Soil & Tillage Research 115–116 (2011) 1–15

Contents lists available at ScienceDirect

Soil & Tillage Research

journal homepage: www.elsevier.com/locate/still

Review

A review of the changes in the soil pore system due to soil deformation: A hydrodynamic perspective

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ARTICLE INFO

Article history:
Received 8 February 2011
Received in revised form 9 June 2011
Accepted 12 June 2011

Keywords:
Soil deformation
Soil pore connectivity
Pore geometry
Water retention curve
Hydraulic conductivity function

ABSTRACT

Compaction and shearing, as well as the rearrangement of soil aggregates and clods due to shrinkage, among other processes, can strongly affect the pore geometry of agricultural soils. These soil structural changes directly affect soil water movement by altering the hydraulic properties that are commonly described by the soil water retention curve (WRC) and the unsaturated hydraulic conductivity function (HCF). This review focuses on recent advances in the understanding and evaluation of changes in hydraulic functions in relation to compacted soil. The development of hydromechanical models due to recent advances with more sophisticated methods enables quantification of the effects of compaction on the hydraulic conductivity functions at the pore scale of aggregated soil. However, it remains unclear how to up-scale the dynamic, in terms of inter-aggregate pore models, into the continuum-scale dual-porosity models in the form of effective parameters, particularly regarding effective hydraulic properties for the preferential flow domain. While hydromechanical models fail to describe water flow and hydraulic conductivity at the relevant scales and water saturation ranges, the continuum-based flow models rely on effective parameters that are mainly empirical or are based on fitting model results to data. Input data usually do not address temporal changes in the arrangement of aggregates induced by soil compaction and shrinkage. This review presents a concept that summarizes the changes in structural and textural porosity upon compaction. It suggests focusing on the extension of existing hydraulic and hydromechanical models to include the pore structural changes that account for the movement and rearrangement of soil aggregates and the resulting changes in the soil hydraulic properties which basically manifest the effects of shearing and compaction on water flow.

Contents

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doi:10.1016/j.still.2011.06.002

1. Introduction

Cultivated soils are particularly affected by soil deformation (Berli et al., 2004; Håkansson and Reeder, 1994). It has frequently been recognized that compaction reduces total porosity and increases bulk density while also reducing the proportion of larger pores that play an important role in water movement and solute transport, nutrient availability, aeration, and crop productivity, owing to the limitation in root growth (e.g., Domżał et al., 1991; Mapfumo et al., 1998; Arocena, 2000). Agricultural field operations employed at various levels of mechanization depend heavily on wheeled tractors as a source of traction power. It is common practice to use the same tractor for different operational requirements. Hence soils with different load-bearing capacities are exposed to repeated compressive stress of the same magnitude. This results in the formation of a dense soil layer which has low hydraulic conductivity and aeration (e.g., Lipiec and Hatano, 2003).

Soil structure may be significantly altered due to either mechanical stress (e.g., tractor traffic, tillage and cattle trampling) or hydraulic stress from natural wetting and drying cycles. While many studies have investigated soil compaction caused by wheel traffic, few investigations (e.g., Cui et al., 2010; Keller et al., 2002; Berli et al., 2004) deal with the extent of soil compaction on agricultural land due to tracked heavy machines. Track-type tractors have the potential to cause less soil compaction because the tracks usually have a greater surface area than wheels for tractors with equivalent power ratings (Brown et al., 1992). Pagliai et al. (2003) have shown that the effect of soil compaction with a rubber-tracked tractor was limited to the surface layer only.

In addition to wheeled and tracked machines, stock trampling affects soil in different ways, depending on several conditions: (i) trampling intensity; (ii) soil moisture (iii) soil type; (iv) plant type; (v) field slope and (vi) land use type (e.g., Zhao et al., 2010; Krümmelbein et al., 2006). Trampling-induced soil compaction is characterized by its spatial heterogeneous distribution. It mainly affects pore geometry (or structure) at the soil surface (Nie et al., 2001; Vzzotto et al., 2000) and topsoil matrix (Alaoui and Helbling, 2006). The depth of soil compaction induced by pugging depends on animal weight, soil moisture, hoof size and kinetic energy.

The structure of soils can also change continuously through water menisci forces during drying and wetting cycles (hydraulic stress) (Peng et al., 2007). In non-rigid soils, two shrinkage components with vertical and horizontal directions can be quantified. Vertical shrinkage results in soil subsidence, while horizontal shrinkage produces soil cracks. Pre-existing soil cracks induce surface water to reach more directly the water level as preferential flow (Liu et al., 2003) and potentially lead to unintended contamination of ground- and surface waters by agrochemicals (e.g., Borggaard and Gimsing, 2008; Jarvis, 2007).

To date, a number of methods have been used to demonstrate the occurrence and effects of soil compaction. The most commonly used parameters are bulk density and total porosity (Boone, 1988; da Silva et al., 1994), macroporosity (Alakukku, 1996), penetration resistance (PR) (Pagliai, 1998), air permeability (Ball, 1981; Reszkowska et al., 2011), saturated hydraulic conductivity, K_{sat} (Alakukku, 1996), pre-consolidation pressure (Horn, 1981; Kirby, 1991), dye surface density (Kulli et al., 2003; Alaoui and Helbling, 2006; Alaoui and Goetz, 2008) and infiltration capacity (Alaoui and Helbling, 2006; Blanco-Canqui et al., 2010).

The change in bulk density due to compaction is a global indicator that integrates information about the total change in the volume of voids of soil under consideration. But it cannot account for changes in the volume distribution of these voids, their connectivity, or the changes in this connectivity (e.g., Vogeler et al., 2006), especially during elastic compaction, which has a significant impact on hydraulic properties (Lenhard, 1986). In fact, it has been shown that compaction results in not only discontinuity of the structure between top and subsoil that promotes ponding at the soil surface (Gysi et al., 1999; Kulli et al., 2003; Alaoui and Helbling, 2006) but also a change in the proportion of textural and structural porosity (e.g., Bruand and Cousin, 1995; Richard et al., 2001).

Soil compaction and shearing caused by external factors affect pore geometry and consequently water retention (Wu et al., 1997; Richard et al., 2001; Tarantino and Tombolato, 2005; Dexter et al., 2008; Gregory et al., 2010; Li and Zhang, 2009; Wójciga et al., 2009; Matthews et al., 2010). In most soils, especially fine-textural soils, there is also a decrease in the pore volume as the soil shrinks (Simms and Yanful, 2001; Braudeau et al., 2005; Whalley et al., 2008; Li and Zhang, 2009). Both deformation and shrinkage have significant implications for soil physical functions (Gregory et al., 2010).

The purpose of this paper is to review recent advances in the understanding and evaluation of functional changes in compacted soil with respect to water retention and water flow.

2. General considerations

Soil pore systems can be defined as the ensemble of the voids or spaces existing in a given volume of soil. Water flowing through connected pores involves the notion of structural hierarchy (Brewer, 1964; Hadas, 1987; Dexter, 1988; Dexter et al., 2008). The mean sizes of the pores separating the aggregate particles at progressively higher levels are themselves progressively bigger (e.g., Dexter et al., 2008). The spatial arrangement or clustering of primary soil particles into secondary units called aggregates or peds, (and clods often in cultivated soil) is known as *soil structure*.

The solid phase of soils consists of discrete units called primary soil particles. These particles may vary widely in size, shape and composition. In this study, we define *textural (or matrix) porosity* as the pore space between the primary particles or intra-aggregate pores (or micropores).

Structural porosity is the pore space between micro-aggregates or aggregates and contains macropores such as inter-pedal voids, biopores or desiccation cracks that have a significant effect on the water flow and solute transport processes and offer greater potential for undisturbed growth of *plant* roots.

Biopores and the network of inter-aggregate pores, cracks and other structural voids can act as preferential flows through which most of the volume of the bulk soil matrix can be bypassed during infiltration from storms (e.g., Beven and Germann, 1981; Gerke, 2006; Jarvis, 2007). Efficiency of preferential flow paths is measured by pore morphology (Brewer, 1964), geometry (Perret et al., 1999a,b), hydrodynamic aspect and functionality (e.g., Bouma et al., 1977; Beven and Germann, 1982; Chen et al., 1993), equivalent pore diameter (Sekera, 1951; Luxmoore, 1981), or size (Velde et al., 1996). In soil science, the term "*macropore*" has been used to represent the whole spectrum of pores ranging from inter-aggregate pores, shrinkage cracks, and fissures to earthworm and root channels (Tisdall and Oades, 1982; Elliott, 1986; Oades, 1993; Amezketa, 1999; Braudeau et al., 2004; Bronick and Lal, 2005; Abou Najm et al., 2010). In this review, the term "macropore" is used to define the non-capillary biopore in which water flows under gravity. A mean diameter >30 μm is used to define macropore if no diameter is specified; otherwise, it is set according to the specific study cited in this review.

The dry soil bulk density and associated total porosity is the most general and frequently used parameter to characterize the state of compactness and physical properties of soil (Panayiotopoulos et al., 1994; da Silva et al., 1994; Assouline, 2006a, 2006b). However, its usefulness in characterizing compaction effects on storage and transport of water and air in terms of connectivity and continuity is not obvious. This is due to the fact that at given bulk density for the same soil, the pore geometry and continuity can differ due e.g. to soil management practices. For this reason bulk density alone is not always a sensitive indicator of soil compaction effects, particularly with respect to transport properties (e.g. Gebhardt et al., 2009; Horn et al., 2003; Lipiec et al., 2003). However, in the case of various soils, a particular bulk density may indicate an extremely compact state in one soil with reference to its natural uncompacted state, but a very loose state in another one due to differences in texture and organic matter content. To overcome this problem, the actual bulk density is expressed as a percentage of the reference-compaction state of a given soil known as "*degree of compactness*" or "relative compactness" (Håkansson, 1990; Håkansson and Lipiec, 2000). In the same way, the pore space can be quantified by the void ratio frequently used in soil mechanics and soil physics and defined as the volume of the pores per unit volume of solid. The fact that the denominator is constant enables the void ratio of different types of pores to be compared, even in soil where pore space may vary with shrinkage/swelling processes or under compaction/shearing (e.g., Dexter et al., 2008). Zhang et al. (2006) also used water volume ratio expressed as the volume of water per unit volume of solid phase, which does not depend on the changes in soil bulk density and is appropriate variable to use for swelling soils. These relative compaction parameters are more useful than bulk density or total porosity in studies of the effects of field traffic on soil structure and consequently on root and crop response (Canarache, 1991; Håkansson and Lipiec, 2000). By using the relative compaction instead of the bulk density performance and applying the concept of the least limiting water range, LLWR (defined as the ideal soil water content range, in which the limitations for root growth were due to the availability of water, air, and PR were minimal) are enhanced (da Silva et al., 1997).

3. Hydrodynamic aspects

Hydrodynamics here are defined as interactions between forces of moving fluid and soil structure related to soil pore aggregate systems. In fluid dynamics, hydrodynamics describe the liquids in motion and their various properties such as velocity, pressure, density, and temperature, as functions of space and time (Corey, 1994). Not only particles but also soil aggregates and clods are subject to movement under intense external perturbations, as during repeated wheeling, leading to the formation of a coherent and very compacted soil horizon (Horn et al., 2003). The non-rigidity of soil systems can easily be observed in the field; most soil properties change dynamically with time (e.g., Horn, 2004; Horn et al., 2003). The dynamic involves coupled interaction between soil mechanical and hydraulic processes (e.g., Horn et al., 2003; Horn and Smucker, 2005). The management of arable soils basically attempts a soil structure optimization between sufficient stability for cultivation and favourable soil ecological conditions for crop growth and environmental functioning (e.g., Roger-Estrade et al., 2000).

3.1. Pore system

A dual-porosity medium is defined as a pore system continuum, characterized by two types of pores: structural and textural. Inter-aggregate and intra-aggregate soil porosities have been used to designate the dual soil pore system. For preferential flow and soil hydraulic modelling (e.g., Gerke and van Genuchten, 1993), the two-pore domain concept was applied for flow in structured soil, assuming still rigidity of the soil. With respect to describing the effects of compaction on mechanical properties of structured soil, different studies carried out on soil compaction showed that up to a certain level of compaction, depending of the amplitude of the strength, the aggregates remain rigid and only the inter-aggregate structure is affected by the compaction. In fact, macroporosity is more sensitive to compaction than total porosity (Alakukku, 1996). Li and Zhang (2009) reported that the volume of the compressible inter-aggregate pores is closely related to the final void ratio of the compacted soil and that changes to inter-aggregate pores are dominant during compaction, but changes to intra-aggregate pores are dominant during saturation and drying. In fact, in clay soil the volume of intra-aggregate pores was found to agree closely with the volume of pore water, providing support for the assumption that in an unsaturated aggregate microstructure the clay aggregations are saturated (Monroy et al., 2010).

The pore geometry has a large influence on the compressibility of soils. In fact, soils with a high proportion of vertically oriented pores are less susceptible to compaction (even more for cylindrical pore shapes e.g., Schäffer et al., 2008) than soils with predominantly horizontal pores (Hartge and Bohne, 1983) and the inter-aggregate pores in tilled soils (Schäffer et al., 2008). Due to low stability, the tillage macropores (inter-aggregate pores) are not efficient except in the period immediately after tillage when they are not effectively isolated (i.e. not "relict") (Dexter and Richard, 2009). As a consequence, tillage leaves the structural pores that are mostly planar (i.e. micro-cracks) as the principal pathways for hydraulic conductivity in most field situations. This highlights the important role of the cylindrical (bio)pores in soil that may still enable sufficient soil drainage and aeration when other pores have already collapsed. Jégou et al. (2002) showed that soil compaction helps to close numerous pores, reduces the mean length of burrows, and increases the number of fragmented burrows. They concluded that soil compaction affects the functionality of burrow systems to a large extent by affecting their continuity.

It is known that the first pass of a wheel causes a major portion of the total soil compaction (Bakker and Davis, 1995), whereas repeated traffic with low axle loads can affect the subsoil and the effects can persist for a long time (Balbuena et al., 2000). Servadio et al. (2001) found that a massive structure was evident for samples subjected to four passes of a rubber-tracked (RT) tractor, characterized by few and thin, elongated pores (300–500 μm). Lipiec and Håkansson (2000) reported that the number of passes increases the degree of compactness from 81 for zero passes to 93 for eight passes, also reducing the macroporosity from 13.4 to 5.6 (% v/v). Pagliai et al. (2003) reported that after four passes of a wheeled tractor, the traffic caused considerable damage to soil structure, resulting in a complete disappearance of the elongated pores (>500 μm) and the reduction of the transmission pores (50–500 μm). They reported that the damage was more pronounced after the passage of a rubber-tracked tractor, where the transmission pores were reduced from 24.2 to 2.2%. This modification resulted in a more compacted massive structure in which most present pores were completely isolated in the soil matrix. Horn et al. (2003) noted full homogenization of the pore system with increasing wheeling. Kim et al. (2010) conducted field experiments on Mexico silt loam with field treatments of uniformly Compacted (C) and Non-Compacted (NC) plots. They found an increase in bulk density of C treatment of 8%, and a decrease in saturated hydraulic conductivity of 69%. The CT-measured number of pores also decreased by 71%, the number of macropores (>1000 μm diameter) by 69% and coarse mesopores (200–1000 μm diameter) by 75%, with the most pronounced effect in the upper 10 cm of the soil layer. Similar results were reported by Udawatta and Anderson (2008) and Udawatta et al. (2008). This damage may reach deeper soil horizons. Dumbeck (1984) carried out traffic experiments on arable land with heavy excavators (weighing up to 4.7×10^4 kg, mean stress in the contact area up to 100 kPa) under dry (-300 to -1000 hPa soil water potential) and wet (≈ -60 hPa soil water potential) soil conditions. He found a decrease in the amount of macropores to 0.65 m in the dry and 1 m in the wet soil.

Gebhardt et al. (2009) reported that a decrease of coarse pores (>50 μm) in their sandy Podzol was accompanied by a relative increase of intermediate pores (10–50 μm) already at loading of 70 kPa and medium pore size (0.2–10 μm) of 230 kPa. Despite the decrease of wide coarse pores in the Podzol due to compaction, neither significant change in K_{sat} nor in bulk density was shown. On the other hand, in contrast to the clay-rich soils, sandy soil showed a distinct decrease in the air conductivity subsequent to compaction, as a growing fraction of the remaining coarse porosity increasingly filled with water and then no longer conducted air. In fact, it was shown that the high coarse porosity of the sands is predominantly texture-dependent and enhancement of the stability of these soils during compaction due to grain-to-grain contact points occurs without resulting in an entire loss of their coarse porosity (Gebhardt et al., 2009). This implies a change in the matric potential with saturation under stress conditions until no more air is conducted. In contrast, more finely textured soils are extremely susceptible in terms of harmful compaction. The authors concluded that soil-protection strategies should be focused on such vulnerable soils. In another study on silty clay loam (Blanco-Canqui et al., 2010), a decrease of the same pores (>50 μm) by wheel compaction was accompanied by considerable reduction of cumulative infiltration, saturated hydraulic conductivity, soil water retention at 0 kPa, plant-available water, and effective porosity.

Another important aspect of pores is the tortuosity of the pore system, since continuous large pores play an important role in allowing roots, gas, and water to penetrate into the soil. The expanding use of third generation X-ray microtomography enhances new applications of CT for more detailed descriptions of soil compaction processes, including local changes in soil structural pore space characteristics and deformation (Peth et al., 2010), decreases in connectivity of macropores (Schäffer et al., 2007, 2008), and increases in the fraction of spherical pores (and thus the convexity of the pore space) with no clear effect on the orientation of pores (Schäffer et al., 2007).

3.2. Water flow and solute transport

Appropriate methods for investigating the dynamic of flow consist of direct measurements of water and solute transport in soils. The most common and established methods for this purpose employ various types of water samplers or lysimeters (e.g., porous ceramic suction cups, zero tension pan lysimeters, gravity drainable soil columns, passive capillary fiber glass wick samplers (PCAPs), weighing lysimeters and tile drainage samplers).

Most of the transport processes in soil occur through its pores via the soil solution which refers to soil water that contains various agricultural chemicals (fertilizers and/or pesticides). The decrease of the macropore space combined with the distortion of structural porosity due to compaction induces solute transport in small and low mobile transport domains that results in slower solute migration (Heitman et al., 2007).

A recent study by Besson et al. (2011) using retardation factor (the ratio between piston-flow velocity and the mean solute transport velocity) showed that the retardation of solute movement can be due to both dense seedbed aggregates and the presence of a plough pan. Limited water and solute transport in tightly compacted zones may promote solute fluxes towards preferential pathways as wormholes or activated macropores (provided that water ponding occurs at the soil surface) (Kulli et al., 2003) and more permeable zones (Coquet et al., 2005). This results in solutes by-passing the shallower soil surface horizon where transformation processes (degradation of pollutants by biological activity) and retention processes are the most active as a consequence of environmental pollution (Coquet et al., 2005). Some authors (Coquet et al., 2005, Besson et al., 2011) reported that the redirected solute flow resulted in an "umbrella" (shadow) effect in deeper soil. In a study of Besson et al. (2011) the "umbrella" effects on breakthrough curves (the long tailing) and associated low permeability in transport were observed at a depth of 0.15 m as well as at depths of 0.30 and 0.45.

The effect of soil compaction on solute transport is related to the type and nature of the given chemical compound. Sadegh-Zadeh et al. (2008) reported that after irrigation the movement of nitrogen and potassium downward is reduced as the compaction level increases, but the movement of phosphorus is increased because of high moisture and movement by diffusion.

Although the measurement methods may capture both matrix and macropore flow under both saturated and unsaturated soil conditions (Boll et al., 1992), they are tedious and time consuming. Therefore, numerous models were developed to investigate water flow and solute transport through preferential pathways, crossing the vadose zone. As indicated in review papers, predicting movement of water and chemicals is mostly based on the Darcy/Richards one-dimensional flow equation (e.g.; Lipiec et al., 2003; Šimůnek et al., 2003; Jarvis, 2007), and the effect of soil compaction is considered by changing hydraulic conductivity, water retention, root growth and physicochemical and biological transformation of the chemicals (Lipiec et al., 2003). However, such modelling approaches of solute transport in soil may still be found inappropriate when solutes are transported by preferential flow processes (Coquet et al., 2005).

3.3. Water retention curve

Soil moisture status is the most important factor influencing soil compaction processes (Soane and van Ouwerkerk, 1994). Zhang et al. (2006) applied two treatments, C1 and C2 (corresponding to an increase in bulk density by 10 and 20%, respectively) to two silty loam sites, Heyang (Chromic Cambisol) and Mihzi (Calcic Cambisol). They found that water retention curves for both the surface (0–0.05 m) and subsurface (0.10–0.15 m) layers at the two sites were significantly changed by tested levels of soil compaction. They found that a high level of compaction C2 significantly decreased the water content of the surface layer at tensions of <2 kPa for Heyang and $\leq$8 kPa for the Mihzi site.

In other studies, analysis of water retention curves showed that compaction results in a decrease of water content at high matric

Table 1
Relevant studies showing mixed changes in micro- and macropore domain including the increase of "lacunar porosity".

Authors	Soil texture/ classification	Decrease of pore class of diameter D_1 (μm) (corresponding load and/or water potential)	Increase of pore volume class of diameter D_2 (μm) (corresponding load and/or water potential)	Effect on hydraulic parameters
Bruand and Cousin (1995)	Clay loam/chromic luvisol	$D_1 > 5.6$ (compaction tests; 50 and 200 kPa, water potential of −1 kPa)	$0.18 < D_2 < 5.6$ (compaction tests; 50 and 200 kPa, water potential of −1 kPa)	K_{sat}^{+}
Richard et al. (2001)	Silt/orthic luvisol	$D_1 > 4$ (tractor of 81.4 kN, tyres pressure of 200 kPa)	$1 < D_2 < 4$ (tractor of 81.4 kN, tyres pressure of 200 kPa)	K^{+} (water potential <−15 kPa)
Tarawally et al. (2004)	700–780 g kg^{-1} clay/ Rhodic ferrasol	$D_1 > 50$ μm (seven passes of a 10 Mg tractor)	$D_2 < 0.5$ (seven passes of a 10 Mg tractor)	TP^{-}
Koliji et al. (2006)	Sandy loam/n.d.	$2.5 < D_1 < 35$ (compaction test; suction of 400 kPa)	$0.15 < D_2 < 0.9$ (compaction test; suction of 400 kPa)	n.d.
Schäffer et al. (2007)	Restored soil/ eutric cambisol	D_1 (defined as pores drained at field capacity) Decrease of D_1 by 2% for 2 passes and by 8% for 10 passes	D_2 (=fine to intermediate pores, diameter not defined). Increase of $D2$ by 2% for 2 passes and by 5% for 10 passes	BD+
Gebhardt et al. (2009)	Medium sand/podzol	$D_1 > 50$ (compaction tests; 70 kPa)	$10 < D_2 < 50$ (compaction tests; 70 kPa) $0.2 < D_2 < 10$ (compaction tests; 230 kPa)	K_{sat}^{-}

TP: total porosity; BD: bulk density; K: hydraulic cond.; K_{sat}: saturated hydraulic cond. "−": not effective in reflecting treatment effect "+": effective in reflecting treatment effect "n.d." not defined.

potentials (from 0 to −10 kPa) and an increase of water content at low matric potentials (from −250 to −1550 kPa) (Walczak, 1977; Domżał, 1983; Kutílek and Nielsen, 1994; Ferrero and Lipiec, 2000). However, a slight effect occurs in the intermediate potential range. This reflects the fact that, under compaction, as the proportion of large pores decreases, the proportion of small pores increases (Assouline et al., 1997; van Dijck and van Asch, 2002) or remains unaffected (Matthews et al., 2010). The two processes are linked by an "inflection point" (Dexter, 2004a) above which, for soil drying, mainly structural pores are emptying and below which mainly textural pores are emptying. Based on mercury porosimetry analysis, Richard et al. (2001) showed that the decrease in the volume of pores retaining water between −5 and −20 kPa corresponds to a decrease in the volume of pores larger than 4 μm. In the same way, they attributed the increase in pore volume retaining water at potentials ranging between −20 and −80 kPa to increase in the volume of pores between 1 and 4 μm. They suggested that the increase in water retained at potentials between −20 and −80 kPa is due to the formation of relict structural pores being remnants of structural pores distorted during traffic and accessible only through the necks of textural (lacunar) pores (Bruand et al., 1997; Richard et al., 2001). This new structure resulted from the destruction of structural pores. Relevant studies on this topic are listed in Table 1. Applying pressures of 50 and 200 kPa for a water potential of −1 kPa, Bruand and Cousin (1995) reported a drastic decrease in the structural porosity of a diameter larger than 5.6 μm, resulting in increases in lacunar porosity of 16 and 33% respectively when compared to original values. As a consequence, water retention curves (WRCs) are mainly altered at low pressure heads (from 0 to 10 kPa) and at high pressure heads (from 300 to 1500 kPa) (Kutílek and Nielsen, 1994; Ferrero and Lipiec, 2000). Compacted soils present flattened WRCs, with a reduction of the slope of the WRC at the inflection point (Dexter, 2004a; Assouline et al., 1997; Hayashi et al., 2006). It was also suggested that relict structural pores and structural pores could interact to determine together the hydraulic properties of soil as water retention curve and hydraulic conductivity function (HCF). Richard et al. (2001) reported that this point corresponds to a potential equal to −20 kPa for a silt soil.

3.4. Saturated hydraulic conductivity

Water flow in aggregated soils depends mainly on pore structure and hydraulic properties of soil aggregates that are modified by soil compaction. In reduced tillage, K_{sat} has often been found to increase, despite a higher bulk density, while in other experiments a small increase in bulk density may decrease the conductivity by several orders of magnitude (Arvidsson, 1997). However, in compacted soil, increase in bulk density and decrease in the pore volumes with equivalent pore diameter >150 μm or >60 μm (depending on the site) resulted in significantly lower K_{sat} values (Zhang et al., 2006). Also, reduction of similar pores (>50 μm) due to trampling was reflected in decreased saturated hydraulic conductivity (Zhao et al., 2010). These values are predominantly governed by the abundance of relatively large pores and their continuity (Pagliai et al., 2003; Lipiec and Hatano, 2003; Zhao et al., 2010). Therefore, change in this group of pores tends to have a strong influence on K_{sat} values. In the study by Kim et al. (2010), a regression equation with CT-measured macroporosity, the area of largest pore and porosity explained nearly 80% of variability in the saturated hydraulic conductivity of compacted and non-compacted soil. In contrast, Gebhardt et al. (2009) reported that K_{sat} values do not reflect the compaction effect (Table 1).

Other studies showed that soil compaction and increased contribution of finer pores result in lower K_{sat} of soil and lower sorptivity values of soil aggregates (Ferrero et al., 2007). The increased contribution of finer pores, combined with the increased number of contact points between soil particles, leads to greater internal aggregate strength (Horn et al., 1994; Ferrero et al., 2007) and lower wettability (Goebel et al., 2004). The hydraulic properties of soil aggregates play an important role under field conditions where large inter-aggregate pores are drained earliest and the water flow is then influenced by the properties of the aggregates themselves (Horn and Smucker, 2005).

Tractor passes can form an anisotropic soil pore system due to simultaneous movement of aggregates or particles forward and downwards and to wheel slippage (Pagliai et al., 2003; Peng and Horn, 2008). The changes can form a platy structure in the upper few centimetres with elongated pores that are oriented parallel to the soil surface. These pores are not vertically continuous and induce mainly horizontal fluxes of water and gas (Horn et al., 2003; Pagliai et al., 2003; Dörner and Horn, 2006, 2009). In a recent study by Dörner and Horn (2009) found that transport properties such as hydraulic conductivity and air permeability show a more defined anisotropy than mechanical properties. The tendency for horizontal water and air fluxes (and hence solute- and particle-bound nutrient transport) due to the development of a platy structure was also shown on grazed soil by the anisotropy of the K_{sat} (Krümmelbein et al., 2006; Reszkowska et al., 2011) and air permeability (Reszkowska et al., 2011).

The results of Gallipoli et al. (2003) showed a significant increase in the degree of saturation during shearing and attributed this change to the combination of an inflow of water to the sample and a decrease in the total void volume as the sample compressed during shearing. It was suggested that the increase in the degree of saturation was assumed to result from water-filling of additional voids because of a decrease in both the void sizes due to compression and the number of voids below a critical size corresponding to water-saturation at the applied suction. The authors argued that in such a soil with bimodal pore size distribution, the increase in the degree of saturation was accentuated by dilation of small water-filled micro-voids within relatively densely packed soil, despite the compression of the more open macrostructure. This phenomenon means that deformation does not always result in an increase in bulk density.

3.5. Unsaturated hydraulic conductivity

Experimental data relating the effects of soil compaction on unsaturated flow are limited. Some studies have shown that soil compaction results in greater unsaturated hydraulic conductivity (K) of soil for the whole range of matric potential up to -100 MPa (Assouline et al., 1997; Richard et al., 2001) and for soil aggregates (Lipiec et al., 2009). This increase was ascribed to increased connectivity between smaller pores filled with water in a compacted soil than in a loose one (De Cockborne et al., 1988; Guérif et al., 2001; Lipiec and Hatano, 2003; Lipiec et al., 2009; Richard et al., 2001). Using microtensiometers, Türk et al. (1991) found that greater water uptake rates in the denser aggregates can also be caused by larger differences in matric potential between inner and outer parts (i.e., increased hydraulic gradient). As shown by Richard et al. (2001), for water potentials less than -15 kPa, K in the compacted layer was higher than that in an uncompacted layer. The ratio between K in the two layers was 2.1 at -80 kPa. This observation can be explained by the fact that the reduction in the pore space between aggregates and the deformation of the aggregated soil increases the contact surface area between the aggregates (Gupta et al., 1989). Thus, K would be increased by compaction when water is mainly within the aggregates. On the contrary, Zhang et al. (2006) reported that K was not affected because there were no significant differences among treatments at pore sizes <60 μm for one site (Heyang) and at pore sizes <15 μm for a second site (Mizhi). They explained that the water volume ratio corresponding to their K measurements was beyond the range where compaction had significant effects on pore sizes and spaces.

3.6. Infiltration capacity

Infiltration capacity is the maximum rate at which water soaks into or is absorbed by the soil through the soil surface (Ward and Robinson, 1990). In a hydrological context, many studies have dealt with the effect of soil compaction on soil hydrologic properties (e.g. Alakukku, 1996; Assouline et al., 1997; Betz et al., 1998); however, few investigations have been concerned with the effect of soil compaction on infiltration (van Dijck and van Asch, 2002; Håkansson, 2005; Kim et al., 2010). van Dijck and van Asch (2002) demonstrated that compacted subsoil under vineyards and orchards reduces neither steady-state infiltration rates nor total infiltration, despite the high infiltration rate of 85 and 105 mm h^{-1} for a maximum ground contact pressure by tractors of around 70 kPa. This fact was confirmed by measurements of K_{sat} that were only 1–3 times lower in subsoil than in topsoil. Possibly, larger ratios between the K_{sat} of top- and subsoil than those observed by the authors are required for the development of a saturated zone on top of the compacted subsoil layer. Quantitative analysis of infiltration under different saturation levels may offer more appropriate information on the effects of compaction on infiltrability. From this perspective, hydrodynamic variation of water content in response to brief irrigations at different depths was proposed to investigate the effect of soil compaction on infiltration capacity (Alaoui and Helbling, 2006). Delaying infiltration and enhancement of capillary rise to the soil surface due to compaction leads to higher soil water potential (greater wetness) that can be detrimental to trafficability and to workability (Müller et al., 1990; Boone and Veen, 1994). In comparison, under a pressure of 320 kPa and an infiltration rate of 24 mm h^{-1}, soil compaction significantly reduced the infiltration rate to nearly 0 mm h^{-1} in the topsoil of a trafficked plot (Alaoui and Helbling, 2006). Indeed, during the infiltration experiments, water infiltrated the soil on the control plot directly and instantaneously, whereas it ponded on the surface of the compacted soil. In the latter, variation of the water content was observed neither at depth during infiltration or during drainage, showing the damage of both functions—water transport and aeration. These observations are supported by the results obtained by Kulli et al. (2003) and Gysi et al. (1999) for similar soil and by Kim et al. (2010) for claypan soils due to the wet conditions created by the subsurface claypan horizon. By using dye-tracer experiments, Kulli et al. (2003) demonstrated that only very few flow paths are continuing below the topsoil directly underneath the area of the passage of the wheel. Mosaddeghi et al. (2007) reported that air permeability is a useful index of soil physical quality, especially at low matric suctions when largely dependent on connectivity of air-filled pores. The above results show the usefulness of investigating infiltration capacity over a vertical soil profile to evaluate the connectivity and tortuosity of the pore system with respect to compaction.

The infiltration process can be investigated by simulating soil moisture variation in response to a brief irrigation event using the kinematic wave approach (e.g., Germann and DiPietro, 1996; Alaoui et al., 2003). Semi-quantitative analysis of the infiltration and aeration capacities (Fig. 1) may be carried out from the evaluation of both $\Delta\theta_i$ (=$\theta_{max} - \theta_{init}$), which shows the magnitude of the increase of mobile soil moisture during infiltration, and $\Delta\theta_d$ (=$\theta_{max} - \theta_{end}$), which shows the magnitude of the decrease of mobile soil moisture during drainage. The values of $\Delta\theta_i$ and $\Delta\theta_d$ provide information on rapid increase and decrease in water content, respectively.

Preferential flow may occur for a short time period, varying from about 10 min to about 20 h, depending on soil structure and soil texture (e.g., Alaoui et al., 2003; Germann and DiPietro, 1996). The delimitation between the micropore and the macropore domain can be determined by dual-porosity modelling. Resulting WRC and the partitioning between soil moisture contributing to rapid flow (and preferential flow if it bypasses volume fractions of the porous matrix) and the soil moisture contributing to matrix flow are schematically presented in Fig. 1. Here, the water content serving as a reference can be obtained by a compression test using a standard oedometer and is a minimum value corresponding to the critical void ratio (Alaoui and Helbling, 2006). These authors found a reference value of 20% for a loamy soil.

4. Modelling water retention and water dynamic in compacted soil

4.1. Hydromechanical models

Hydromechanical models are models that link pore space geometry in a deformable porous media to its hydraulic properties

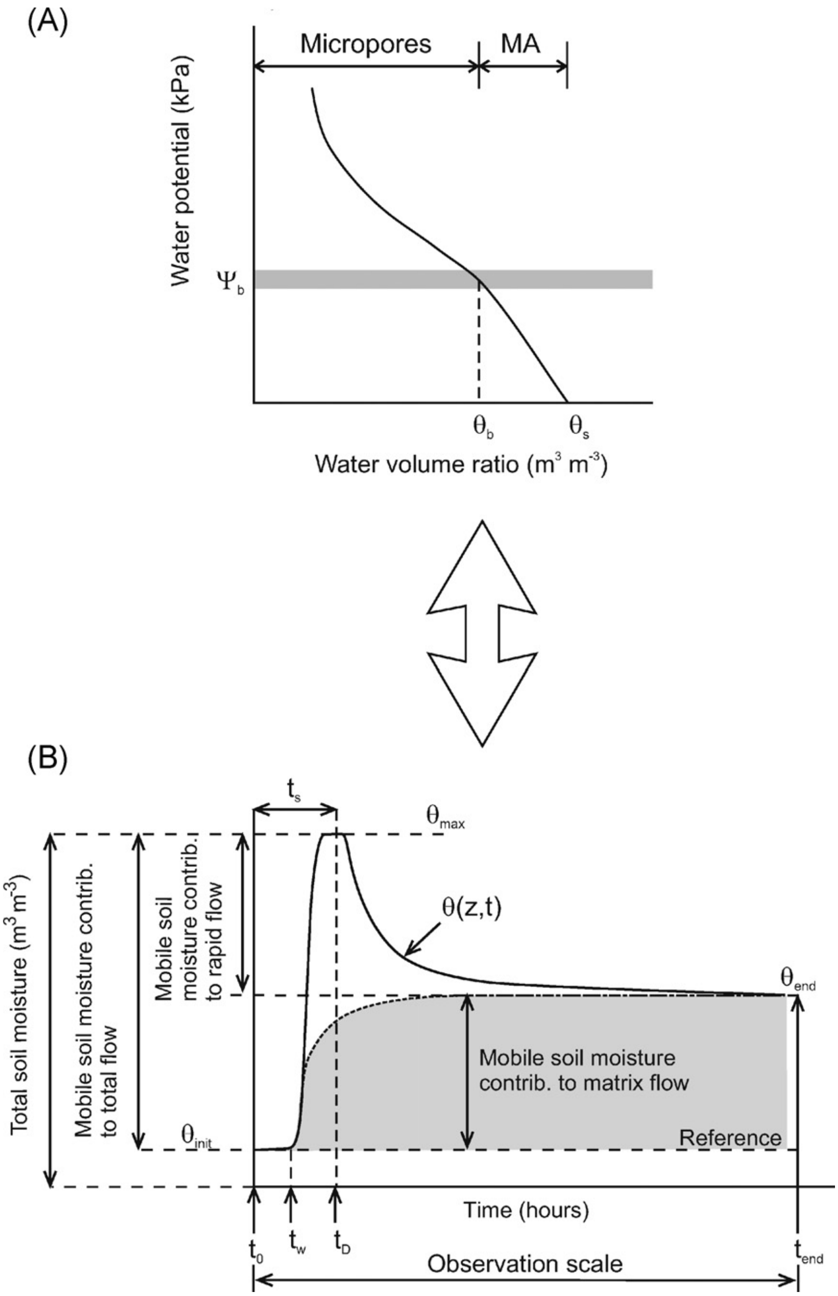

Fig. 1. (A) Water retention curve in structured soil; θ_b is the boundary water content between micropores and macropores; Ψ_b is the boundary water potential between micropores and macropores; θ_s is saturated water content. (B) Evolution of total soil water content $\theta(z,t)$ measured at depth z and time t during infiltration and drainage; t_o is the beginning of infiltration; t_{end} is the cessation of drainage; t_W is the arrival of the wetting front; t_s is the duration of infiltration; t_D is the arrival of drainage front; θ_{init} is the initial soil water content prior to infiltration; θ_{max} is the maximal soil water content measured during infiltration; θ_{end} is the final soil water content ending rapid flow corresponding to t_{end}. The reference is defined by the oedometer compression test (see text for explanation).

(i.e. water retention and hydraulic conductivity) and fluid (i.e. water) flow over various space and time scales. The pore space geometry depends on the arrangement of soil particles and aggregates. Thus, organisation of soil particles is a dominant influence on the hydraulic properties of unsaturated soils (Nimmo, 1997). Eggers et al. (2007) showed that pores within a cubic pack of aggregates deform isotropically even under anisotropic stress conditions.

Or (1996) proposed a model of liquid phase sintering of glass compacts to describe rapid wetting-induced densification of aggregates, assuming that (i) reduction in soil strength upon wetting is similar to the effect of high temperature on glass powders (Scherer, 1977, 1984), and (ii) capillary forces induce viscous flow. This model reproduced the soil structural changes of aggregated silt loam during wetting-drying cycles quite well by using the viscosity of soil as a fitting parameter. It overcomes many of the drawbacks related to other established geotechnical approaches (Sigala, 1968; Keller, 1970; Ghavami et al., 1974) which do not address either the question of how to adjust empirically for contact area between aggregates or the changes in pore size distribution which greatly affect soil hydraulic properties. Ghezzehei and Or (2000) extended Or's aggregate coalescence model (Or, 1996) by accommodating arbitrary wetting-drying conditions and by deriving constitutive viscous flow relationships from first principle. This was achieved by integrating energy source considerations as well as geometrical configurations. The non-linearity of wet soil viscous flow behaviour was also accounted for by introducing a Bingham rheological model. Similarly, Sisavath et al. (2000) presented analytical approximations based on the hydraulic radius concept (Dullien, 1979) and other approximations attributed to de Saint-Venant (1879) and Aissen (1951) to estimate pore hydraulic conductance from pore geometrical properties.

Starting from the idea that the contacts between aggregates are highly conductive close to saturation and become bottle-necks as water potential decreases, Carminati et al. (2008) showed that the hydraulic conductivity of a pair of aggregates, $K_{\mathrm{pair}(h)}$, with matric head h of the aggregate soil matrix, depends greatly on the contacts between the aggregates. Combining this concept with an analytical model of contact mechanics, Berli et al. (2008) predicted the hydraulic conductivity of a deforming pack of aggregates.

In order to extend the above analyses to more complex pore systems, there is a need to link these processes with easily observable and well-established sample scale deformations and flow behaviours. Eggers et al. (2007) proposed an analytical and finite element model to describe fluid permeability through a stack of deforming spherical aggregates. Tuller and Or (2001) determined the cross-sectional conductivity of liquid films by solving the Navier–Stokes equation in the two-dimensional cross-section of different liquid configurations, assuming a linear relationship between flux density and hydraulic gradient. Carminati et al. (2008) introduced the roughness of the contact aggregates in the calculation of hydraulic conductivity assuming a mean feature of aggregate and contact cross-sections. It was shown that the roughness of the contact aggregates is not only crucial for the stability of unsaturated granular media (Hasley and Levine, 1998) but also for hydraulic behaviour of soil resulting in rapid drainage of the contact region (Carminati et al., 2008). The model proposed by Carminati et al. (2008) is limited to unconsolidated aggregated media such as freshly tilled soils.

Or et al. (2006) developed a simple analytical model for a soil unit cell that links deformation with hydraulic conductivity of unsaturated soil on the basis of the findings by Eggers et al. (2006, 2007). Berli et al. (2008) proposed a hydromechanical model for change in aggregate contacts and their impact on inter-aggregate and intra-aggregate fluid flow on the basis of the investigations by Carminati (2007). Results showed good agreement between measured and calculated K for hydraulic head h <-1 m, whereas this model fails in describing K close to saturation (>-1 m). This discrepancy can be attributed to the simplification of assumptions about the macropore necks that are reduced to a single diameter and the neglected impact of the capillary bridges which greatly contribute to the flow across the inter-aggregate contact near saturation (e.g., Carminati et al., 2008).

Despite the noticeable progress of these models in depicting hydraulic conductivity measurements relatively well, some questions are still open; for example, how to conceptualize (i) the transition from the scale of individual aggregates to up-scaled pore continuum of dual-porosity models, and (ii) the changes in the flux between the domains (in both directions), depending on soil texture and loading.

There is now scope to account for both hydromechanical and thermal conditions in deformable soils. Recently, Salager et al. (2010) proposed a predictive relation expressing the change in soil suction with water content, temperature and density as described by void ratio. This relationship predicts the water retention curve at a specific temperature from the knowledge of the water retention curve at reference temperature. Hence, the number of experimental tests required to characterize the thermo-hydraulic behaviour of a given soil is greatly reduced.

4.2. Water retention curve

The WRC describes the relationship between two fundamental state variables of soil water: the matrix pressure head, including capillary head (ψ), and the water content (θ). This relationship describes flow processes over a large range of capillary potential. Approaches to model the effect of the increase in the soil bulk density on the WRC are very limited (Assouline, 2006a). Rajkai et al. (1996) used a pedotransfer function to predict the WRC. They obtained improved results by incorporating fitted cumulative particle-size data and bulk density. Inserting a penetration resistance into the pedotransfer functions improved the estimates of the WRC, based on soil texture and bulk density. Additional more conceptual approaches for estimating the effects of the changes in bulk density on the WRC involve the formation of dense seal layers at the surface of a bare soil exposed to high-energy rainfall (Baumhardt et al., 1990; Mualem and Assouline, 1989). The approach of Baumhardt et al. (1990) is based on three main steps: (i) experimentally estimate the K_{sat} of the seal layer; (ii) compute the seal saturated water content on the basis of the K_{sat} estimate using the Kozeny–Carman relationship, and the water entry value using the Poiseuille equation, and iii) determine the Brooks and Corey (1964) exponent λ for the seal layer, assuming a proportional increase in the initial water content of the seal with bulk density. Although these approaches capture the changes in the WRC caused by compaction, they do neither account for the effects of soil structure on the pedotransfer function estimates, nor for changes in soil pore distribution caused by compaction (Assouline, 2006a). The last author used two expressions of the WRC, relating their parameters to the bulk density of a compacted soil. Data were calibrated and validated against experimental WRC data of soils at various levels of compaction. These relationships gave a relatively good prediction of the effect of bulk density on the WRC.

Cornelis et al. (2001) reported that the pedotransfer functions predict moisture content well near saturation and permanent wilting point. The former is mainly dependent on total porosity and the latter on bulk density and clay content. The highest prediction errors were at moisture conditions close to field capacity, due to the mostly different morphology of pore volume. It was shown that use of neutral networks (NNs) to estimate soil water retention performed slightly better than the regression-based PTFs (Pachepsky et al., 1996; Koekkoec and Booltink, 1999). The performance of both NNs and regressions was comparable when van Genuchten's (van Genuchten, 1980) equation was fitted to data for each sample, and the parameters of this equation were obtained from texture and bulk density. Results showed that the changes in volumetric water contents at given potentials affect the hydraulic conductivity.

4.3. Hydraulic conductivity function

The experimental evidence suggests that K_{sat} is mostly determined by large pores, which are greatly reduced when the soil bulk density increases (Carter, 1990; Lipiec et al., 1998; Håkansson and Lipiec, 2000; Assouline, 2006b). Consequently, drastic reductions in K_{sat} with increasing bulk density have been reported (e.g., Håkansson and Medvedev, 1995). Therefore, soil compaction effects can also be simulated directly by considering bulk density or porosity (Walczak et al., 1997), or indirectly by changing inputs for K_{sat}, K, soil water retention, penetrometer resistance, root depth and root distribution (Rajkai et al., 1997; Warner et al., 1997; Eckersten et al., 1998). When based on the bulk density or porosity, the methods above provide useful information on the state of compactness and do not relate soil structure to water movement. Lipiec et al. (2003) reported that bulk density was not a sensitive indicator for the estimation of K_{sat} as it contains no information regarding pore geometry and continuity.

Compaction has a great influence on macropore flow, but there have been few attempts to model these effects. In this respect, Dexter and Richard (2009) developed a model for water retention in bi-modal soils (soils having structural and matrix porosity only) into a model for tri-modal soils, including the effect of the macropores. This model is based on the exponential (Bolzmann) water retention function. When combined with Marshall's pore model for hydraulic conductivity, it can predict the K_{sat} on n-modal soils. They reported that the number of structural pores – and not their size – decreases as soil becomes denser. This shows that the multi-modal nature of soil pore size distributions has a big influence on water retention and on K_{sat}. The correlation between coarse pores (>10 μm) and K_{sat} reported by Gebhardt et al. (2009) illustrated their crucial importance for water movement under saturated conditions. In fact, the Hagen–Poiseuille law expresses the dependence of water flux through a simple pore in terms of the fourth power of the pore radius. Using the dual-porosity MACRO model (Jarvis, 1994), Alaoui and Helbling (2006) showed the dominance of macropore flow for their compacted loamy soil, especially in the topsoil. The volume of macropores, representing only 0.23–2.00% of the total soil volume, transported approximately 74–100% of total water flow. The results show the very weak lateral exchange between macropores and micropores in the topsoil due to the compacted soil matrix, as shown by the dye tracer images.

Dexter (2004b) used S (slope of the soil water retention curve at its inflection point) as a measure of the micro-structural porosity of the soil, which can be also used as a "matching point" in the studies of K (Eq. (1)). Their results are promising because at the inflection point, the largest water-filled pores are also at the peak of pore size distribution and therefore dominate hydraulic conductivity. Their approach also introduces the possibility of examining the likely trends of $K(i)$ (K of soil at the inflection point) in terms of soil composition and bulk density using pedo-transfer functions. They combined Eq. (1) with the pedo-transfer functions for the parameters of van Genuchten (1980) and the water retention equation of Wösten et al. (1999). It was shown that increases in $K(i)$ with increasing bulk density were predicted for different soil textures (sand, loamy sand and sandy loam) at low values of bulk density. This is to be expected because compaction (i.e. increasing bulk density) reduces the number of conducting pores. Dexter (2004b) suggested the relationship:

$$\log K(i) = -A + B \log \left[\frac{S}{h_i}\right]^2, \qquad (1)$$

where h_i is the suction at the inflection point and A and B are coefficients, depending on soil type. The author showed that to improve the precision of the estimation of $K(i)$, the suction values allowing its calculation must be in the range of 0–5 hPa, since macro-structural features empty in this range, and that predictions of K using saturation as a matching point can be highly inaccurate (see for instance Jarvis, 2008). Consequently, bulk density is not a good indicator of soil structure because it is not directly correlated with K_{sat}, which depends greatly on soil macro-structural features such as cracks and biopores and especially on their connectivity (Dexter, 2004b).

4.4. Dual-porosity models

Flow and transport in structured porous media are frequently described using double- (or dual) porosity models that take into account two flow domains; the basic model concepts are reviewed elsewhere (Gerke, 2006).

Typically, dissipation of momentum dominates flow in well-structured soils at high moisture contents and at high infiltration rates, whereas diffusion of capillary potential dominates flow in homogeneous soils over a broader range of moisture contents and at low infiltration rates. Therefore, approaches to flow in structured soils can be divided in two major groups:

i) Approaches based on Richards' (Richards, 1931) equation for water flow in the matrix domain (i.e., diffusion of potential energy) in combination with other approximations for macropore flow. The two domains are considered independently of one another with some account for interactions, and dual-permeability models assume flow in both pore domains (e.g., Gerke and van Genuchten, 1993). In this regard, Jarvis (Larsbo and Jarvis, 2005) developed a physically based model, MACRO that simulates water and solute transport in macroporous soil. Water flow in micropores is calculated with Richards' (Richards, 1931) equation, while macropore flow is simulated using the kinematic wave theory (Beven and Germann, 1982; Alaoui et al., 2003).
ii) Approaches using the kinematic wave theory (Lighthill and Whitham, 1955) based on the dissipation of momentum (Germann and DiPietro, 1996). A power function relates the water flux to the mobile volumetric water content.

Matthews et al. (2010) using the Pore-Cor void network model including Euler beta distribution to describe the sizes of the narrow interconnections, known as throats, revealed a change from bimodal to unimodal throat size distributions on compaction, as well as a reduction in sizes overall.

4.5. Shrinkage characteristic curve

Swelling and shrinking clay soils change in volume with soil-water content changes. The macroporosity, and to a lesser extent the microporosity, of swelling and shrinking soils is affected by their shrinkage behaviour. The magnitude of the change in bulk volume is usually described by the soil shrinkage characteristic curve, which represents the specific volume change of soil relative to its water content (Haines, 1923; Stirk, 1954; Braudeau, 1988).

The methods of shrinkage measurements helped to develop shrinkage curve models with different parameter sets (Braudeau et al., 1999). Braudeau (1988) proposed a conceptual model of soil shrinkage on the basis of the model of Sposito and Giráldez (1976), where clay aggregates in the soil are assumed to shrink as clay paste, and where the slope of the normal shrinkage is related to aggregate fabric and aggregate stability. The model proposed by Braudeau (1987) contains five main zones: linear and curvilinear residual shrinkage, basic shrinkage, and linear and curvilinear structural shrinkage. The endpoints of these zones representing the points of transition are considered as characteristics of the shrinkage process. The linear zones are modelled by straight line equations, and the curvilinear zones by exponential (XP) (Braudeau, 1988) or polynomial (PL) (Giráldez et al., 1983; Tariq and Durnford, 1993) parametric equations. The parameters represent the coordinates of the endpoints of the zones.

Recently, Cornelis et al. (2006) proposed to test the performance of different models frequently used to assess the shrinkage characteristic curve. For describing the shrinkage characteristic curve, it was shown that the multi-equation models proposed by McGarry and Malafant (1987) and Braudeau et al. (1999) and the modified model of Chertkov (2000, 2003), as well as the Groenevelt and Grant (2001, 2002) simple equation, fit the measurements well but differed in complexity and the number of parameters.

In addition to describing the shrinkage processes, the shrinkage characteristic curve may describe the state of compaction. A slope of 0–0.2 $cm^3 g^{-1}$ means that soil has a nearly constant volume when it dries and is therefore rigid or compacted (Braudeau and Bruand, 1993; Boivin et al., 2006). When the slope is greater than 1 $cm^3 g^{-1}$, soil is very unstable (Boivin, 1990). A more recent study by Schäffer et al. (2008)

showed that trafficking and associated destabilization of soil structure increased the slopes of the shrinkage curve in the basic and structural shrinkage domains by about 30% and more than 150% after 10 passes, respectively. Gregory et al. (2010), using X-ray computed tomography scanning, found that compression from 50 to 200 kPa mainly affected the >30-μm pores, changing a dual porous system of inter- and intra-aggregate pores to one mainly dominated by intra-aggregate pores. Shear-deformed soils retained more water than the compressed soil and shrank more, so that they remained tension-saturated at low (negative) matric potentials. The authors developed a function to predict the soil saturation state as a function of matric potential and porosity.

It was also demonstrated that combining the shrinkage and desorption experiments to calculate the equivalent pore radii of drainage plasma-macropores provides useful information about the changes in pore diameters upon compaction (Boivin et al., 2006). Therefore, there is a need for further investigations to assess soil compaction and its variability in situ.

Several processes, such as wetting/drying, freezing/thawing, and tillage help to alleviate the effects of soil compaction (Håkansson, 2005). In fine-textured soils, freezing and drying cause crack formation and disintegration of large clods into smaller aggregates. Roots and microbial exudates help to stabilize crack and aggregate surfaces. Research showed that the effects of the freeze and thaw and drying and wetting cycles are more frequent and intense near the soil surface (e.g. Arvidsson and Håkansson, 1994; Halvorson et al., 2003) and therefore the persistence of compaction effects increases rapidly with depth and in regions without freezing. In a study by Taboada and Lavado (1993), the damaged soil macropores (mainly >60 μm) regenerated and aggregate stability recovered during the subsequent period of surface water ponding, when soil swelling increased macropores in the grazed area but not in the ungrazed area.

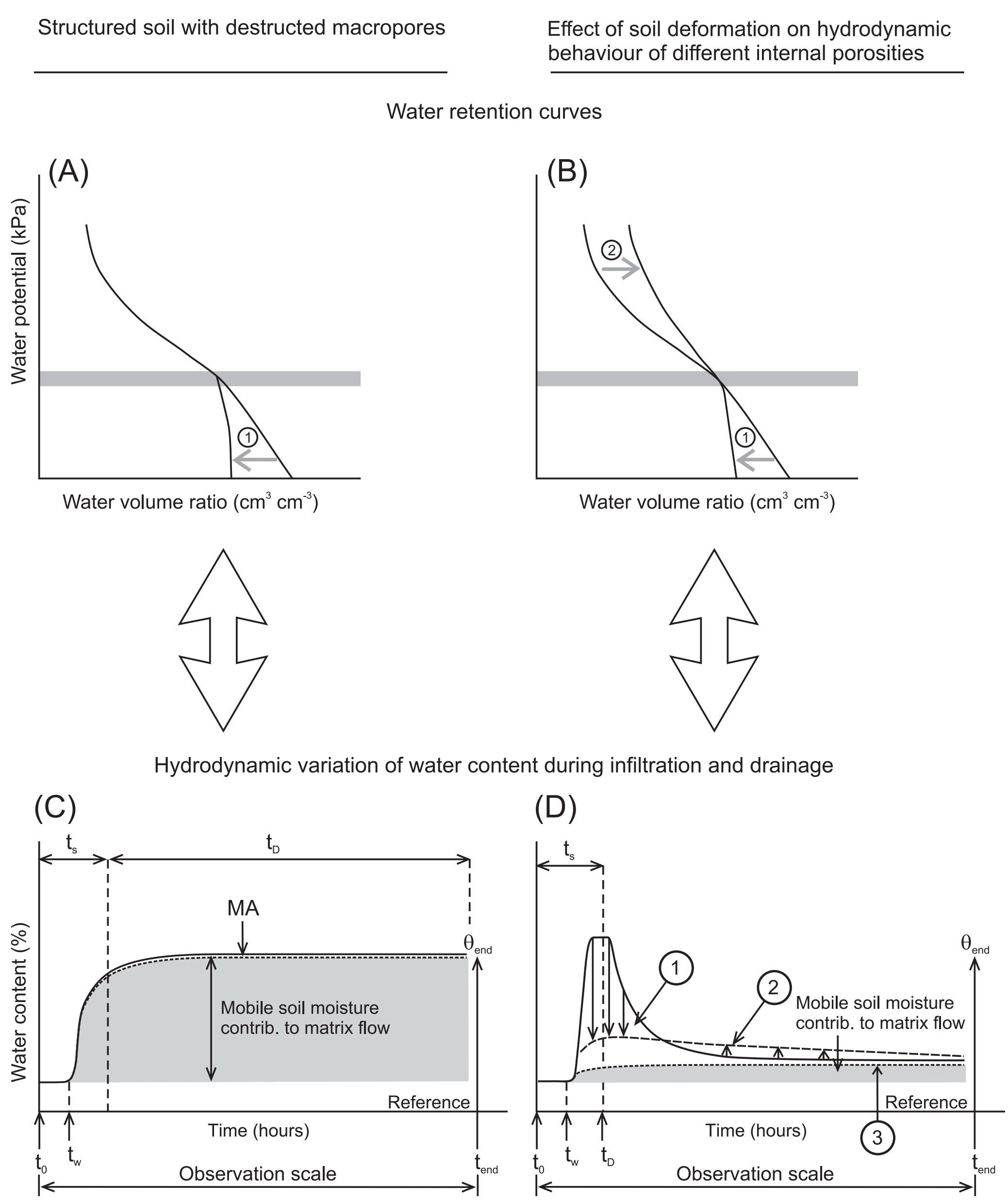

Fig. 2. Simplified outline of the effect of soil compaction on the soil structure from a hydrodynamic point of view; (A) and (B) water retention curves; (C) and (D) hydrodynamic variation of soil moisture; (1) reduction of structural porosity, (2) appearance of relict structural porosity, and (3) boundary between matrix and macropores; MA: macroporosity.

The action of the natural processes is related to tillage practices. Capowiez et al. (2009) reported that up to 30% of the compacted zones observed at the scale of soil profiles were colonized under reduced (RT) and only 10% under conventional tillage. A recent study by Weisskopf et al. (2010) indicates that omitting soil tillage after compaction results in the development of a dense, blocky structure, with the blocks separated by shrinking cracks and penetrated by earthworm holes. The authors observed that with the evolution of a network of shrinking cracks, especially stimulated under ley, air permeability values increased considerably, while macropore development still lagged behind, leading to a totally different ratio macropore volume: air permeability in the tilled vs. untilled soils. In coarse-textured soils without swelling and shrinking, processes are generally more persistent than in fine-textured soils.

5. A conceptual model

Results from the literature allow us to propose a general conceptual model where the implications of soil deformation on hydrodynamic behaviour can be synthesized. The basis of this conceptualization resulted from modelling water or solute transport in unsaturated soil. This implies at least assessing WRC measurements and hydraulic and physical parameters.

A heterogeneous structure in undisturbed (structured) soil generally creates suitable conditions for ecosystem services. Changes in the WRC and HCF affect the water transport processes. According to the dual-porosity modelling of flow and on the basis of WRC and the partitioning between the micropore and macropore domains, three main hypothetical schematic diagrams can be distinguished (Figs. 2 and 3):

i) Damage to macropores, illustrated by the decrease in the water potential >-20 kPa and domination of matrix flow as shown in Fig. 2A and C. This type of behaviour results from dynamic loading as observed by Pagliai et al. (2003), Servadio et al. (2001) and Zhang et al. (2006) among others. The matrix remains active and provides slow water transport. Soil hydraulic properties, such as saturated hydraulic conductivity and pore shape by image analysis techniques (e.g., Richard et al., 2001) and dye tracer and infiltration experiments (Kulli et al., 2003; Alaoui and Helbling, 2006), may provide a good illustration of the destruction of structural porosity.

ii) The effect of soil deformation shown in Fig. 2B and D reflects a more complex process, showing coupled hydraulic and mechanical aspects. A decrease in structural porosity results in an increase in relict structural. Such an effect is induced by dynamic loading of wheeled and tracked heavy machines. The volume of the relict structural pores is indicative of soil compaction and its effect on behavioural properties (e.g., Richard et al., 2001). From a hydrodynamic point of view, this effect leads to a change in the WRC and corresponding hydrodynamic variation in θ. Depending on the degree of compaction, a transition from macropore flow to intermediate or matrix flow occurs. Backscattered electron scanning images and mercury porosimetry highlighted this change, whereas bulk density and total porosity did not (Richard et al., 2001).

iii) The destruction of the matrix in the form of dense soil layer in the topsoil caused by trampling in the pastures (Fig. 3) (Alaoui and Helbling, 2006). In this case, vertically oriented macropores are resistant to vertical compression (e.g., Kirby and Blackwell, 1989; Langmaack et al., 1999; Lee and Foster, 1991) and transport all water downwards, providing rapid flow characterized by a drastic draining of macropores during drainage. Hydraulic properties such as saturated water conductivity

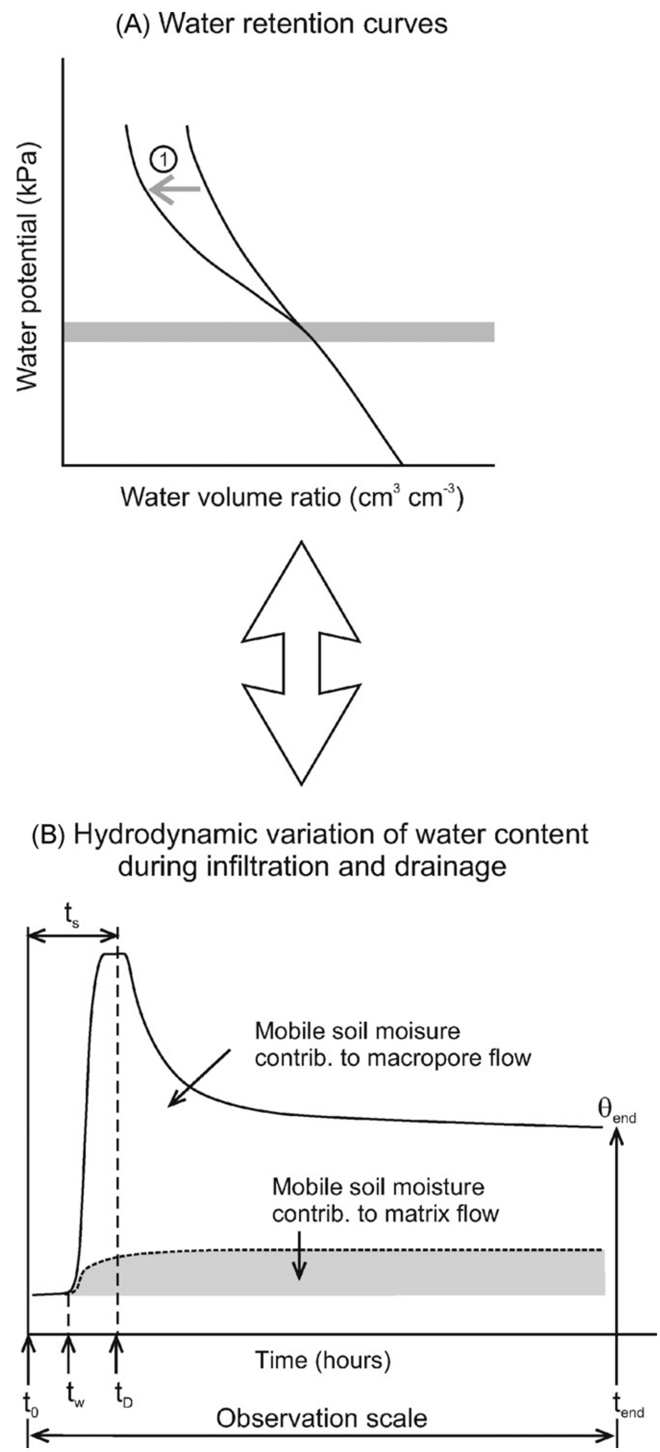

Fig. 3. Simplified outline of the effect of soil compaction on the soil structure (i.e., soil pasture); reduction in the matrix porosity; (A) water retention curve; (B) hydrodynamic variation of soil moisture.

(Zhang et al., 2006; Kim et al., 2010), pore volume distribution analyses (Alaoui and Goetz, 2008), dye tracer experiments (Kulli et al., 2003) and water flow dynamics during irrigation experiments describe such effects of pore structure deterioration (Alaoui and Helbling, 2006).

The concept outlined above may be verified by the following measurements:

(i) Direct measurements of water flow or (ii) modelling it on the basis of the key hydraulic and physical soil parameters (i.e., Ψ, K_{unsat} and $\theta(t)$). Fitting these variables with a dual-porosity model will help to divide total water content into two categories, mobile (contributing to preferential flow) and immobile (contributing to matrix flow). Modelled water flow in such a dual-porosity medium provides the level of compaction in each domain.

6. Conclusions

The aim of this paper is to review recent advances in the understanding and evaluation of functional changes in compacted soil with respect to water retention and water flow.

Among other processes, compaction, shearing and the rearrangement of soil aggregates and clods due to shrinkage can greatly affect the pore geometry of agricultural soils. These soil structural changes greatly affect soil water movement by altering the hydraulic properties that are commonly described by the soil water retention curve (WRC) and the unsaturated hydraulic conductivity function (HCF). These functions are particularly useful because they describe the effects of compaction on the functionality of soil structure, rather than only the state of compaction.

The sophistication of the methods developed recently (i.e. X-ray computed tomography) provides accurate information on the pore system. These methods make it possible to conceptualize aggregate contact size evolution and consequently quantify the effects of compression on hydraulic conductivity functions at the aggregate pore scale. However, it remains unclear how to up-scale the dynamic inter-aggregate pore models into the continuum-scale dual-porosity model: indeed, what are effective parameters to this end? In particular, what are effective hydraulic properties for the preferential flow domain and changes in the fractions and geometries during the dynamic structural changes due to external perturbations? The review strongly suggests extending the existing hydraulic and hydromechanical model to include the pore structural changes that account for the movement and rearrangement of soil aggregates and the resulting changes in the soil hydraulic properties that basically manifest the effects of shearing and compaction on water flow.

Based on modelling water flow, we also present a conceptual framework that summarizes the effects of external perturbations on soil. This framework may enable us to conceptualize compaction effects in a cost-effective manner, since it only requires WRC measurements. There remains the challenge, however, of how to link these data to the pore geometry and then to its evolution upon compaction.

Acknowledgment

This study was supported by the Swiss Federal Office for the Environment (FOEN), grant no. 04.1214.PJ/H343-0419.

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