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Soil physical quality
Part I. Theory, effects of soil texture, density, and organic matter, and effects on root growth

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Abstract

A soil physical parameter, *S*, is defined. It is equal to the slope of the soil water retention curve at its inflection point. This curve must be plotted as the logarithm (to base e) of the water potential against the gravimetric water content (kg kg^{-1}). The value of *S* is indicative of the extent to which the soil porosity is concentrated into a narrow range of pore sizes. In most soils, larger values of *S* are consistent with the presence of a better-defined microstructure. Much previous work has shown that this microstructure is responsible for most of the soil physical properties that are necessary for the proper functioning of soil in agriculture and the environment. The use of *S* is illustrated with examples of soils with different texture, density (or degree of compaction), and organic matter (OM) content. The effects of *S* on root growth in soil are investigated, and *S* is shown to be a better indicator of soil rootability than bulk density. It is suggested that *S* can be used as an index of soil physical quality that enables different soils and the effects of different management treatments and conditions to be compared directly.

Keywords: Compaction; Density; Organic matter; Root growth; Texture; Water retention curve

1. Introduction

Soil quality is a subject that is receiving increasing attention (e.g., Wilson and Maliszewska-Kordybach, 2000). Soil quality is usually considered to have three main aspects: physical, chemical, and biological. It is considered to be important for the assessment of the extent of land degradation or amelioration, and for identifying management practices for sustainable land use. Although this paper deals specifically with soil physical quality, it should be realised that this has big effects on chemical and biological process in the soil, and, therefore, it plays a central role in studies of soil quality.

Soil physical quality is manifest in various ways. Examples of poor physical quality are when soils exhibit one or more of the following symptoms: poor water infiltration, run-off of water from the surface, hard-setting, poor aeration, poor rootability, and poor workability. Good soil physical quality occurs when soils exhibit the opposite or the absence of the conditions listed above. Often, a soil exhibits several

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or all of these physical problems simultaneously. The reason is that all of these symptoms have a common cause—poor soil structure.

Several authors have commented that there has been no single measure of soil physical quality (e.g., Dexter and Czyz, 2000), and that it has been necessary to somehow integrate observations of a range of properties to obtain an overall assessment. For example, the "agrophysical index" is a weighted average of the values of 10 soil physical properties (Canarache, 1990).

This series of papers proposes the use of an index of soil physical quality, S. It is intended to be easily and unambiguously measurable using standard laboratory equipment. Its use is illustrated with examples showing different soils, different contents of soil organic matter (OM), different degrees of compaction, levels of sodicity, effects of tillage at different soil water contents, hard-setting behaviour, and predictions of hydraulic conductivity.

Its use is illustrated with experimental data from Australia, England, Poland, Spain, Sweden, the Netherlands, and the USA.

The wide applicability of the index of soil physical quality, S, is illustrated with real measurements made on real soils and also with predictions obtained from the use of pedo-transfer functions.

2. Theory

An example of a water retention curve of soil is shown in Fig. 1A. If a soil sample is drained progressively from saturation, the largest pores empty first, and this is followed by the emptying of progressively smaller pores. At any given water potential, ψ, the water menisci are in pores with an effective cylinder radius, a, given by

$$a = -\frac{2\sigma}{\psi}, \tag{1}$$

where σ is the surface tension of the water menisci. In cracks (planar pores), the menisci are in pores of width, t, given by

$$t = -\frac{\sigma}{\psi}. \tag{2}$$

In the remainder of this paper, water suctions, h, are used such that $h = -\psi$. Values of h have the

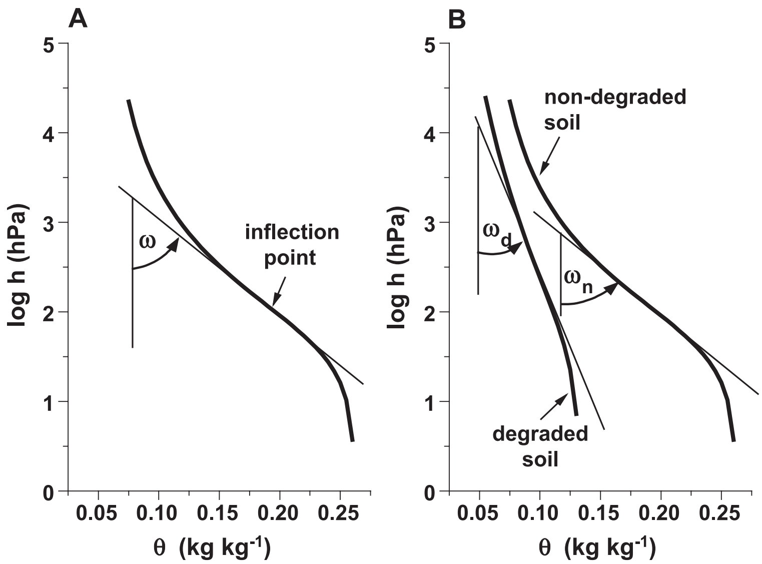

Fig. 1. (A) Example of a soil water retention curve showing the inflection point and the slope, tan ω, of the tangent to the curve at the inflection point. (B) Water retention curves of the same sandy clay loam soil at two different bulk densities. Soil physical degradation occurs when the soil is compacted, and this reduces the slope of the retention curve at the inflection point.

mathematical advantage that they are positive in unsaturated soil. In Fig. 1A, the curve of water content against log h has only one characteristic point, which is the inflection point. At the inflection point, the curvature is zero. The curve at the inflection point has only two characteristics: firstly, its position (θ_i, ln h_i), and secondly its slope, $S = \mathrm{d}\theta/\mathrm{d}(\ln h)$.

It may be noted that $S = C_i h_i$ where C_i is the differential (or specific) water capacity which is widely used in the soil water literature (e.g., Hillel, 1980). The subscripts i refer to the values at the inflection point. The significance of the inflection point was discussed by Dexter and Bird (2001) who identified θ_i with the optimum water content for tillage. This latter subject is explored further in Part II of this series.

In the mathematics and in the calculations presented here, the slope is calculated as $S = \mathrm{d}\theta/\mathrm{d}(\ln h)$. However, for easier comprehension, the graphs are plotted as the logarithm to the base 10 of h against θ, which gives a slope of $T = \mathrm{d}\theta/\mathrm{d}(\log h)$. It should be noted that values of T are larger than the corresponding values of S by a factor of $\log_e 10 = 2.3026$. Therefore, values of slope measured from graphs may need to be corrected before being compared with values obtained by calculation using the equations presented here. Although S is always negative, it is often more convenient to use the modulus of S in discussions.

Of principal interest is the slope of the water retention curve at the inflection point. This can be measured directly by hand from the curve if there are many accurate measurement points. However, it is more convenient to fit the curve to a mathematical function and then to calculate the slope at the inflection point in terms of the parameters of the function. This is the approach adopted here. Pairs of experimental points (θ, h) are fitted to the van Genuchten (1980) equation, which is given in Eq. (A1). This equation is probably the most widely used worldwide, and pedo-transfer functions are available for estimation of its parameters. Its use provides a consistent and objective method for data analysis. Equations giving the value of the slope $S = \mathrm{d}\theta/\mathrm{d}(\ln h)$ at the inflection point of the fitted equation are presented in Eqs. (A7) and (A8).

As a simplification, soil porosity can be considered as having two parts: textural porosity and structural porosity (e.g., Guérif et al., 2001). The textural porosity occurs between the primary mineral particles, whereas the structural porosity comprises microcracks, cracks, bio-pores, and macrostructures produced by tillage. Textural porosity is little affected by soil management, whereas structural porosity is sensitive to management factors such as tillage, compaction and cropping. The thesis proposed in this paper is that most of the slope, S, of the water retention curve at the inflection point is mostly due to microstructural porosity, and, therefore, S governs directly many of the principal soil physical properties. Soil with only textural porosity has only very poor physical quality. Except for some sands, such soils usually exhibit poor workability, low water infiltration rates, etc. Therefore, the presence of structural pores and a corresponding large value of S are essential for good soil quality. A review of soil structure and its consequences was presented by Dexter (1988).

The pores that are just draining at the inflection point of the water retention curve can mostly be classified as structural pores or microcracks which can be seen as elongated pores on two-dimensional sections. To a first approximation, it is possible to say that for soil drying between saturation and the inflection point, it is mainly structural pores that are emptying. However, for soil drying below the inflection point, it is mainly textural pores that are emptying. This concept is supported by the findings of Richard et al. (2001) who showed clearly that when a silt loam soil is compacted, it is the structural porosity (i.e., pores which are larger than that which corresponds with the inflection point) which are destroyed, whereas there are simultaneous much smaller gains in textural porosity (sizes smaller than that which corresponds with the inflection point). The results of Richard et al. (2001) support the use of the inflection point of the curve of the logarithm of h against θ for the measure S.

Drying of soil causes shrinkage due to the effective stresses generated by the pore water pressure and the surface tension in the water menisci (e.g., Towner and Childs, 1972). This shrinkage is equivalent to compaction, and this can be essentially irreversible if the soil is dried to a lower water content than at any time in the past. As a result of this effect, drying of a soil (e.g., in the process of measuring the water retention characteristic) can change the pore size distribution and consequently the shape of the water retention curve relative to what it was in situ (Katou et al., 1987; Baumgartl et al., 2000). This effect is unlikely to be important for the results presented in this paper which are mostly for arable top soils. Such soils will have frequently dried to water contents below those of their inflection points. Sub-soils in Mediterranean climatic regions can also have been expected to have dried to the wilting point of crops down to the depth of crop rooting. In contrast, some clayey sub-soils in humid, temperate regions (such as northern Europe) may not have dried intensively, and changes in the shape of the water retention curve upon further drying could be important. However, humid, clayey sub-soils are not the subject of this paper, and so this effect is not considered further in this work.

A comparison has been made of values of S obtained by hand and from the fitted van Genuchten equation for 24 Polish soils. For the "hand" measurements, water retention curves were drawn through the mean values of water content at the 11 different values of suction, h. Then the inflection point (where the curvature is zero) was identified, a tangent to the curve was drawn at this point, and its slope was measured. Comparison of the slopes obtained by the "hand" method and those obtained by calculation from the parameters of the fitted van Genuchten equation is shown in Fig. 2. The regression line in Fig. 2 is given by:

$$\log S(\text{van G}) = -0.728 + 0.585 \log S(\text{hand}),$$
$$(\pm 0.073)(\pm 0.063)$$
$$r^2 = 0.79,\ p < 0.0001 \tag{3}$$

This shows that, on average, the agreement between the two methods is not too bad for values of S smaller than about 0.05, but becomes increasingly worse for larger values of S. For individual soils, the difference between the methods can be large. When making the "hand" measurements, it immediately becomes apparent that drawing the curve, identifying the inflection point and measuring the slope are all very subjective. The method of fitting the van Genuchten (or some other) equation has the considerable advantage of being objective and exactly reproducible by different scientists.

Functions other than that of van Genuchten (1980) can be used to describe the water retention curve. Many of these functions have been discussed in terms of their advantages and disadvantages by Assouline et al. (1998). Any one of these functions could also be fitted to experimental data and used to calculate S. Each of these functions would give different estimations of S depending on its goodness-of-fit around the inflection point and some conversion equation, such as Eq. (3) would be needed to compare the results obtained by different methods.

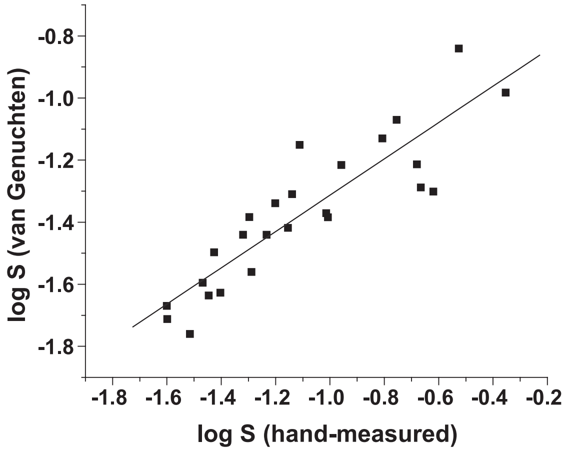

Fig. 2. Comparison of values of log S obtained by two different methods: measurement by hand on graphs of water retention data and calculation using parameters of the fitted van Genuchten equation in Eq. (A8).

A water retention curve if plotted as the logarithm of potential against water content always exhibits at least one inflection point. If the soil exhibits a well-defined hierarchical soil structure (Dexter, 1988), then there may be two or more inflection points corresponding to the drainage of the pores at different levels in the structural hierarchy. A hypothetical soil with a tri-modal structure and a real soil with an apparent bi-modal structure are discussed in Ross and Smettem (1993). Schjønning (1992) also reported bi-modal pore size distributions for Danish soils derived from glacial deposits but reported that the uni-modal van Genuchten model described the pore size distribution of water-sedimented soils reasonably well. However, real data sets for water retention are rarely of sufficient quality to enable the fine detail of the curves to be quantified. In such cases, the fitting of a single van Genuchten curve, as done here, at least takes some account of the presence of microstructure by integrating over the different structural levels. In any case, accurate information on the pore size distributions of soils cannot be obtained from water retention data because of the complicating effects of pore connectivity (Bouma, 1977, p. 30; N.R.A. Bird, personal communication). Irrespective of any fine structure of water retention curves, a larger value of S indicates that part of the porosity is organized into a narrower range of pore sizes, such as a network of microcracks.

Because of the above effects, the method used must always be stated clearly. In the work presented here, the water retention curves have been fitted to the van Genuchten equation with the Mualem (1986) constraint ($m = 1 - 1/n$), unless otherwise specified.

Examination of water retention curves in the literature shows that soil physical degradation always leads to a change in the shape of the curves. It can be noticed especially, that the value of the water content at saturation, θ_{sat}, becomes smaller and that the slope of the retention curve at the inflection point, $S = \tan\,\omega = d\theta/d(\ln\,h)$ also becomes smaller. This is illustrated in Fig. 1B in which the slope $\tan\,\omega_d$ for physically degraded soil is less than the slope $\tan\,\omega_n$ for non-degraded soil. A small slope corresponds to structureless (homogeneous) soil, whereas a large slope corresponds to soil which is structured and which has many pores, such as microcracks, with an effective width of $t_i = \sigma/h_i$.

The main hypotheses that are explored in this series of papers are:

(1) that the slope, S, at the inflection point is a measure of soil microstructure that can be used as an index of soil physical quality, and
(2) that several important soil physical properties can be estimated directly from the value of S.

3. Effects of soil texture

Here, the term "texture" is used to refer to the size distribution of the primary mineral particles in the soil. Different textures give rise to different pore sizes in the soil. When only one size of particle is involved, pore size (e.g., diameter) is proportional to the particle size. However, when a mixture of particle sizes is present, as in soils, the situation is much more complicated because phenomena such as "infilling" of smaller particles in the spaces between larger particles can occur.

For cases where the water retention characteristics of soils with different texture have been measured, it is possible to look at the average effects on the value of the index of soil physical quality, S, directly. An example of this is shown in Fig. 3 where data for 335 Swedish soils have been used (Andersson and Wiklert, 1972, Table 2 and Figs. 58 and 59).

A general downwards trend of S with increasing clay content can be seen in Fig. 3. However, this is not unreasonable because soils with higher clay content are often more difficult to manage. The peak in S between about 30% and 50% clay content is interesting and may be associated with increased development of microstructure in soils with this range of textures.

A useful indirect way to investigate the effects of different soil textures on soil hydraulic properties is to use pedo-transfer functions. These are regression equations that enable the soil hydraulic parameters to be estimated from soil texture. Texture is conveniently quantified in terms of the content of sand, silt and clay. These are defined in the FAO/USDA classification system as particles with effective spherical diameters of >50, 50–2, and <2 μm. Since these are defined so as to make a total of 100%, only two of them are needed to specify the system. Additionally, the organic matter content and the bulk density of the

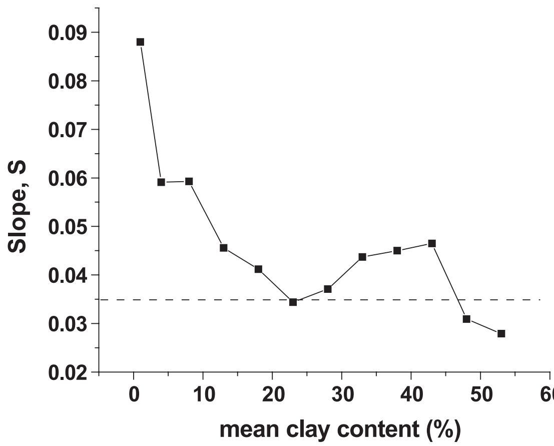

Fig. 3. Values of the slope, S, as a function of the mean clay content of Swedish soils as calculated from the data of Andersson and Wiklert (1972). The broken line shows the "critical value" of $S = 0.035$, as discussed in the text.

soil need to be specified. In this paper, the pedo-transfer functions used are those of Wösten et al. (1999), which are based on the analysis of water retention data of 5521 soil horizons.

Effects of soil microstructure are automatically included in the pedo-transfer functions. This is because the 5521 soil horizons which were used in their development were in their natural condition. Therefore, the predictions from pedo-transfer functions refer to soils with typical amounts and kinds of microstructure. In contrast, the effects of soil macrostructure are not taken into account in pedo-transfer functions because macropores empty at suctions that are too small to be included in water retention measurements. The nature, accuracy, and reliability of pedo-transfer functions has been discussed by Wösten et al. (2001).

For the purposes of illustration of the possible effects of soil composition, we use the following equation which uses the mean of the coefficients obtained by fitting data from 28 Polish top soils with $r^2 = 0.35$ and from 91 Dutch clay soils with $r^2 = 0.78$ (Wösten, personal communication) in which ρ is the soil bulk density in Mg m^{-3}, and C and OM are clay and organic matter contents expressed in % or kg (100 kg^{-1}):

$$1/\rho = 0.590 + 0.00163\ C + 0.0253\ \mathrm{OM} \tag{4}$$

Also, there is the result from analysis of 210 Polish soils from the tilled layer which shows that organic matter content is positively correlated with soil clay content:

$$\begin{aligned}\mathrm{OM} &= 1.58 + 0.048\ C,\\ &\quad(\pm 0.05)(\pm 0.007)\\ r^2 &= 0.19,\ p < 0.0001\end{aligned} \tag{5}$$

A positive correlation between C and OM is usually found. For example, mean data presented by Clayden and Hollis (1984) for the six sub-groups in the UK soil classification system give for 1654 arable soils:

$$\begin{aligned}\mathrm{OM} &= 3.38 + 0.050\ C,\\ &\quad(\pm 0.45)(\pm 0.018)\\ r^2 &= 0.66,\ p = 0.05\end{aligned} \tag{6}$$

and for 1332 soils under permanent (managed) grass:

$$\begin{aligned}\mathrm{OM} &= 5.42 + 0.083\ C,\\ &\quad(\pm 0.59)(\pm 0.023)\\ r^2 &= 0.77,\ p = 0.02\end{aligned} \tag{7}$$

Equations such as Eqs. (5)–(7) are expected to vary significantly between different geographic regions because of the trends of prevailing moisture and temperature as has been shown for the case of Europe (see Fig. 4 in Jones, 2002).

Table 1
Typical values of S for the 12 FAO/USDA soil texture classes together with the parameters used in their calculation

FAO/USDA texture class	p.s.d. (%)		OM (%)	ρ (Mg m^{-3})	θ_{sat} (kg kg^{-1})	α $(hPa)^{-1}$	n	S
	Clay	Silt						
cl	60	20	4.47	1.249	0.395	0.0217	1.103	0.0296
sa cl	42	7	3.61	1.334	0.335	0.0616	1.139	0.0317
si cl	47	47	3.85	1.309	0.362	0.0220	1.104	0.0273
cl l	34	34	3.22	1.376	0.324	0.0400	1.127	0.0285
si cl l	34	56	3.22	1.376	0.325	0.0226	1.129	0.0290
sa cl l	27	13	2.89	1.414	0.299	0.0727	1.169	0.0326
l	17	41	2.41	1.474	0.278	0.0314	1.208	0.0354
si l	14	66	2.26	1.492	0.269	0.0134	1.245	0.0385
si	5	87	1.83	1.552	0.243	0.0045	1.392	0.0485
sa l	10	28	2.07	1.518	0.258	0.0400	1.278	0.0405
l sa	4	13	1.78	1.559	0.239	0.0534	1.406	0.0488
sa	3	3	1.73	1.566	0.226	0.0671	1.581	0.0594

Values of organic matter (OM) content were estimated using Eq. (5) and then values of bulk density (ρ) were estimated using Eq. (4). The values of the parameters θ_{sat}, α, and n of Eq. (A1) were calculated using the values for p.s.d., OM, and ρ in the pedo-transfer functions of Wösten et al. (1999).
sa = sand, si = silt, l = loam, cl = clay.

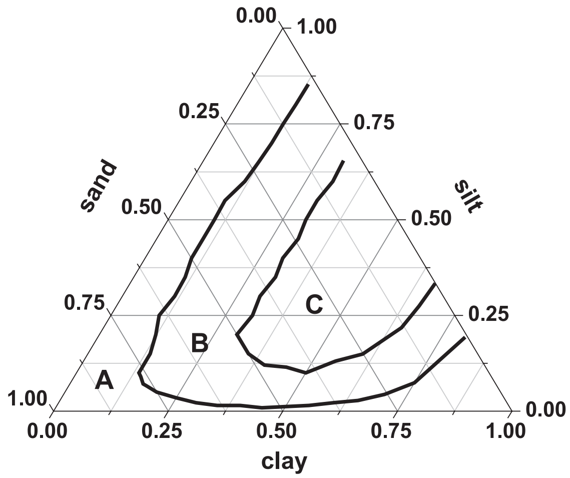

Fig. 4. Ternary diagram showing contours of constant values of S for soils of different texture. Area A is where $S>0.04$, area B is where $0.03<S<0.04$, and area C is where $S<0.03$. The slight bumpiness of the curves was caused by rounding-off errors in the "goal-seeking" program which was used to generate them.

In what follows, Eqs. (4) and (5) are used in all cases for illustrative purposes.

Eqs. (4) and (5) can be useful to study trends for cases in which values of bulk density are not available. In this case, Eq. (4) can be used to estimate ρ. For cases where only the clay content is known, then firstly Eq. (5) may be used to estimate the OM content, and then the known value of C together with the estimated value of OM may be used in Eq. (4) to estimate ρ.

This has been done for the FAO/USDA texture classes. Here, the mean values of clay, silt, and sand for each class were read from the centre of the area for that class when plotted on the standard texture triangle. Then mean typical values of OM and ρ were calculated using Eqs. (4) and (5). Then all of these values were used to estimate the parameters of the three-parameter van Genuchten equation for soil water retention. The value of the slope, S, at the inflection point was then calculated using Eq. (A8). All the values estimated and used are shown in Table 1.

Instead of looking at texture classes, it is instructive to look at continuous functions of texture as can be plotted on ternary diagrams. This is done in Fig. 4, which shows contours for two values of the index of soil physical quality, S.

4. Effects of compaction

Compaction is a reduction of the volume of a given mass of soil. When soil is compacted, the volume of pores is reduced. However, not all pores are reduced similarly. The largest pores are usually lost or reduced in size first (e.g., Richard et al., 2001), and compaction ceases when the soil has become strong enough to withstand the applied stress without further failure. This preferential loss of the largest pores has the effect of changing the pore size distribution and hence the water retention characteristic (e.g., Katou et al., 1987).

The effects of compaction will be illustrated with data from a Spanish soil with 27% clay content and 12.5% silt content at different bulk densities. These data are for profile number SE0504 in the microLEIS data base (De la Rosa et al., 1992; De la Rosa, 2002). The water retention data were fitted to the four-parameter van Genuchten equation, and the values of S were calculated using Eq. (A8). A graph of S against density, ρ, is shown in Fig. 5A. It can be seen that the value of the index of soil physical quality, S, decreases with increasing density.

The regression line in Fig. 5A giving S as a function of soil bulk density, ρ, is $S=0.1171-0.0583\rho$, $r^2=0.74$. These experimental results can be compared with predictions from the pedo-transfer functions. In the pedo-transfer functions, a constant value of organic matter content of 1.5 kg (100 kg)$^{-1}$ was assumed, but

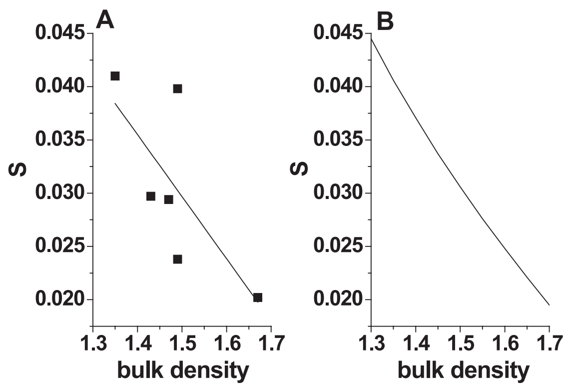

Fig. 5. (A) Experimental values of S as a function of bulk density for Spanish soil SE0504 (a sandy clay loam) shown with the fitted regression line. (B) Predicted values of S for soil of the same texture calculated using pedo-transfer functions as described in the text.

this is not critical because most of the effect of organic matter is indirect through its effect on bulk density (see Eq. (4)). With the pedo-transfer functions, the value of the density can be increased while keeping everything else constant. The predictions from the pedo-transfer functions are plotted in Fig. 5B and can be seen to be almost identical with the regression line through the measured data from profile number SE0504.

It is also possible, using the pedo-transfer function approach to investigate the sensitivity of the index of soil physical quality, S, to bulk density, ρ, for the different FAO/USDA texture classes which are given in Table 1. These results are given in Fig. 6.

However, increase of soil density does not always lead to a reduction of S. Consider the special case of a bed of aggregates in which each aggregates has a small value of S and in which the inter-aggregate pores are too large to retain water and hence to appear on the water retention curve. When such a bed is compacted, the aggregates crumble (Dexter, 1975) and smaller inter-aggregate pores are formed with non-accommodating faces (Pagliai and Vignozzi, 2002). These new inter-aggregate pores may be able to retain water and will result in an increased value of S. This is an example of compaction improving the value of the index of soil physical quality.

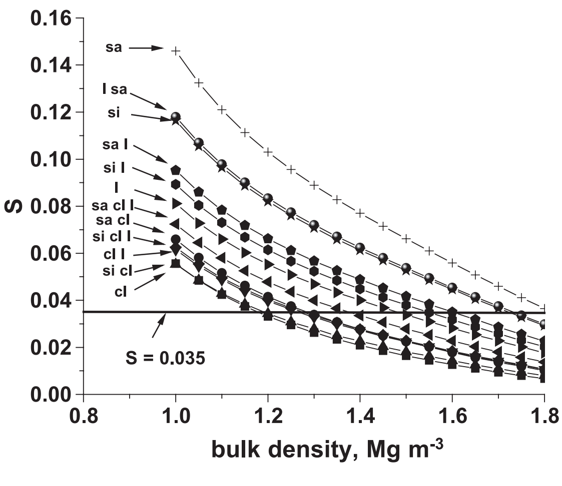

Fig. 6. Predicted values of the index of soil physical quality, S, as a function of soil bulk density. These values were calculated from pedo-transfer functions for the 12 FAO/USDA soil texture classes as described in Table 1. Also shown is a line corresponding to $S=0.035$, which is assumed to correspond to the boundary between good and poor soil physical quality as discussed in the text.

The typical, “critical” values of bulk density at which the curves intersect with the value of $S=0.035$ fit with the experience of this author with pot and field experiments in several countries. These critical bulk densities are also in broad agreement with those presented for Romanian soils by Canarache (1990, pp. 52–54). As stated elsewhere, individual soils may differ significantly in behaviour from these typical values and trends.

5. Effects of soil organic matter content

The effects of soil organic matter content on S have been investigated using experimental data for two soils: a loamy sand soil from Grabów in Poland and a silt loam soil from Highfield at the Rothamsted Experiment Station in England.

At Grabów, the soil has 4% clay, 25% silt, and 71% sand. Plots with different management treatments were established in about 1978, and the results presented here were measured on samples collected from 10 cm depth about 20 years later. More comprehensive details of the soil, treatments, methods, and experimental results are given in Dexter et al. (2001).

At Rothamsted, the Highfield soil has 25% clay, 65% silt, and 10% sand. Plots with different treatments were established in 1949, and the results presented here were measured on samples collected from 10 cm depth about 48 years later. More comprehensive details of the soil, treatments, methods, and experimental results are given in Watts and Dexter (1998) and Dexter and Bird (2001).

Water retention was measured at 12 different water potentials using sand tables and ceramic pressure plate extractors for the samples with different contents of organic matter. These retention data were then fitted to the van Genuchten (1980) equation using non-linear curve fitting. The parameters of the van Genuchten equation which were obtained are presented in Watts and Dexter (1998), Dexter and Bird (2001), and Dexter et al. (2001). An exception is the Highfield arable treatment for which the water retention data have now been fitted to the van Genuchten equation with the residual water content set equal to zero. This puts the data for all the treatments onto a common basis (the resulting values for the arable plot are: $\theta_{sat}=0.2568$ kg kg^{-1},$\alpha=0.0070$ (hPa)$^{-1}$, $n=1.1504$). These fitting parameters were then used to calculate S using Eq. (A8) given in Appendix A. The resulting values of S are plotted against organic matter content in Figs. 7 and 8 for the Grabów and Highfield soils, respectively.

The regression equations for the lines in Figs. 7 and 8 are:

$$\text{for Grabów}: \quad S = -0.009 + 0.038\ \text{OM}, \quad (\pm 0.011)(\pm 0.009)$$
$$r^2 = 0.90,\ p = 0.052 \tag{8}$$

$$\text{for Highfield}: \quad S = 0.0063 + 0.0069\ \text{OM}, \quad (\pm 0.0015)(\pm 0.0004)$$
$$r^2 = 0.99,\ p = 0.00039 \tag{9}$$

Comparison of these equations shows a greater effect of organic matter content on measured values of S for the low clay content soil (Grabów) than for the higher clay content soil (Highfield).

Eqs. (8) and (9) may be used to estimate the values of organic matter content for which $S = 0.035$. These values are OM = 1.2% and OM = 4.2% for the Grabów and Highfield soils, respectively. These values can be considered to be the critical contents of organic matter for these soils. Organic matter contents below these values can be seen in the field to coincide with the occurrence of aspects of poor soil physical quality. In the case of Grabów, these are mainly hard-setting and poor water infiltration, and in the case of Highfield, this is mainly poor workability.

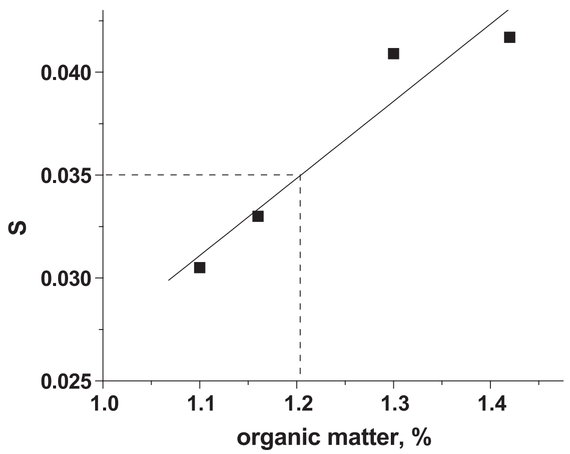

Fig. 7. Experimental results from field plots at Grabów, Poland, illustrating a linear dependence of S on soil organic matter content. The broken lines show the value of $S = 0.035$ and the corresponding organic matter content.

Loveland and Webb (2003) have reviewed the literature on critical levels of soil organic matter and concluded that there are no values below which the soil structure suddenly collapses. Nevertheless, there is a huge amount of literature that shows that decreases in organic matter content are associated with increasingly adverse soil physical conditions. Therefore, the value of $S = 0.035$ may be considered as being an arbitrary but useful value for assessing soil physical quality.

It is possible to compare Eqs. (8) and (9) with predictions from pedo-transfer functions. The measured values of clay, silt, and organic matter content and bulk density were put into the pedo-transfer functions of Wösten et al. (1999) and the resulting estimates of θ_{sat} and n were used to calculate values of S using Eq. (A8) in Appendix A. Regression of the resulting estimated values of S produced the following equations:

$$\text{for Grabów}: \quad S = 0.009 + 0.021\ \text{OM},$$
$$r^2 = 0.955,\ \text{and} \tag{10}$$

$$\text{for Highfield}: \quad S = 0.0049 + 0.0049\ \text{OM},$$
$$r^2 = 0.970. \tag{11}$$

Comparison of Eqs. (8) and (9) with Eqs. (10) and (11) shows that the field soils are showing effects of organic matter that are greater by factors of 1.5–1.8 than are predicted by the pedo-transfer functions. Predictions of the intercepts (for zero organic matter) are closer.

The likely explanation of this is that the organic matter content that is plotted in Figs. 7 and 8 is not the only factor influencing S. Other important factors include crop effects and machinery impacts. As an example, in Fig. 7, the two right-hand points refer to plots that include grass in the crop rotation. Also, the first and third points refer to plots that receive no fertilization whereas the second and forth points refer to plots with a high level of fertilization. In Fig. 8, the two right-hand points were from permanent grass plots. The left-hand point in Fig. 8 refers to soil that is kept bare without plants. Therefore, at both Grabów and Highfield, there are trends of greater values of S in the management

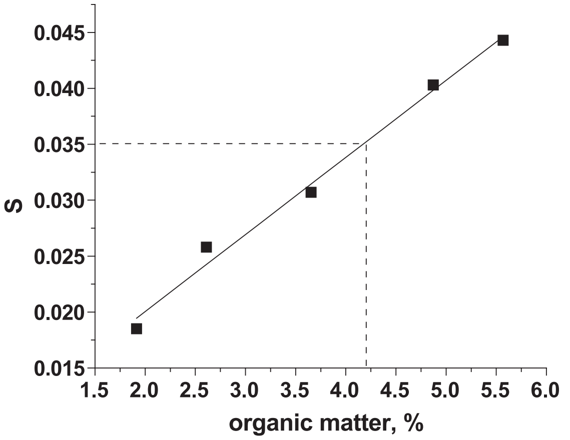

Fig. 8. Experimental results from field plots at Highfield, Rothamsted, England, illustrating a linear dependence of S on soil organic matter content. The broken lines show the value of $S = 0.035$ and the corresponding organic matter content.

systems with greater plant production and/or with lower levels of machinery impact.

This fact is consistent with the view that part of the soil physical benefit attributed in the literature to soil organic matter is due to other factors such as the associated increased crop rooting intensity and reduced machinery impacts when, for example, grass is grown. It could be argued that pedo-transfer functions provide the best method of studying the effects of organic matter alone because they can be developed using data from arable agricultural soils which, on average, are managed conventionally (i.e., using typical crop rotations, typical tillage machinery and typical levels of vehicular traffic). The pedo-transfer functions of Wösten et al. (1999) used in this paper were based on data, the great majority of which were from arable soils, with only few data coming from pasture or forest soils (J.H.M. Wösten, personal communication). In contrast, data such as those presented in Fig. 7 or 8 coming from a range of management systems on a single soil, do not show the effects of organic matter alone but include the effects of different management systems.

Nevertheless, it can be concluded that the use of systems of land management that lead to greater contents of organic matter in the soil also result in higher values of soil physical quality as measured by S. It is likely that the separation of the effects of the different factors on soil physical quality can be aided significantly by the use of S.

6. Effects of sodicity

Most agricultural soils have clay particles, the surfaces of which are saturated with calcium ions. Such soils are stable in water because calcium-saturated clays do not disperse in water but remain flocculated. However, in arid and semi-arid climatic regions and in areas where saline water is used for irrigation, some of the calcium ions can be displaced and replaced by sodium ions. When more than 5–15% of the adsorbed ions are sodium, the soil is described as sodic. Sodic soils exhibit large swelling and shrinkage as they wet and dry. In the presence of free water, the clay particles may disperse completely. Sodic soils are completely structureless and homogeneous and as such represent the extreme case of physical degradation. Consequences are that sodic soils are hard-setting, impermeable, non-friable, and, hence, essentially unworkable.

Much of the water retention data for sodic soils published in the literature cannot usefully be evaluated because of the common usage of volumetric water contents without the corresponding density values. Volumetric water contents for soils of different density cannot be meaningfully compared because they do not have a common basis. In contrast, water ratios (volume of water per unit volume of solids) can be compared. Gravimetric water contents, such as are used throughout this paper, are nearly as good because errors are due only to small differences in particle density between soils. Examples of the use of correct units may be seen in Crescimanno et al. (1995) and in Fig. 26 on pp. 46 of Bresler et al. (1982).

The use of bulk density (mass per unit volume), ρ, is also problematic for soils with changing density because it has the variable quantity (volume) in the denominator. This is why the specific volume, $1/\rho$, was used in Eq. (4).

Interpretation of water retention experiments on sodic soils is problematic because reductions in water content with increases in suction, h, at the wet end of the water content scale within which swelling and shrinkage occur are not due to the drainage of pores as described by Eqs. (1) and (2), but are due mainly to reductions in the separation of the particles. In spite of this complication, the results presented in Fig. 2 (top) in Crescimanno et al. (1995) for water retention curves at different values of exchangeable sodium percentage (ESP) show the trend of decreasing S with increasing ESP, as might be expected.

It may be that for sodic soils, the slope S of the water retention curve at the inflection point cannot be used directly as a measure of soil structural quality and cannot be used to predict other soil physical properties in the same way as appears to be possible for non-swelling soils. Possible effects of sodicity on S could form the bases of hypotheses for future experimental testing.

7. S and root growth

Jones (1983) reviewed the literature which was then available on root growth in soils at a water potential of -330 hPa with different clay contents and different densities. From these published data, Jones (1983) produced two lines (in his Fig. 5) the first of which is the boundary between soil with "few" roots and "no" roots. This line is based on results from 107 soil horizons described as having "no" roots. In fact, 10% of horizons having "no" roots fall above the line given by:

$$\rho = 1.985 - 0.00857C,\ \text{Mg m}^{-3}, \tag{12}$$

where C is the clay content of the soil in kg $(100\ \text{kg})^{-1}$.

The second line is the boundary between soil with "many" roots and "few" roots. This line is based on results from 106 horizons described as having "many" roots. In fact, 10% of horizons having "many" roots fall above the line given by:

$$\rho = 1.882 - 0.00830C,\ \text{Mg m}^{-3}. \tag{13}$$

With the use of the pedo-transfer functions, it is possible to compare his results with the corresponding S values.

Regression lines for some different values of S are:

$$\rho = 1.809 - 0.0114C,\ \text{Mg m}^{-3},\quad r = -0.961,$$
$$\text{for } S = 0.030,\ \text{and} \tag{14}$$

$$\rho = 1.960 - 0.0105C,\ \text{Mg m}^{-3},\quad r = -0.977,$$
$$\text{for } S = 0.020. \tag{15}$$

A simplified version of the results of Jones (1983) is shown in Fig. 9A. Comparison with the predictions from the pedo-transfer functions shown in Fig. 9B

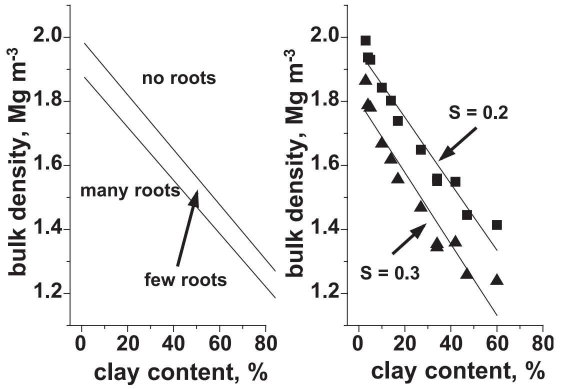

Fig. 9. (A) Simplified version of the results of Jones (1983) showing the critical values of soil bulk density for root growth as a function of soil clay content. (B) Predicted values of soil bulk density as a function of clay content for the 12 FAO/USDA soil texture classes for fixed values of $S = 0.2$ and 0.3. These values as functions of S were obtained using pedo-transfer functions as described in the text.

shows that to a first approximation, root growth will not occur for values of $S < 0.020$ and that only a few roots will grow if $0.020 < S < 0.030$. Adequate root growth typically requires values of $S > 0.030$.

The results in Fig. 9A and B can be considered to be the same to within experimental error in view of the scatter of the original values summarized by Jones (1983) and the fact that the experiments reviewed by Jones (1983) had used American soils whereas the pedo-transfer functions of Wösten et al. (1999) were based on west European soils.

All predictions from pedo-transfer functions indicate only typical values or trends, and exceptions are common. A good example is that of a uniform sand that has been well-sorted by the action of wind or water. Such a sand, if the particles approximate to spheres, can be packed to a maximum density of only 1.69 Mg m^{-3}, and root growth would cease at a density lower than this. This clearly illustrates that the intercept of $\rho = 1.99$ Mg m^{-3} given by Eq. (12) is not a good prediction for all soils with zero clay content. Examples of sands for which root growth ceases at relatively low densities occur in Denmark (P. Schjønning, personal communication).

It may at first look as though the connection between S and root growth is entirely coincidental. However, there is a logical, physical connection. When roots grow through soil, they compress cylindrically the soil around them (Cockroft et al., 1969; Bruand et al., 1996). Indeed, soil compressibility can be used to assess soil rootability (Römkens and Miller, 1971). Soil compressibility is influenced strongly by the presence of structural pores (Dexter and Tanner, 1973). Since the quantity S is a measure of the amount of microstructure in the soil, soil compressibility may be expected to be positively correlated with S. This is consistent with the finding by Katou et al. (1987) that the shape of the water retention curve is indicative of the susceptibility of soil to compression. Hence, root growth is logically expected to be positively correlated with S. Experimental and theoretical investigations of the quantitative connection between S and soil compressibility is a subject of on-going research by the author.

From all of the above, it is clear that in most soils, larger values of S indicate soil that is more suitable for root development. Therefore, these results support the concept of S as an indicator of soil physical quality which appears to be independent of soil texture with the possible exception of some sands.

8. Reference values for S

Considerable practical experience together with the evidence presented here suggests that the boundary between soils with good and poor soil structural quality occurs at value of approximately $S = 0.035$. No sudden change in soil properties occurs at this value, but this value is consistent with field experience with widely differing soils. Values of $S < 0.020$ are clearly associated with very poor soil physical conditions in the field.

9. Conclusions

The soil physical parameter, S, defined in this paper can be considered as an index of soil physical quality that is consistent with observations on soil compaction, on effects of soil organic matter content, and on root growth. The underlying reason is that S is a measure of soil microstructure that controls many key soil physical properties.

Soil quality has been of increasing concern in recent years. It is usually considered to have three main components: physical quality, chemical quality, and biological quality. Of course, these are not independent because, for example, the biological status of soil depends very strongly on the prevailing physical and chemical conditions. However, there is little doubt that an improved measure of soil physical quality could contribute greatly to the overall assessment of soil quality. It is hoped that the parameter, S, which has been introduced here will prove to be useful for this purpose.

Soil physical quality, as defined, can be easily and unambiguously measured. Nearly every soils laboratory has the equipment necessary to determine the water retention curve of soil samples. This measurement is more uniform and reproducible than many other soil physical measurements.

In this paper, the water retention curves were fitted to the van Genuchten (1980) equation. This has the advantage that this produces an analytical expression for the index of soil physical quality, S. This, in turn, has the advantage that pedo-transfer functions that have been published for the parameters of the van Genuchten equation can be used. This approach enables typical trends in the effects of several factors on the index of soil physical quality, S, to be investigated. Even without fitting any mathematical function to water retention data, it is still possible to obtain an estimate of S from the slope at the inflection point by measuring it directly on a graph. However, it has been shown that this can give a different estimate of the numerical value of S. This illustrates the importance of specifying the methodology used.

The index S provides a scale that can be used to compare easily different soils or the effects of different soil physical conditions and management practices. Applications of S in agricultural soil mechanics and in soil hydrology are explored in Parts II and III of this series of papers.

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Appendix A. Derivation of the position and slope of the inflection point of the van Genuchten water retention curve

The van Genuchten (1980) equation for water retention is

$$\theta = (\theta_{sat} - \theta_{res})[1 + (\alpha h)^n]^{-m} + \theta_{res}, \tag{A1}$$

which may be plotted as curves of log h against θ, as shown in Fig. 1.

We may write:

$$\frac{d\theta}{d\ln(h)} = \frac{d\theta}{dh}\frac{dh}{d\ln(h)}, \tag{A2}$$

where ln(h) is the natural logarithm of h. Therefore,

$$\frac{d\theta}{d\ln(h)} = -mn(\theta_{sat} - \theta_{res})\alpha^n h^n [1 + (\alpha h)^n]^{-m-1} \tag{A3}$$

At the inflection point:

$$\frac{d^2\theta}{d\ln h^2} = -mn(\theta_{sat} - \theta_{res})\alpha^n \{nh^{n-1}[1 + (\alpha h)^n]^{-m-1} + h^n(-m-1)\alpha^n nh^{n-1}[1 + (\alpha h)^n]^{-m-2}\}h = 0 \tag{A4}$$

Therefore, the modulus of the water potential at the inflection point is:

$$h_i = \frac{1}{\alpha}\left[\frac{1}{m}\right]^{\frac{1}{n}} \tag{A5}$$

Substitution back into Eq. (A1) gives the water content at the inflection point as:

$$\theta_i = (\theta_{sat} - \theta_{res})\left[1 + \frac{1}{m}\right]^{-m} + \theta_{res} \tag{A6}$$

The slope, S, at the inflection point is obtained by substituting Eq. (A5) into Eq. (A3), which gives:

$$S = -n(\theta_{sat} - \theta_{res}).\left[1 + \frac{1}{m}\right]^{-(1+m)} \tag{A7}$$

or, if the Mualem (1986) constraint, $m = 1 - 1/n$, is applied:

$$S = -n(\theta_{sat} - \theta_{res}).\left[\frac{2n-1}{n-1}\right]^{\left[\frac{1}{n}-2\right]} \tag{A8}$$

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