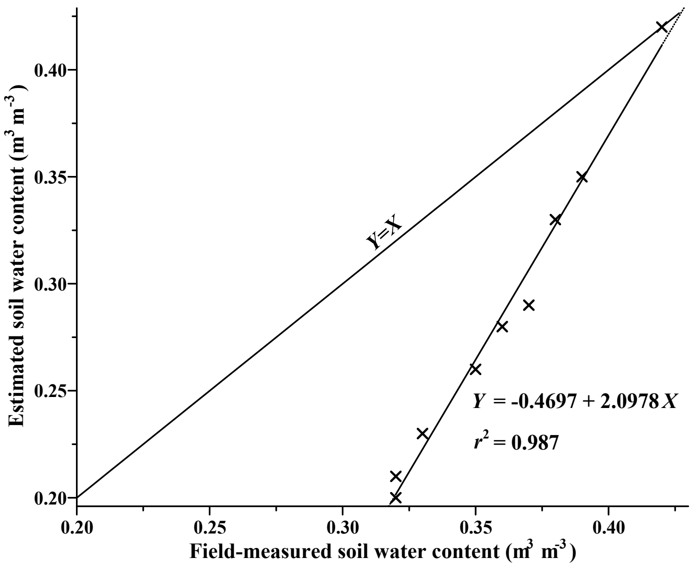

Figure 5
Estimated soil-water content vs. field-measured soil-water content at the same soil-water potential

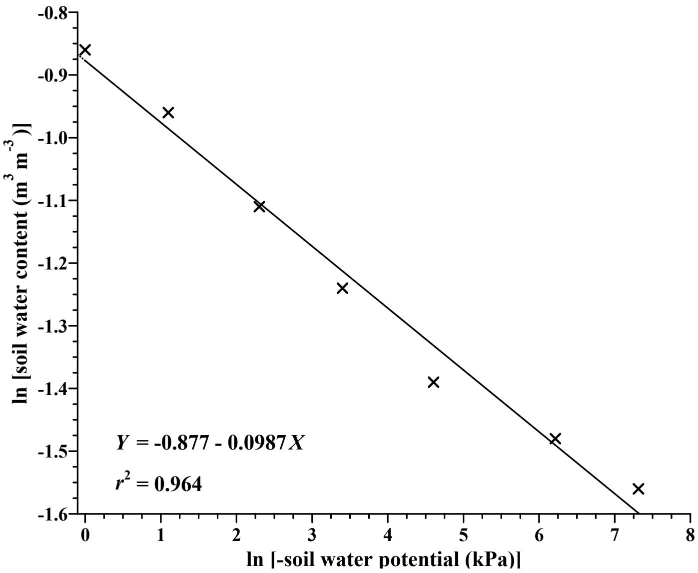

Figure 6
Estimated soil-water content (ln θ) vs. the known soil-water potential (ln (-Ψ)) fitted to develop the linear regression

field-measured soil-water content at a given soil-water potential (Fig. 5). The slope and intercept were also statistically different from one and zero respectively. However, the relationship is highly significant (linear regression coefficient of determination of 0.995) and the slope and intercept were used as the multiplier and offset to adjust the equation to adequately estimate the soil-water content at the corresponding soil-water potentials.

The retentivity relationship for the estimated values (Eq. (11)) was developed to estimate values of soil-water content at a given soil-water potential. The constants a and b (Eq. (1)) were calculated using the graph ln θ *vs* ln (-Ψ) (Fig. 6).

Conclusions

The results obtained using laboratory and estimated soil-water content values demonstrate statistical differences from the soil-water content measured in the field. The variation in soil-water content was mainly due to the difference between the measurement methods, but in part the difference is also due to soil variability and the treatment of the soil sample between the time when the samples were taken from the field and laboratory measurement. With great care, laboratory measurements could yield a good estimation of soil-water content limits, if the errors encountered in the field and in the laboratory were minimized. Practically, it is relatively simple and feasible with the available sensors to measure the drained upper limit in the field. However, it is much more difficult to measure the lower limit in the field with the readily available sensors since the soil-water potential limit of -1 500 kPa may not be reached. In such situations, the best option is to measure the lower limit in laboratory. The use of regression equations, which allow soil-water content estimates using some easily measurable soil parameters, could be useful for estimating the soil-water content limits when the time, cost and labour needed to undertake the field and laboratory measurements are considered. The regression equations that were developed by Hutson (1986) showed a linear regression coefficient of determination of 0.995 with systematic error of 97%. If the

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equation was calibrated against the gravimetric soil-water content at the corresponding soil-water potential, it could yield a good estimate of soil-water content.

In this study, the drained upper limit and lower limit were defined using the laboratory-measured values of soil-water content at -33 and -1 500 kPa. However, many workers do not recommend the laboratory method, if direct measurement in the field is possible. The laboratory-measured values were taken, since the soil-water content was measured within the matric potential range from saturation to -1 500 kPa. In the field, measurements were made from -4 to -119 kPa at a soil depth of 100 mm, -5 to -81 kPa at 300 mm and -8 to -30 kPa at 600 mm. The drained upper limit (0.39 $m^3 \cdot m^{-3}$) and the lower limit (0.31 $m^3 \cdot m^{-3}$), which was equal to 0.08 $m^3 \cdot m^{-3}$ resulted in the PAW value of 0.08 $m^3 \cdot m^{-3}$. The upper and lower limits could then be used for monitoring the soil-water content using a soil-water content profile probe to determine the timing and amount of irrigation.

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