

CHAPTER 7

Effects of Compaction on Soil Hydraulic Properties

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SUMMARY

Compactive processes affect soil hydraulic properties and associated soil water flow. Soil water retention and transport properties are altered in response to changes in pore space geometry. Soil water flow is affected not only by soil hydraulic properties but additionally by the distribution of sources and sinks of water in the soil system. Compaction can alter soil pore geometry, and can also affect sources and sinks of water by changing surface configuration, and crop rooting distribution. This paper reviews the literature and presents data and relationships showing the effects of compaction on soil hydraulic properties and water flow, presents numerically modeled water flow for some management systems, and identifies future directions for research.

INTRODUCTION

Compaction significantly influences soil hydraulic properties, infiltration, soil water retention, soil water flow, and hydrologic response (Klute, 1982; Onstad and Voorhees, 1987). Unfortunately, comparative soil hydraulic properties or soil hydrologic components for various management systems are often not consistent (Culley et al., 1987b). Given specific soil management operations, few general statements can be made to describe the effects of compaction on soil hydraulic properties and soil hydrology (Baker, 1987). Measurement and interpretation are often difficult because compaction and soil loosening action usually take place spatially within the same unit implement width. Zonal loosening often requires zonal compaction in a horizontal/vertical arrangement. Thus, inconsistencies are often attributed to variance in climate, topography and spatial/temporal variance of the soil, because the spatial character of tillage/traffic (Cassel, 1983) was not recognized. Measured tillage effects on soil hydraulic properties often disagree because investigations are not consistent as to how, when, and where in the soil

profile measurements were made. Results obtained in loosened and compacted zones may have been wrongly lumped together. A better understanding of the effects of management on soil hydraulic properties and hydrology is badly needed; it requires joint efforts in theory development, field measurement and modeling of systems with validation. Field traffic has a fundamental influence on soil hydraulic properties, although its adverse effect on soil permeability is not always as obvious as demonstrated in the accumulation of surface water in wheel ruts (Fig. 1).

The objectives of this paper are to: (1) review some fundamental principles of water retention and transmission; (2) highlight observations of soil hydraulic properties and hydrologic components for various compactive and tillage situations; (3) report results of modeling efforts for predicting the hydrology of soils; (4) suggest future research directions.

Water flow and associated water content of the soil have many indirect effects on plant rooting and growth, aeration, and nutrient availability. Our approach will be to concentrate on those hydraulic factors needed to predict water flow and associated water contents in a soil affected by compaction.



Fig. 1. Ponded water in wheel ruts illustrates the greatly reduced permeability of compacted soil (foreground) when heavy rain followed the mechanical harvesting of potatoes, whereas there was no ponded water in the unharvested area (background).

PORE SPACE, WATER RETENTION AND WATER TRANSPORT

Water retention and transport in soil via the non-solid or the pore spaces, underscores the importance of porosity (see Chapter 5). Total porosity, n , is defined as the ratio of non-solid volume to total volume, and can be calculated using the following equation:

$$n = 1 - (\rho_d/\rho_s) \quad (1)$$

where ρ_d = dry soil bulk density and ρ_s = density of soil solids. Bulk density and total porosity, which are affected by compactive and tillage operations, are commonly measured (Allmaras et al., 1977; Gantzer and Blake, 1978; Akram and Kemper, 1979; Bauder et al., 1981; Reicosky et al., 1981; Hill and Cruse, 1985; Potter et al., 1985; Voorhees et al., 1985; Culley et al., 1987b; Allmaras et al., 1988). Intra- and inter-aggregate porosity were determined in relation to tillage-induced soil structure (Allmaras et al., 1977); yet additional information is needed to estimate soil water retention and transport functions.

The size, shape, continuity and tortuosity of pores in structured soil all contribute to the water retention and transport characteristics (Hill et al., 1985). Therefore, total porosity alone should not be expected to correlate with either the water retention or hydraulic conductivity function (McBride et al., 1987; Kluitenberg et al., 1988).

The height of water rise in a capillary tube, h , is described by Jurin's equation:

$$h = (2 \gamma \cos \psi) / r \rho g \quad (2)$$

where γ = surface tension of water; ψ = contact angle between liquid and solid; r = pore radius; ρ = density of water; g = gravitational acceleration. The Jurin equation can be applied to soil by assuming that a given value of h is the matric potential at which all pores greater than the associated radius, r , must drain. Based on this analogy, soil water retention at a given matric potential is dependent upon soil pore size and geometry. The soil water retention curve is expressed as a continuous functional relationship between volumetric water content and negative pressure required to remove water from a soil (Hillel, 1980).

Numerous investigators have theoretically estimated a relative soil hydraulic conductivity, given a measured water retention function (Marshall, 1958; Millington and Quirk, 1961; Brutsaert, 1967; Green and Corey, 1971; Campbell, 1974; Mualem, 1976; Van Genuchten, 1980). A measured saturated or unsaturated hydraulic conductivity, K , is used to convert relative to absolute estimated values. Others have reported empirical equations using texture and bulk density to estimate soil water retention and hydraulic conductivity (Clapp and Hornberger, 1978; Gupta and Larson, 1979; Arya and Paris, 1981; Rawls and Brakensiek, 1982; Saxton et al., 1986). Wu et al. (1990) developed an empirical

technique to account for both particle size and aggregation effects on the water retention function. Unfortunately, these techniques need more evaluation for application to a wider range of soils, states of aggregation and compaction.

Van Genuchten (1980) presented the following equations to define mathematically the soil water retention relationship (eqn. (3)) and unsaturated hydraulic conductivity (eqn. (4)):

$$\theta = \theta_r + (\theta_s - \theta_r) \left[\frac{1}{1 + (\alpha h)^n} \right]^{(1-1/n)} \quad (3)$$

$$K(h) = K_s \frac{1 - (\alpha h)^{n-1} [1 + (\alpha h)^n]^{(1-n)/n^2}}{[1 + (\alpha h)^n]^{n-(1/2n)}} \quad (4)$$

where θ = actual water content; θ_s and θ_r = saturated and residual water content, respectively; $K(h)$ = unsaturated hydraulic conductivity; h = matric potential; K_s = saturated hydraulic conductivity; α and n = parameters describing the shape of the soil water retention function. The parameter n is closely related to pore size distribution (Horton et al., 1987). This mathematical form (Van Genuchten, 1980) is the most general formulation available; others are restrictive in that the fit is best at some special range of the water retention function.

Often the water retention function and/or the hydraulic conductivity function (of water content) are not available for a test soil over the whole water content range. Wösten and Van Genuchten (1988) suggested that simultaneous fitting of the functions in eqns. (3) and (4), using the available incomplete data, can provide good estimates of the water retention and conductivity functions for simulations of the water flux.

SOIL WATER DIFFUSIVITY RESPONSE TO BULK DENSITY

Libardi et al. (1982) estimated soil water diffusivity, D , of various soils for a range of bulk density values, provided D is known for at least one bulk density. The following equation was used to represent D ($\text{cm}^2 \text{s}^{-1}$) for soil i :

$$D_i(w, \rho_d) = A m_i^2 \exp(Bw) \quad (5)$$

where w = a dimensionless water content equal to $(\theta - \theta_o)/(\theta_s - \theta_o)$; ρ_d = dry bulk density; A and B = constants studied in detail by Brutsaert (1979); m_i ($\text{cm s}^{-0.5}$) = the slope in a plot of distance (cm) from the water source to the wetting front in horizontal infiltration (Bruce and Klute, 1956), as a function of the square root

of the time (s). The volumetric water content, θ , has a value θ_0 for the initial air-dry condition and a value θ_s near the water source. The following equation can be used to relate m_i and ρ_d :

$$m_i = a_i + c_i \rho_d \quad (6)$$

where a_i and c_i = empirical coefficients. Rearranging eqn. (6) and substituting c as the average value of c_i gives:

$$m_i - a_i = c \rho_d \quad (7)$$

where c (= -0.464) and a_i are values which translate the observations to a common ordinate. Libardi et al. (1982) presented the combined results obtained from 13 soil types ranging in textural classification from sand to clay.

If m_i is obtained for one soil at one bulk density value, a_i can be estimated using eqn. (6) and the value of c from eqn. (7). Hence, eqn. (5), written as a function of w , becomes:

$$D_i(w, \rho_d) = A (a_i - 0.464\rho_d)^2 \exp(Bw) \quad (8)$$

The derivative of eqn. (8) with respect to ρ_d shows how changes in bulk density affect soil water diffusivity:

$$(\partial D_i / \partial \rho_d) = 0.928 A (a_i - 0.464\rho_d) \exp(Bw) \quad (9)$$

FIELD AND LABORATORY MEASUREMENT OF HYDRAULIC PROPERTIES

Field and laboratory techniques for measurement of the unsaturated hydraulic properties of soil are described, respectively, by Green et al. (1986) and Klute and Dirksen (1986). The solution of unsaturated flow problems generally has required experimental determination of the relationship between hydraulic conductivity and water potential or water content. Field methods used to obtain these relationships include the instantaneous profile method, steady-flux methods (with sprinkler irrigation or artificial crusts), sorptivity measurements, and use of tension infiltrometers (Clothier and White, 1981; Ankeny et al., 1988; Elrick et al., 1988; White and Perroux, 1987, 1989). Soil profile and steady-flux techniques often require installation of tensiometers or neutron probe access tubes, which may limit sample numbers. Internal drainage rates in the subsoil often limit the range of applicable θ . When rapid field techniques and straightforward calculations are needed for measuring unsaturated hydraulic properties of the soil, especially at a number of sites and soil depths, tension and positive head infiltrometers have proven successful. They are especially useful for

comparing soil management treatments.

Saturated and near-saturated hydraulic properties, especially those in the tilled layer, are of particular interest but are difficult to predict without *in-situ* measurements. Spatial variability encountered in compaction studies due to both intrinsic soil properties and management effects (e.g., compaction or tillage), often necessitates intensive sampling to reach experimental objectives. Description of field-scale water or solute movement also requires that the distribution of hydraulic properties be known. Increasing interest in near-saturated hydraulic properties has prompted improved methods of measuring infiltration under negative water potential (Elrick et al., 1988; Ankeny et al., 1988; Perroux and White, 1988). An experimental approach to quantify compaction effects on soil hydraulic properties is to measure steady-state unconfined infiltration rates with ponded and tension infiltrometers (Ankeny et al., 1990b).

Unconfined, saturated infiltration rates are measured by ponding water in a ring pressed a short distance into the soil. The sharpened ring defines the infiltration surface area and prevents lateral surface flow of ponded water. After a steady-state saturated rate is measured, sand is applied to the infiltration surface to establish hydraulic continuity between the tension infiltrometer and the soil surface. Both saturated and unsaturated water flow are three-dimensional and usually reach steady-state rates rapidly (typically within 30 min). Dense and dry soils require the most time to reach a steady-state rate. At the exact location of the infiltration measurements, soil cores for laboratory measurements of desorption or hydraulic conductivity may conveniently be taken afterwards because the soil water content is then much higher. Confined laboratory measurements of unsaturated hydraulic conductivity can be made by a method of steady-state head control using a device suggested by Klute and Dirksen (1986) or as modified by Ankeny et al. (1991). Tension infiltrometers with a larger contact area (Perroux and White, 1988) may be used with a membrane in direct contact with the soil.

Steady-state infiltration rates can be used directly to compare compaction treatments (Ankeny et al., 1990b). In turn, these rates can be used to estimate saturated and unsaturated hydraulic conductivities (Ankeny et al., 1991). There are several practical advantages in estimating conductivities from unconfined measurements: (1) the estimates are independent of antecedent water potential or content; (2) only steady-state infiltration rate measurements are needed; (3) both ponded and tension measurements are taken on the same soil surface area; (4) flow through longer macropores is not interrupted by driving a ring or isolating a monolith; (5) calculations are straightforward. Capillary lengths as well as hydraulic conductivities can be calculated (White and Sully, 1987; Ankeny et al., 1991).

Available instrumentation also allows fast, accurate field determination of sorptivity. For example, an automated tension infiltrometer (Ankeny et al., 1988)

can be used in the field or laboratory to measure sorptivity, a measurement of water uptake by soil in the absence of gravitational forces (Philip, 1957), at different surface-applied water potentials. Improved precision of a transducer-equipped tension infiltrometer allows measurement of sorptivity even at low infiltration rates. An alternative method is to use two tension infiltrometers that have a large difference in contact area (White and Sully, 1987; Sauer et al., 1990). Macropore contributions to sorptivity can also be determined using a tension infiltrometer. Sorptivity measurements can be used to model the effect of antecedent water content on soil erosion, and water flux into soil peds from cracks and other macropores in a clay soil.

COMPACTIVE EFFECTS ON SOIL HYDRAULIC PROPERTIES

The soil compaction process increases soil bulk density and decreases total pore space; as a result, water-related soil properties are significantly altered. Another effect of loading on a soil is shear without a significant change in volume. This type of soil strain, associated with traffic and implement use on wet soils, most significantly alters water-related properties. According to Koolen and Kuipers (1983), compaction and/or shear may be produced by traction, transport, and depth control devices on machinery (tires, wheels, tracks, and sliding plates). There are therefore many possibilities for changing water relations in soils as related to tillage and traffic. The number of soil water-related properties chosen for measurement should be sufficient to simulate the water flux process; a single water-related characteristic rarely suffices to explain the full impact of compaction on water relations of an arable soil.

Measured water retention curves and predicted unsaturated hydraulic conductivity for Barnes loam packed at several bulk densities, demonstrated that compaction decreases total porosity (Reicosky et al., 1981). However, unsaturated water contents were larger for a wide range of matric potentials in compacted versus non-compacted soil (Fig. 2). Gupta et al. (1989) have shown a similar response to soil compaction when volumetric water content is related to water potential. Moreover, their relationship was based upon model computation. When the gravimetric water content is related to water potential, the effect of compaction is characteristically related to soil texture (Gupta et al., 1989). The original work of Hill and Sumner (1967) explains that the characteristic influence of compaction on the water retention curve is based on three combined effects: (1) prominence of large pores; (2) distribution of smaller pores; (3) overall reduction in pore volume caused by the compaction. Hill et al. (1985) characterized and explained water retention curves related to tillage systems without traffic-induced compaction, but this technique can also be used to describe compaction effects on soil water retention.

Hydraulic conductivity as a function of soil water content generally decreases with compaction (Fig. 2); however, over some of the compactive range, hydraulic

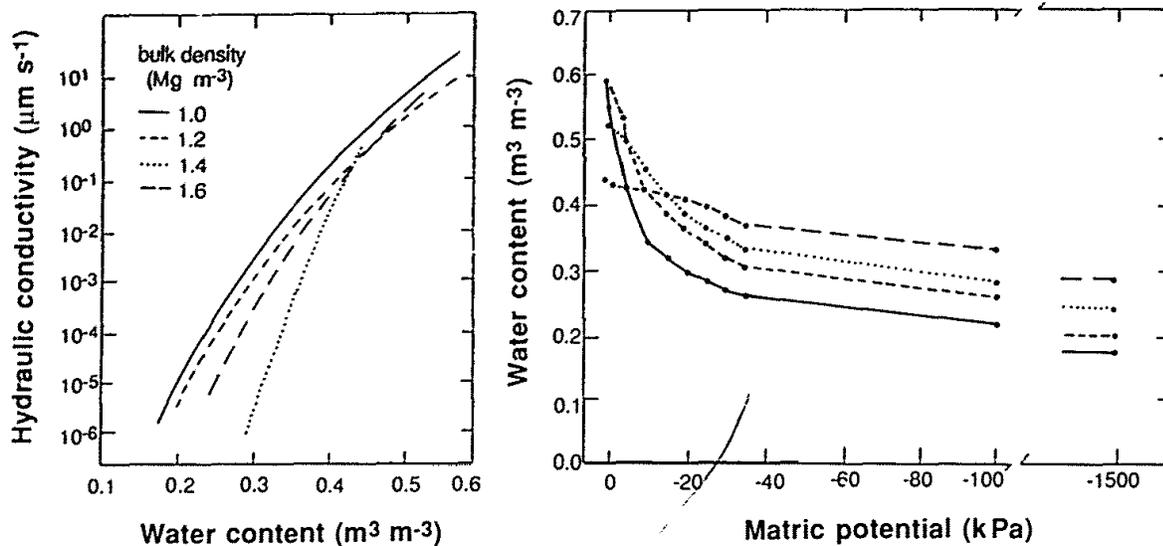


Fig. 2. Unsaturated hydraulic conductivity and soil water characteristics for Barnes loam packed to known bulk densities (from Reicosky et al., 1981).

conductivity as a function of matric potential may increase with compaction (Mapa et al., 1986; Horton et al., 1989). Saturated hydraulic conductivity is especially sensitive to soil compaction caused by field traffic (Table 1). Dawidowski and Lerink (1990) showed that stress produced during traffic has a relatively major influence on saturated hydraulic conductivity compared to the influence on total pore space, while soil water content at the beginning of compaction had a relatively stronger influence on total pore space than on hydraulic conductivity (Fig. 3).

Changes of these two hydraulic properties during uniaxial compression in undisturbed cores were sensitive to both applied stress and soil water content at the time of compaction. The LGP and HGP curves describe the soil response to field traffic at planting (and during traffic for harvest of root crops) using low- and high-pressure tractor tires, respectively (Fig. 3). Differences between uniaxial stress and field traffic were related to soil structure. In a triaxial experiment, where shear was involved during deformation with very small volume changes, the saturated hydraulic conductivity was reduced from 2.6 to $0.06 \mu\text{m s}^{-1}$ with an axial deformation of only 10% (Dawidowski and Koolen, 1987).

A maximum bulk density of laboratory-compacted soil samples generally occurred when soils were compacted at water contents near field capacity (Akram and Kemper, 1979). Soils containing a wide texture range were compacted with loads equivalent to 340 kPa. The soil compacted at field capacity had infiltration rates $<1\%$ of those compacted at air-dry water contents. Dawidowski and Lerink (1990) and Lerink (1990) made similar comparisons. Walker and Chong (1986) reported that sorptivity responded to soil compaction in a manner similar to void ratio (Fig. 4). Sorptivity was more sensitive to changes in soil structure than in

TABLE 1

Summary of means and coefficients of variation for hydraulic conductivities of a Webster silty clay loam after different treatments, measured near the soil surface (after M.D. Ankeny, 1990, unpublished)

Tillage	Position	Traffic	Matric potential (-kPa)	Hydraulic conductivity ($\mu\text{m s}^{-1}$)	CV ¹ (%)	
No-till	Interrow	No	0.00	166.4	52	
			0.30	8.4	42	
			0.60	3.0	57	
			1.50	1.1	46	
		Yes		0.00	26.6	96
				0.30	2.7	116
				0.60	0.9	83
				1.50	0.3	4
	In-row	No		0.00	257.4	70
				0.30	8.8	50
				0.60	3.5	60
				1.50	1.1	44
Chisel	Interrow	No	0.00	219.3	44	
			0.30	28.5	42	
			0.60	11.6	42	
			1.50	2.0	57	
		Yes		0.00	33.7	98
				0.30	2.8	116
				0.60	1.1	147
				1.50	0.4	116
	In-row	No		0.00	168.0	88
				0.30	13.2	68
				0.60	5.0	69
				1.50	1.9	67
Plow	In-row	No	0.00	593.2	84	
			0.30	12.7	33	
			0.60	4.5	27	
			1.50	1.2	46	

¹n = 16 for trafficked and non-trafficked; n = 8 for in-row measurements.

soil water content. Mapa et al. (1986) report sorptivity to be sensitive to temporal changes in soil hydraulic properties.

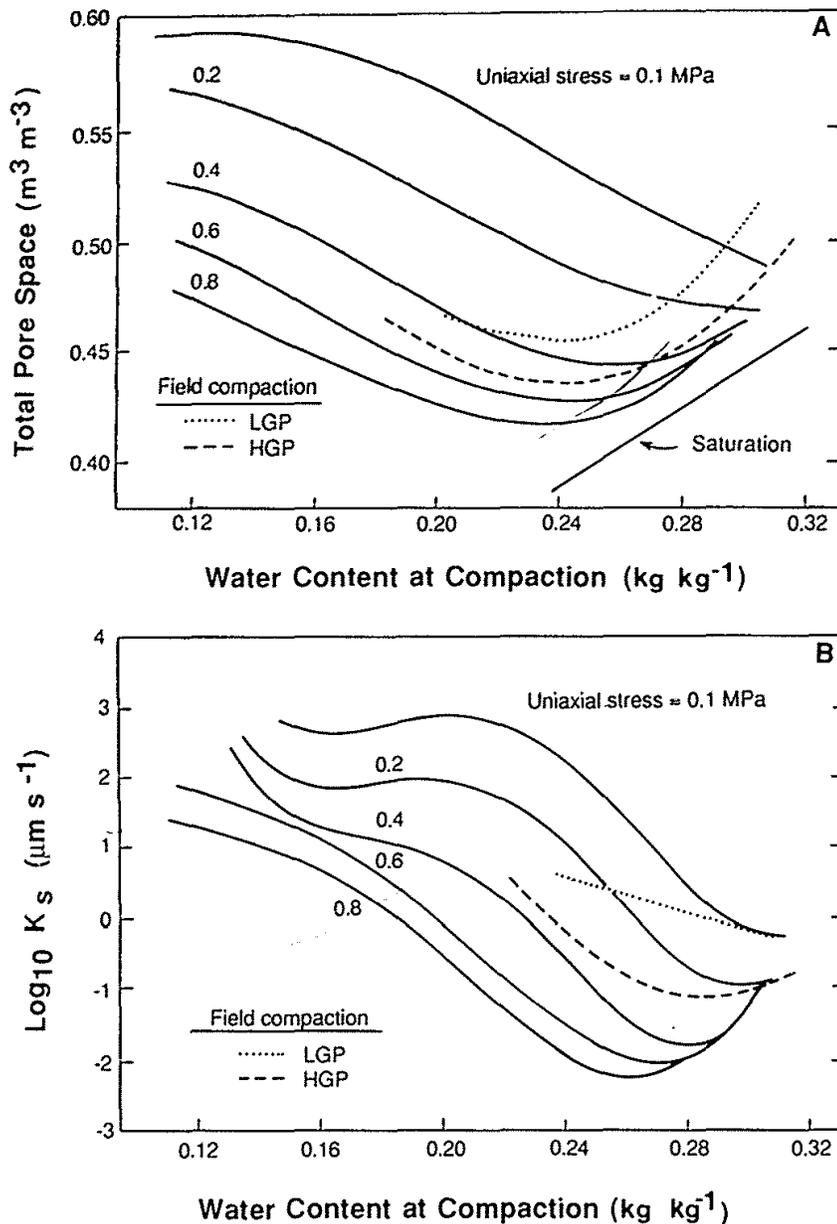


Fig. 3. Changes in total pore space (A) and saturated hydraulic conductivity (B) when a sandy clay loam, after field traffic with low ground pressure (LGP) or high ground pressure (HGP) tires, was subjected to different uniaxial stresses at various water contents. Solid lines refer to compression of aggregate mixtures (from Dawidowski and Lerink, 1990).

Fig. 5 shows the effect of tillage and wheel traffic on infiltration rates at different nominal pore diameters (diameter is inversely proportional to negative water potential) in a silty clay loam. As sequentially smaller macropores emptied, infiltration rates decreased. Compaction reduced infiltration rates in both no-till and chisel treatments. Compaction also changed the slope of the lines in Fig. 5. A decrease of the intercept with constant slope shows proportional reduction in both larger and smaller macropores in which there is water flow. The observed decrease in slope shows that flow in larger pores is more affected by compaction

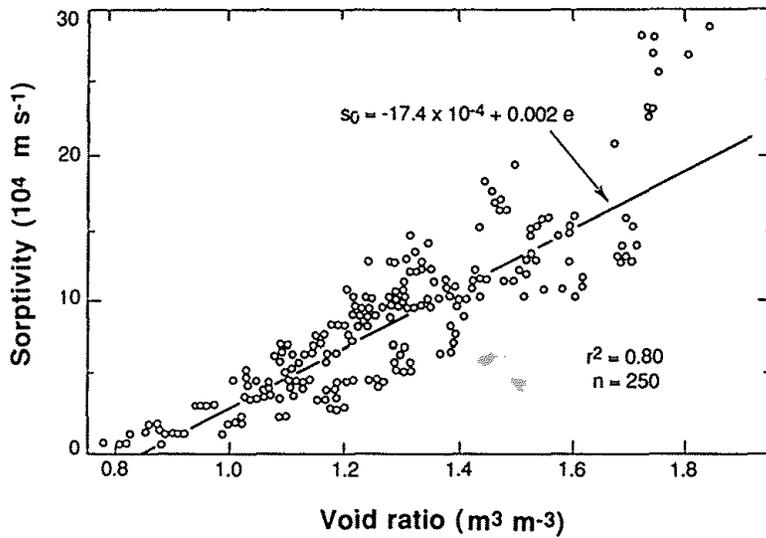


Fig 4. Relationship between sorptivity (S_0) and void ratio (e) in a silt loam soil with different water contents (from Walker and Chong, 1986).

than flow in smaller pores. This observation suggests that compaction destroys more large than small macropores, a phenomenon often measured in desorption curves. Because measured infiltration rates under negative water potential are quite sensitive to compaction, the simplicity and rapidity of these field techniques proves useful in quantifying compaction and other management effects

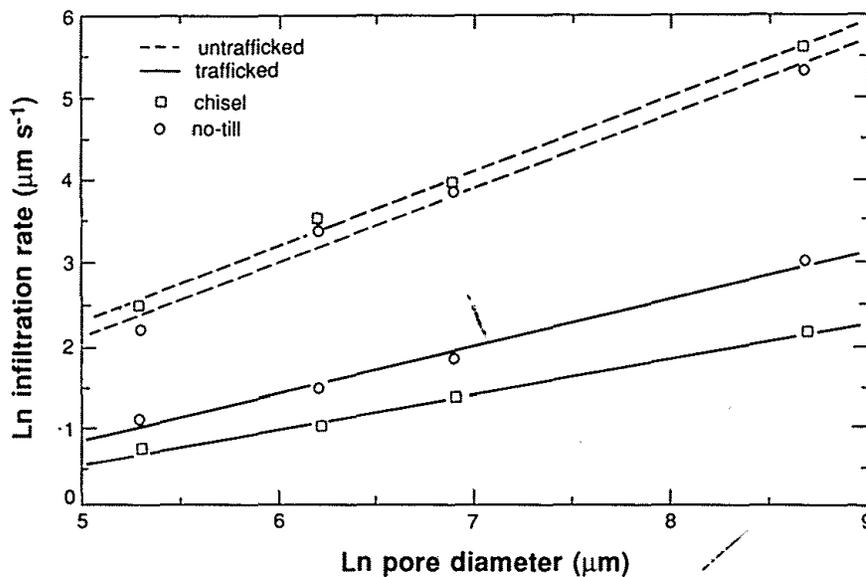


Fig. 5. Influence of compaction on the relationship between unconfined infiltration rate and largest nominal water-filled pore diameter for two tillage treatments on trafficked and untrafficked soil (after Ankeny et al., 1990a).

on hydraulic properties. When there was traffic, the increased separation of datapoints between chisel and no-till treatments (Fig. 5) suggests that infiltration may have responded to the degree of rutting related to antecedent water content or soil strength. Sorptivity also shows great sensitivity to compaction (Walker and Chong, 1986) and can be readily measured in the field. Sorptivity and conductivity measurements might therefore be useful in predicting, as well as measuring, compaction.

The high variability of unconfined hydraulic conductivity measured in a tillage study on a Webster silty clay loam is typical (Table 1). Means within each treatment shown in Table 1 display a response to matric potential. The coefficients of variation are typical for controlled wheel traffic plots. Higher variability when there was traffic is at least partially due to increased measurement error associated with measuring low flow rates. Because traffic has such a profound influence on infiltration properties (Table 1), variability can be higher where measurements are taken without knowledge of the compaction history of the site.

Soil compaction dramatically influences solute transfer, particularly because of the influence on water flow. In fact, solute measurements can infer water flow responses to compaction. Kluitenberg et al. (1988) compared solute breakthrough curves for undisturbed and compacted soil samples. Similarly to Hill et al. (1985), they found that compaction reduces the number of large pores, which is shown in breakthrough curves by reductions in both mean pore-water velocity and hydrodynamic dispersion.

Kiuchi et al. (1990) and Kirkham and Horton (1990) have proposed to use soil compaction as a means to retard nitrate leaching. Where fertilizer is banded, a compacted layer either above or below the band, will direct water flow away from the band into less compacted soil on both sides. In fact, banding fertilizer without formation of a compacted layer would leave a macroporous pathway for enhanced water flux through the banded fertilizer, which is analogous to the flow observed through chisel marks (Pikul et al., 1990). Both laboratory and field studies show that such water flow barriers from selected compaction of soil zones are effective retardants to fertilizer-nitrate leaching. Further investigation of this use of soil compaction to control water and solute fluxes is warranted.

Compaction of soil generally, but not always, increases bulk density, whereas changes in porosity and pore geometry are always produced. Changes of these soil properties must be considered in the context of compaction in a soil layered and variably structured by tillage systems.

LOOSENING EFFECTS OF TILLAGE ON SOIL HYDRAULIC PROPERTIES

Tillage deforms (strains) soil by applying tensile, shear, and compressive stresses. Every tillage tool is unique in its spatial application of stress and therefore, the strain or deformation would also be spatially unique (Koolen and

Kuipers, 1983). When tillage tools are used to loosen soil, various traction, transport, and depth-control devices (tires, wheels, tracks, sliding plates) may produce a zone of soil compaction just below the depth of tillage tool action, alter strain in the soil in response to the tillage tool, and/or recompact bands of soil. Soil structure may also change in response to subsequent wheel traffic without tillage, as well as to biological activity and weather-related inputs of energy. Currently no theoretical basis exists for predicting soil water properties from use of a tillage tool alone or in combination with similar or different tillage tools (Hadas et al., 1988).

For purposes of water flow, a generalized one-dimensional model consists of three soil layers: (1) a tilled and packed A_p layer; (2) an intermediate sublayer rarely tilled but subject to packing; (3) a subsoil unaffected directly by tillage and traffic (Allmaras and Logsdon, 1990). A fourth layer is the surface of the tilled layer subject to sealing and crusting. Such a model systematizes the frequency, spatially and temporally, with which water flow-related parameters must be measured/remeasured. Layering not only affects water flow rates but can also alter water retention of a soil profile after subsequent drainage and redistribution of water (Miller and Bungler, 1963; Miller, 1964; Allmaras et al., 1982).

After tillage and before water has infiltrated, the tilled layer is an admixture of clods, incorporated residue, and tillage voids (unstable voids created by tillage; see Chapter 5) (Soane, 1990; Staricka et al., 1991); the incorporated materials, including broadcast agrichemicals and weed seeds, are located in the tillage voids. The clods may or may not have a wide variation in density depending on soil management prior to the most recent tillage. Vertical uniformity is rare. After some infiltration and biological activity, these tillage voids should resemble flow paths around clods or aggregates.

Rawls et al. (1983) predicted hydraulic properties in the tilled layer and their change during subsequent slaking and natural recompaction. They combined the usual prediction of retention and hydraulic conductivity functions (using organic carbon content, soil texture, effective saturation and pore size distribution) with field-measured porosity changes produced by moldboard plowing, which were also related to soil texture. After first accounting for total porosity produced immediately after moldboard plowing, they then accounted for both seasonal decrease of total porosity (in the absence of wheel traffic) and changes in total porosity for other tillage systems (i.e. chisel, plow-disk-harrow, rotary, plow-pack using traffic) relative to that produced by moldboard plowing.

Brakensiek and Rawls (1983) modeled hydraulic conductivity of the surface crust of the tilled soil, which was made responsive to inputs of rainfall energy (original K_s estimated by Rawls et al., 1983); random roughness of the surface, and a texture-based steady-state matric potential drop across the surface seal. Recent refinements in interpreting the influence of random roughness on depressional storage and soil deposition can improve the Brakensiek and Rawls (1983) model. Moore and Larson (1979) developed a system of routing water

over the soil surface to estimate surface storage by use of point data originally measured to estimate random roughness produced by tillage or compaction. Onstad (1984) demonstrated that measurement of precipitation excess (to fill field-estimated depressional storage) was also required because runoff was initiated before maximum depressional storage volume was attained. Linden et al. (1988) presented a theoretical model to relate ponded area, depressional storage volume, and runoff volume to random roughness and its decline during the application of rainfall energy. They included the concept that soil detachment began after infiltration rate lagged behind rainfall rate and a ponded area was initiated; as roughness declined, the surface seal formed in the depressional areas.

INFLUENCES OF SOIL STRUCTURE ON HYDRAULIC PROPERTIES OF SUBSOILS

The pedal nature of soil structure can be used to predict hydraulic properties in soil layers below the tilled layer. Pedal properties are routinely described by pedologists and may be changed directly by soil tillage and traffic in the upper 30 cm of a soil profile (Bouma et al., 1975; Wang et al., 1985; McKeague et al., 1987). However, pedal properties are more obvious factors in the structure and hydraulic properties of soil layers below the tilled layer. When the subsoil is compacted by heavy axle loads, particularly during harvest (Håkansson et al., 1988), changes in the hydraulic properties of a subsoil may adversely influence hydrologic responses in the tilled layer.

McKeague et al. (1987) define pedality as "the natural organization of soil particles into units (peds) which are separated by surfaces of weakness that persist through more than one cycle of wetting and drying in place." McKeague et al. (1982) estimated K_s from soil morphological observations. Eight classes of K_s , ranging from <0.05 to $>139 \mu\text{m s}^{-1}$, were described on the basis of combinations of texture and morphology. Soils with high sand or clay contents had a high K_s , but biopores and a blocky structure had the largest influence on K_s . The textural relation is not the same as that given by Rawls et al. (1982), in which K_s predictions decreased as sand content decreased and clay content increased. The predictions agree with field measurements of K_s by Topp et al. (1980). Horizons with clay texture had massive and compressed structure, few or no macropores and low K_s ; some of these horizons were in the tilled layer and were often at the base of the A_p or upper part of the B horizon and were associated with tillage pans. The degree of compression and associated K_s estimate were difficult to assess visually.

Micromorphometric information concerning planar voids, tillage voids, and tubular pores in both aggregated and non-aggregated structure was used to predict unsaturated hydraulic conductivity and a water retention characteristic (Bouma and Anderson, 1973). However, it was concluded that direct physical

measurement is much easier.

The estimation of K_s based on observed structure and texture in the field, has most application in soil layers deeper than the depth of tillage tool action but it can also be used in the tilled layer, especially if severe compaction is involved. Both measured K_s and K_s estimated from soil morphology (structure) detected a tillage pan (produced by moldboard plowing and furrow traffic) at 10-25 cm under continuous corn, but not in adjacent hayland or in first year corn after hayland on clayey soils (Wang et al., 1985). These K_s changes (as much as 100-fold reduction) were much greater than any changes of dry bulk density or total porosity; thus, the massive and compressed structure without macropores or a blocky structure could be detected only by observing structure or measuring K_s . In tillage- and traffic-affected layers, a structure that is more pedal than angular blocky or subangular blocky is rarely found; rather the structure may be platy (compressed), granular or massive.

Water movement in these tillage-affected zones (usually not deeper than 45 cm) may be affected by pedal structures in B horizons below 45 cm depth. Compared to coarse prismatic structure, medium subangular blocky structure (both with the same texture) had a much higher K_s and a much larger hydrodynamic dispersion coefficient during saturated flow. Although hydrodynamic dispersion was much reduced in drained soil columns with pulse applications of water, the subangular blocky structure had a much larger dispersion coefficient than the prismatic structure (Anderson and Bouma, 1977).

MULTI-DIMENSIONAL ASPECTS OF COMPACTION

Compaction is often associated with two-dimensional hydrologic effects because some form of tool action or trafficking occurs on or within the tilled layer (Lindstrom et al., 1981; Reicosky et al., 1981; Voorhees and Lindstrom, 1984; Voorhees et al., 1985). Local slope configuration and orientation, such as in ridge tillage (Hamlett, 1987; Van Es et al., 1988), and strips of heavier residue cover, accentuate these hydrologic effects. They are usually most intense just after tillage; their degree of influence is often tempered with time.

A wheel track may cause significant non-uniform drying during an evaporation period (Reicosky et al., 1981). Matric potential profiles at a selected time after initiation of surface evaporation for both "dry" and "wet" wheel track compaction methods, show clearly that matric potential varies in a two-dimensional fashion, with depth and horizontal location (Fig. 6). Both the soil surface and the soil beneath the wheel track are sinks for water because water potential gradients decrease toward the surface and laterally towards the wheel track region. Vertical and lateral fluxes of water were not determined experimentally, but this experiment shows clearly the need for further two-dimensional experimental as well as two-dimensional numerical studies to characterize traffic compaction effects on soil water flow.

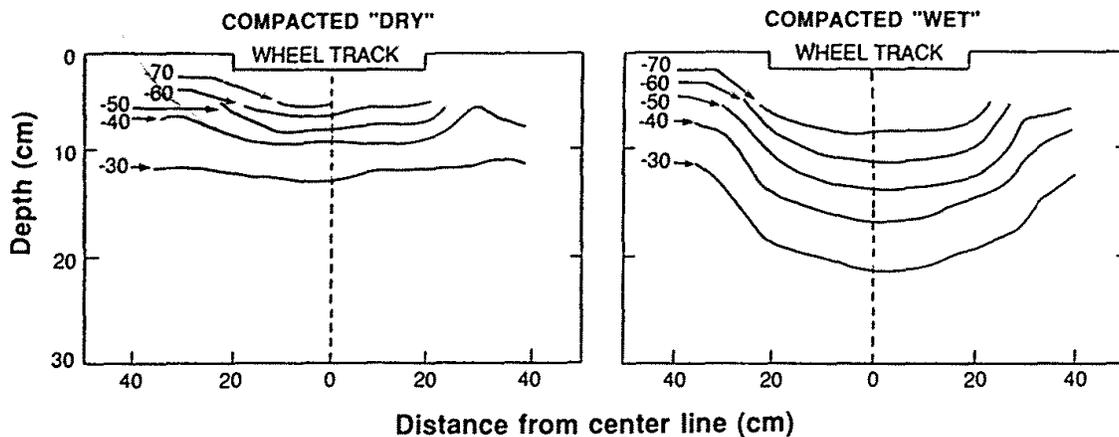


Fig. 6. Matric potential (kPa) distributions in "dry" and "wet" compacted wheel tracks, after 14 days of drying (from Reicosky et al., 1981).

Variability of soil water properties in tilled soils has been studied spatially (Hamlett et al., 1986; Cressie and Horton, 1987) and temporally (Mapa et al., 1986). Matric potential and infiltration of soil water showed spatial variability in both tilled and no-till systems. Analyses indicated that tillage with larger soil disturbance levels may provide more spatial correlation in physical condition of the soil surface. Water retention in freshly tilled soils changed dramatically with successive wetting and drying cycles (Mapa et al., 1986); associated hydraulic conductivity curves also changed with time as a function of wetting and drying cycles. Wetting and drying cycles and freezing and thawing cycles were also reported to increase infiltration rates into previously compacted soil (Akram and Kemper, 1979).

MODELING SOIL WATER FLOW

Numerous models of soil water flow have been developed, some of which have potential to characterize field compaction effects on soil water flow. Both one-dimensional and two-dimensional models are included.

Van Genuchten (1978) modeled flow in layered soil profiles using several numerical schemes; both finite-difference and finite-element simulations offered advantages depending on initial and boundary conditions. Soil water flow and distributions of water content and potential were predicted for soil profiles with both gradual and abrupt changes in properties with depth.

Tillage effects on soil water flow were predicted using one-dimensional numerical models (Hammel et al., 1981; Mapa et al., 1986; Culley et al., 1987a). Hammel et al. (1981) predicted and measured seed zone volumetric water contents to be higher in conventionally tilled soil than in no-till under fallow conditions with no mechanical disturbance; the no-till condition was too dry for successful establishment of winter wheat. Culley et al. (1987a) used two models,

a simple water budget and an integrated soil-plant-atmosphere model (NTRM), to predict soil water regimes under no-till and conventional tillage of a highly structured Mollisol. Output from the NTRM model agreed better than the simple water budget model with experimental data, showing water contents to be lower under conventional tillage than no-till. Mapa et al. (1986) predicted the impact of temporal changes in hydraulic functions on water movement in the soil by comparing water content profiles predicted with a simulation model using hydraulic input functions obtained before and after irrigation. Water content profiles after infiltration and redistribution differed substantially, depending upon hydraulic properties of the soil. Temporal changes in soil hydraulic properties after tillage can therefore influence soil water flow and soil water content.

For water flow simulation in fine-textured soils with cracks and macroporous channels sensitive to shrink-swell, Jarvis and Leeds-Harrison (1990) have developed a dual-region model that treats water flow in large continuous cracks in dynamic equilibrium with water content changes in aggregates (or blocks of swelling/ shrinking soil). Some unusual soil parameters were used, e.g., stable crack porosity, slope of the shrinkage characteristic, aggregate diameter distribution and sorptivity at the wilting point in each of the soil horizons specified.

Two-dimensional models have been used to study water flow. Chung and Horton (1987) predicted that a partial surface mulch changes surface evaporative water flux, water contents, temperature, and water potential variations compared to bare soil. The mulch, however, was shown to have almost no effect on drainage. Hamlett (1987) used a finite-element model to predict infiltration and redistribution of water and anions in flat and ridged soil surfaces. He showed that water entered uniformly into the flat profile, while more water infiltrated from the base of the ridge than from the ridge top. Thus, a tracer anion in the ridge did not move as deeply as it did under the flat surface condition. Whisler et al. (1982) modified the GOSSYM cotton growth simulation model to account for changes in hydraulic properties produced by cultivation and wheel traffic. The model predicted rooting patterns to vary according to external processes, and the authors state that the predicted root patterns agreed with observations.

Benjamin et al. (1990a,b) described a two-dimensional heat and water flow model that includes variable soil properties and surface configurations. The model allows for compacted and loosened zones of soil in a ridge tillage system (Fig. 7). Soil physical properties in the planted row, the untracked interrow, and the wheel-tracked interrow positions, were used to model the effects of soil variability (also compaction) on subsurface water and heat transport. Soil below the dotted lines in Fig. 7 was assumed to have properties of the untracked interrow. Simulated water and heat flow over a period of 134 h, produced isolines of soil water potential and soil temperature that were sensitive to ridge configuration and traffic compaction in the furrow (Fig. 8). Isopotential response to traffic compaction was noted as deep as 50 cm. Isopotentials under the non-

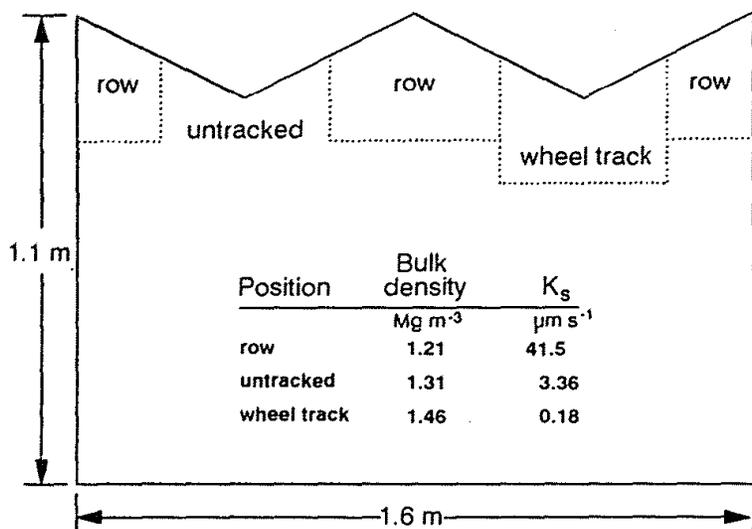


Fig. 7. Soil compaction-related soil properties measured in a ridge tillage system on a Monona silt loam (from Benjamin et al., 1990a).

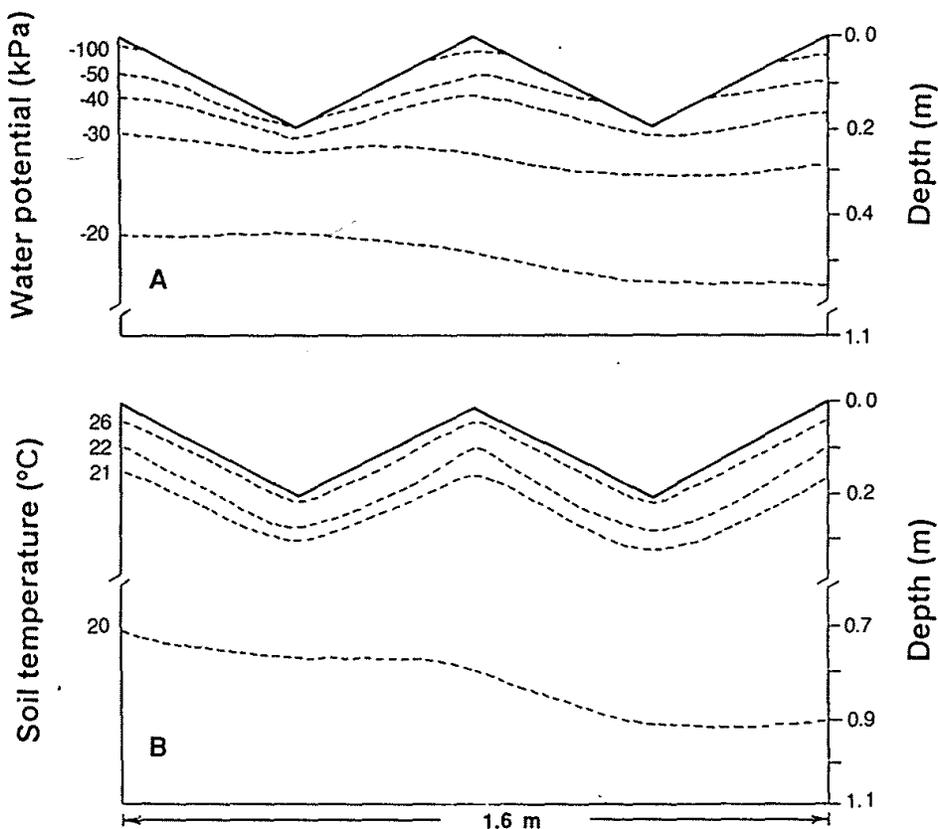


Fig. 8. Predicted soil water potential (A) and soil temperature (B) after 134 h in a ridge tillage system with variable degrees of compaction (cf. Fig. 7) and exposed to a typical diurnal boundary condition at the surface (after Benjamin et al., 1990a).

tracked furrow responded to changes in coupled heat and water flow produced under the tracked furrow. Isothermals more clearly followed the ridge shape and did not show asymmetrical trends due to compaction until deeper than 70 cm. Future research efforts should include studies of variable surface configuration and variable soil properties on soil heat, water and chemical dynamics because it is virtually impossible to maintain constant soil properties horizontally in an arable field.

Most literature on plant response to compaction does not separate the direct from the indirect effects. Research must first identify the direct soil water responses to compaction and tillage and then use simulation with validation to identify the indirect effects. Several especially helpful discussions on this matter are given by Richter (1987), Larson et al. (1989), Rendig and Taylor (1989), and Gliński and Lipiec (1990).

CONCLUSIONS

(1) Among soil properties affected by field compaction and related tillage, perhaps least is known about soil hydraulic properties and processes in spite of their importance.

(2) Although compaction usually does not directly change soil water properties below 30 cm depth, changes of soil water properties above 30 cm depth may dramatically affect water regimes in the subsoil. Rarely is matric flow the dominant mechanism for water flux in these subsoil layers, which contain biopores and planar voids.

(3) Changes of soil water properties in the tilled layer as related to traffic, temporal changes after tillage and extreme layered arrangements, all make it difficult to measure and account for changed soil hydraulic properties and processes in the tilled layer. Matric flow does not dominate because tillage, planar, and biopore voids are usually present in the tilled layer.

(4) A variety of methods are available for *in-situ* measurements and subsequent laboratory measurements on undisturbed samples. Positive/negative pressure infiltrometers are especially useful for studying soil hydraulic properties *in situ* or in the laboratory on undisturbed samples. The proportion of total water flow which passes either through the macropores or through the soil matrix, is influenced by structural changes by tillage and compaction. There is no direct method to estimate soil water properties from a given tillage system and specified soil properties before tillage. However, soil water properties can be estimated, at least generally, from a knowledge of soil structural properties after tillage is completed.

(5) Numerical models are available to predict field compaction effects on soil water regimes in soils that are layered or consist of a matric flux into and out of macro-porous paths. Two-dimensional models utilize surface configuration features related to wheel tracks and residue placement in rows and between rows.

(6) Matrix and macro-porous flow occur simultaneously in all but very sandy soils. Such awareness will improve the flow predictions in the upper soil layers formed by tillage and compaction. Coupled heat and water flow simulations demonstrate relatively greater influences of compaction on the soil water regime relative to the thermal regime close to the soil surface. Soil water regimes can have strong influences on the dynamics of the movement of mineral plant nutrients and pesticide chemicals.

(7) Direct influences of compaction on hydraulic properties may control infiltration and redistribution of water in the soil profile. This water flow process has many indirect and important influences on plant rooting and growth, such as aeration status, soil mechanical resistance to rooting, root extension and uptake of water and plant nutrients. These effects should be included in models predicting the influence of soil compaction on soil hydraulic properties.

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