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Hydrologic Processes in Vertisols

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VERTISOLS AND TECHNOLOGIES FOR THEIR MANAGEMENT

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Chapter 2

PEDOGENESIS

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2.1. INTRODUCTION

The genesis of Vertisols is strongly influenced by soil movement (contraction and expansion of the soil mass) and the churning (turbation) of the soil materials. The name of Vertisol is derived from Latin "vertere" meaning to turn or invert, thus limiting the development of classical soil horizons (Ahmad, 1983). These soils have the capacity to swell and shrink, inducing cracks in the upper parts of the soil and distinctive soil structure throughout the soil (see Chapter 4, Soil Morphology). The formation of these specific features are caused by a heavy texture, a dominance of swelling clays in the fine fraction and marked changes in moisture content (Hubble, 1984). The swell–shrink behavior is attributed to the wetting and drying of the soil mass. Time and frequency of surface cracks, penetrating to a depth of 50 cm or more, have been used to define different kinds of Vertisols (Eswaran et al., 1988).

A large body of research is now available, on the genesis and their unusual properties, most of which emphasize turbation or churning as a fundamental process in the formation of Vertisols. Wilding and Tessier (1988) suggest that there are several lines of evidence to prove that the rate of churning is not sufficient to cause extensive mixing, especially in upper sola. While the genesis of these soils has been considered rather simple, it is more probable that they form under multiple genetic pathways, likely more complex than generally recognized (Blokhuis, 1982).

The purpose of this chapter is to provide a review of the available information about the pedogenesis and pedogenic models proposed for the formation of Vertisols. For further information, readers are also referred to Ahmad (1983), Probert et al. (1987), Wilding and Tessier (1988), and other chapters in this book.

2.2. SOIL-FORMING FACTORS

The following section summarizes the influence of the soil forming factors on the soil-forming processes common to most Vertisols.

2.2.1. Parent material

The materials that form Vertisols can be either allochtonous or autochtonous in origin. The former are geographically more extensive and generally occur in the lower parts of the landscape. Vertisols are known to develop on a wide variety of parent materials such as basalts in Australia (Hosking, 1935), calcareous rocks in the West Indies (Ahmad and Jones, 1969), gneisses and sandstones in India (Bal, 1935), deltaic deposits in the United States of America (Kunze et al., 1963), lacustrine deposits in Trinidad (Brown and Bally, 1968), glacio-lacustrine in Saskatchewan (Mermut and Acton, 1985), marine deposits in Guyana (FAO, 1966), and marls (Duchafour, 1977). It is reported that in all cases (except for the Vertisols developed on lacustrine deposits), the materials were recently deposited and that soil formation was still at its early stages (Ahmad, 1983). In the case of Vertisols developed from lacustrine deposits in Trinidad, it is believed that extensive weathering and clay-mineral synthesis had occurred prior to deposition (Rodrigues and Hardy, 1947).

Vertisols may develop *in situ* from the parent materials mentioned earlier, but in general are associated with transported materials (Eswaran et al., 1988). The smectites in these soils could be derived from the original rock or form as a result of neogenesis or transformations from primary minerals. Studies by Weaver et al. (1971) and Kittrick (1971) have shown that both neoformation and transformations require specific microenvironmental conditions.

The formation and stability of smectites are dependent on the Si activity as well as pH. At high pH, and in the presence of high potentials of Si as well as Mg, smectites develop—a process which is also favored by poor drainage conditions. Transformations of mica (ubiquitous mineral in sediments) through depotassification, dealumination of the tetrahedral sheet and finally, silication of the tetrahedral sheet lead to the formation of smectites. The conditions required for the transformations of other primary or secondary minerals into smectites are also similar. The microenvironmental conditions for this transformation are well established; essentially pH conditions of 7 or higher. Direct transformations of feldspars into smectites have also been reported (Eswaran and Wong, 1978). If leaching occurs in the upper part of the soil with the generation of an acid environment, smectites tend to be destroyed and other soil forming processes are initiated (Eswaran et al., 1988).

Smectites are very pliable and conform to the shapes of any harder minerals such as quartz. Upon drying, they form tight masses with very little intergranular spaces. However, on wetting, they swell but with low porosity within the tight masses. As such, the unique physical properties of smectite-rich soils are related to these specific properties of the mineral. For more information see Chapter 5 (Mineralogy and Chemistry of Vertisols).

2.2.2. Climate

In studying the dark-colored Vertisols, Dudal (1965) classified the climatic phases associated with these soils as desertic, arid, semi-arid, tropical monsoon, equatorial, dry Mediterranean, and sub-humid temperate. Oakes and Thorp (1951) placed lesser emphasis on these climatic variations but claimed that the Vertisols occurred extensively in warm-temperate to tropical regions. Currently, Vertisols are known to occur in wider climatic regions than those described earlier. Examples of Vertisols occurring in much colder climates than visualized previously have been documented in several excellent publications (White and Bonestell, 1960; Mermut and St Arnaud, 1983; Dasog et al., 1987; Mermut et al., 1990). Accordingly, amendments have been made to soil taxonomy (Soil Survey Staff, 1994) to cater for these soils.

In essence, the seasonal variations in precipitation and temperature, which favor the formation of smectitic clays as well as provide many of the physical attributes, characteristic of these soils, would be considered as prerequisites for the formation of Vertisols. The variations in climatic conditions result in weathering of primary and secondary minerals during the wet seasons, but encourage the accumulation of basic cations in the dry seasons (Crompton, 1967). Exceptions to this rule are cases where the smectitic clays are allochthonous, as seen in alluvial, coastal, deltaic and calcareous deposits (Ahmad, 1983). In these situations, the seasonal precipitation and its consequent effect on the development and behavior of the Vertisols becomes important.

Areas where Vertisols are found are characterized by a period when the potential evapotranspiration exceeds precipitation, or a dry period (Table 2.1). All these soils are characterized by a period with sufficient moisture deficit to induce cracking, although the intensity in cool temperate regions, such as in Canada (Dasog et al., 1987), is much lower than in the warmer regions. In spite of climatic variations, there appear to be no differences in the soil forming processes and the properties among these soils. However, marked changes in some

TABLE 2.1

Summary of climates in Vertisol areas (Dasog, 1986)

Location	Average annual				Reference
	Precipitation (mm)	Temperature (°C)	PE[a] (mm)	Dry period (months)	
Montana	381	1–6	870	—	Hogan et al. (1967)
S. Dakota	300–550	5–8	—	—	White and Bonestell (1960)
Texas	850	18	1030	—	Buol et al. (1980)
Australia	554	18	1497	8	Dudal (1965)
India	562–1448	25–29	1400–2250	3–11	Murthy et al. (1982)
Sudan	163	29	4075	12[b]	Dudal (1965)

[a]Potential evapotranspiration.
[b]A month is dry when available water (rainfall + soil water) is less than half of the evapotranspiration.

properties are reported across climatic gradients in Sudan and Morocco (Blokhuis, 1982). Higher rainfall resulted in higher intensity of cracking, thinner layers of surface mulch, increased organic matter contents, increased leaching of salts and carbonates, and lower exchangeable sodium.

2.2.3. Vegetation

There is a lack of information regarding the influence of vegetation on the development of Vertisols. However, it is known that, since Vertisols occur in a wide variety of climatic types, the natural vegetation associated with this soil order is equally variable (Ahmad, 1983). The natural vegetation types are, to a certain extent, limited by the soil properties such as the clay contents, shrink–swell characteristics, and soil structure. Both climate and soil properties limit the vegetation types to grasses and slow-growing, deep-rooting tree species with hard wood. The main features of the natural vegetation in these soils are tolerance to drought, as well as the development of deep roots (to overcome root damage as a consequence of the annual cracking phenomena).

Most Vertisols have had grassland as the native vegetation, but some had formed under forest. Natural vegetation of the Vertisol areas in India are mainly open deciduous or tropical thorn forest as judged from the prevailing climate (Simonson, 1954). Acacia is the most common tree species on Vertisols worldwide. A detailed list of native plant species is presented by Dudal (1965). Diverse vegetation is not a critical factor in the genesis of these soils.

2.2.4. Topography and relief

The effect of relief on the development of Vertisols should be examined at two different scales of observation: (a) macro, and (b) micro.

(1) Macro-relief

Field studies have shown that Vertisols generally occur in areas with slopes not exceeding 5% (Mohr et al., 1972; Young, 1976), since at higher gradients, mass movement of soil would occur. Dudal (1965) does not exclude the possiblity of slopes up to 15%, but acknowledges that such situations are rare. It is reported that Vertisols in Trinidad occupy areas with slopes between 10 and 35% (Ahmad, 1983) but are subject to intense erosion, as expected.

Dudal (1965) stated that broad and level areas of Vertisols often lack an integrated drainage network. Infiltration is slow in these soils, which results in surface ponding of rain water for extended periods. Such waters flow slowly on the surface into depressions (Ahmad, 1983) where they form swamps or marshlands.

(2) Micro-relief

The peculiar type of micro-relief seen in many areas occupied by Vertisols is known as "gilgai", which is aboriginal (Australian) for such a landscape. According to Hallsworth et al. (1955), the development of these gilgai is due to the shrinking

and swelling of the Vertisols. The argument is that on wetting and swelling, the soil mass cannot re-occupy the original volume since surficial material has fallen into the cracks during the dry season. As such, part of the soil mass is forced upwards forming the mounds. The formation of a mound provides a locally preferred site for the further release of pressure, thereby perpetuating the formation of other mounds and depressions in an area (Wilding and Tessier, 1988).

2.2.5. Time

It has been mentioned earlier that Vertisols have developed on a variety of parent materials that are basically young in age, but they may occur on old geomorphic surfaces. It has been suggested that slickensides form very rapidly and that Vertisols were found on geomorphic surfaces as young as 550 years (Parsons et al., 1973). Blokhuis (1982) indicated that the formation of slickensides took place <200 years in Romania. According to Yaalon (1971), slickensides form rapidly and approach equilibrium with their environment in a period ranging from 100 to 1000 years.

Vertisols could be old soils, if they are considered to be an end product of a development sequence, involving soils whose B horizons become so clayey (predominantly smectite) that shrink-swell cycles homogenize the solum and convert the soil to a Vertisol. However, in general it is believed that most of the Vertisols occurring in the world are young and immature.

2.3. SOIL-FORMING PROCESSES

Soil-forming processes that lead to the formation of Vertisols are those which control the formation and stability of smectites in the soil (Eswaran et al., 1988). However, subsidiary processes, such as fluctuations in the moisture status of the soil, accumulation of organic matter, carbonates, gypsum or soluble salts and acidification processes through leaching, result in the differences within the Vertisols. The unique morphological expression of the vertic properties can only occur if there is a flux in the moisture status of the soil. The main processes that have been identified to date with regards to the formation of Vertisols are related to the properties of the soils and as a consequence, the morphology of Vertisols. Hubble (1984) indicated five pedogenic processes common to most Vertisols. These are:

(1) Soil movement by swelling and shrinking that cause shearing and consequently the formation of slickensides and the characteristic subsoil structure,

(2) Lateral and vertical movements of soil materials associated with gilgai formation,

(3) Churning which is believed to be a slow process,

(4) Formation of organo-mineral complexations,

(5) Weathering and redistribution of new products under conditions of restricted leaching alternating with drying.

Most common soil-forming processes that are linked to Vertisols and discussed in the literature are: cracking and swelling, self-mulching, churning, formation of slickensides, sphenoids and bowl structures, kankar formation, dark coloration, formation of smectitic clays, and clay translocation.

2.3.1. Cracking and swelling

This process is attributed to the presence of smectitic clays and alternations in climatic conditions (dry and wet seasons). As a result of this process, Vertisols develop deep and wide cracks in a polygonal pattern. The horizontal diameters of these polygons can be up to 4 m (Buringh, 1968). Dasog and Shashidhara (1993) have shown that the volume and intensity of cracking in Vertisols can be variable depending not only climate but also on the management techniques. A more detailed account of the cracking and swelling is given in Chapter 4 (Soil Morphology).

2.3.2. Self-mulching

Self-mulching phenomena, although common in the top few centimeters in Vertisols, is not a characteristic feature of these soils since it was realized that tillage operations can also influence this feature. The self-mulch layer is generally quite thin (mostly 1–5 cm), and normally forms as a result of cracking and swelling in the absence of overburden pressure. The granules formed are up to 3 mm in average diameter and are hard.

2.3.3. Churning

This process, sometimes referred to as self-swallowing (Buringh, 1968), occurs in combination with self-mulching and is a mechanical process which homogenizes the soil material in the upper part of the soil. The varying extents of infilling of the cracks by surficial material during dry seasons prevents these cracks from completely closing up during wet seasons. As a result, the subsoil is pushed up to the surface and the whole cycle would repeat with time. This homogenization process (pedoturbation) also results in the lack of horizon development in Vertisols giving rise to difficulty in identifying horizons other than the A and C horizons.

The pedoturbation process was once identified as a possible mode for the genesis of Vertisols. It is now recognized that this process is an incomplete genetic model as it fails to account adequately for the following relationships and features commonly observed in Vertisols (Wilding and Tessier, 1988):

(a) systematic depth functions of soluble salts, carbonates and organic carbon (Wilding, 1985; Dasog et al., 1987);

(b) systematic increase in the mean residence time of organic matter with depth just as in non-cracking soils (Yaalon and Scharpenseel, 1972);

(c) maximum development of slickensides below the depth of maximum seasonal cracking and infilling (Yaalon and Kalmar, 1978);

(d) formation of gilgai more rapid than that suggested by the rate of crack infilling (White, 1967; Yaalon and Kalmar, 1978);

(e) the presence of weakly expressed albic and textural B_t horizons in clayey soils with slickensides in Canada (Dasog et al., 1987); and

(f) horizontally-bedded stratigraphic marker zones that are not physically distorted, except close to the slickensides (Wilding and Tessier, 1988).

2.3.4. Formation of slickensides, sphenoids and bowl structures

During the drying cycles, cracks develop, whereas, on moistening, shear stresses form which result in slickensides and/or the smoothened surfaces of sphenoids. The combined occurrence of slickensides and sphenoids is not common for all Vertisols since the conditions for their formation are different. However, both require the material to be in a plastic state. Komornik and Zeitlin (1970) stated that the lateral pressures developed in these soils are much greater than the vertical swelling pressures. Within the soil, the vertical component of the swelling pressures includes the weight of the overlying soil material. Eswaran et al. (1988) stated that the moisture conditions above and below a point within such soils determines the net pressure and the angle of shear.

As such, the near surface horizons develop cracks and may have only a few slickensides since both the horizontal and vertical pressures are small (the net pressure as such being much lower than the shear strength of the material). In deeper horizons, typically from 50 to about 125 cm below the surface, slickensides development is maximum (Wilding, 1985). In these deeper layers, the net presure is much greater than the shear strength of the material and soil movement occurs with swelling (see Chapter 14).

The actual deformation zone appears to be a function of the plasticity of the underlying material at the time when swelling is taking place in the overlying material (Eswaran et al., 1988). As an example, if the underlying material is dry or hard, the position of this layer determines the depth at which slickensides occur. If such a layer is close to the soil surface (approximately 20 cm), slickensides will not occur as the overburden pressure is too low to exert swelling pressures.

Sphenoids, on the other hand, develop as a result of the existence of much lower vertical and horizontal pressures in comparison to that needed for the development of slickensides. In the typical case, sphenoids would be found in between the surface horizons with cracks and deeper horizons with slickensides. Their development has been related to lower clay contents, as well as smaller proportions of smectitic clays in the colloidal fractions (Eswaran et al., 1988). It should be noted that despite the general usage of the term "intersecting slickensides", slickensides do not intersect. If they do, the result is the formation of sphenoids. As such, sphenoids are discrete entities, whereas slickensides are modified surfaces (Eswaran et al., 1988).

Studies have shown that the distribution of sphenoids and slickensides in a pedon follow a cyclic arrangement, referred to as "bow structure" (Dudal and Eswaran, 1988; Eswaran and Cook, 1988). The net effect of tilting the long axes

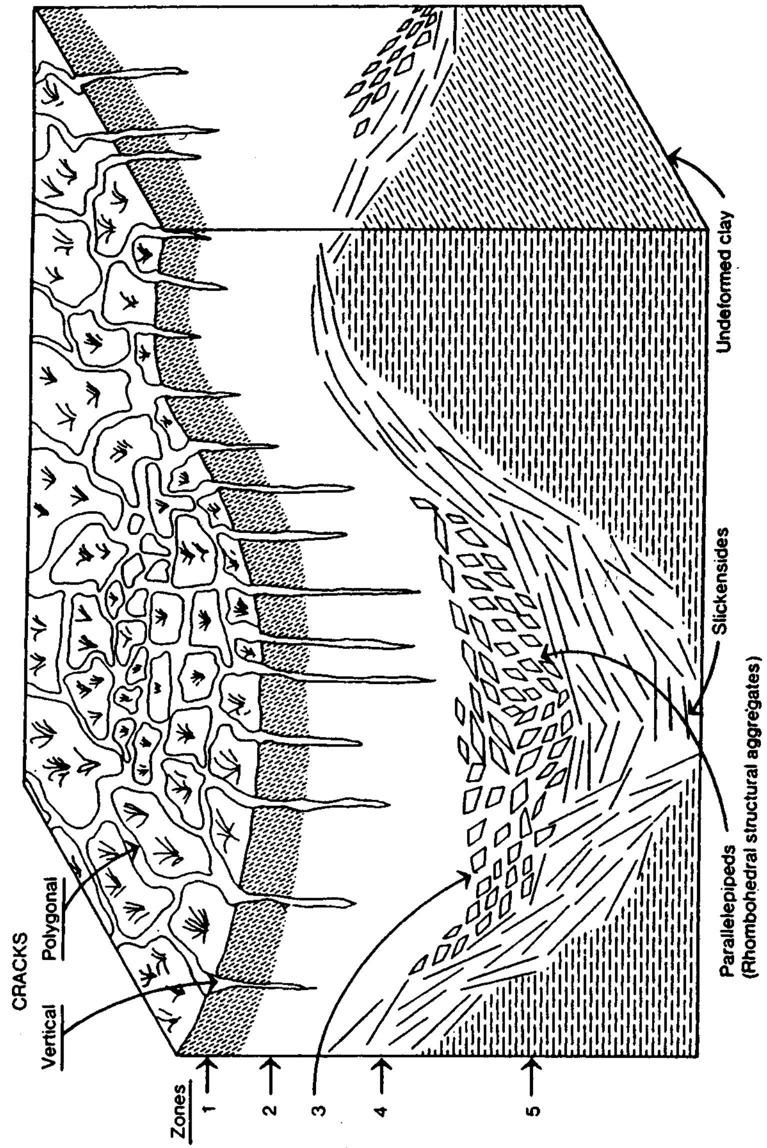


Fig. 2.1. Idealized cross-section of a Vertisol showing the vertical zones in the profile (from Eswaran et al., 1988).

of sphenoids and slickensides in these bow structures forms what has been terme as a synclinorium of the two elements (Eswaran et al., 1988). If gilgai form o the surface, the pattern of the synclinorium appears to follow this microrelief.

As a result of these processes, a typical Vertisol may be subdivided into 5 zon as listed here (Fig. 2.1):

Zone 1 starts from the surface to about 25 cm deep and is subject to intens cracking and formation of large prisms. The material is very hard when dry an may break to coarse, angular blocky elements.

Zone 2 is approximately 10–30 cm thick with essentially very hard and coar angular blocky elements.

Zone 3 is of variable thickness and merge with zone 5 which is underforme clay. This zone is characterized by sphenoids with tilted long axes.

Zone 4 is the zone of slickensides with tilted long axes.

Zone 5 is the zone of underformed clay which is dry and hard, creating th conditions for the formation of slickensides directly above it.

The above-mentioned process is identified as the soil mechanics model for th formation of Vertisols. This process is considered as a viable model to explain th formation of many Vertisols (Wilding, 1985; Eswaran et al., 1988; Wilding an Tessier, 1988).

2.3.5. *Kankar formation*

Kankars (carbonate glaebules or nodules) are basically lime concretions th are found in Vertisols. They are all secondary carbotes (Mermut and Dasog, 1986 Many Vertisols are calcareous in composition and have kankar throughout th profile. Powdery lime pockets may occur at depths and are not affected by th churning processes. In the deeper horizons, it is also not uncommon to find calc horizons (authors observations in India and Indonesia). Drying, in the presen of Fe and Mn, results in the formation of hard concretions whose colors may rang from black to yellowish-gray. Blokhuis (1968/69) suggested that the kankers wi sharp and well rounded boundaries, commonly found at or near the soil surfac are due to upward transportation by pedoturbation process.

Color, mineralogical nature, drainage condition of parent materials, and a of Vertisols are closely associatedd with th size, shape, orientation, chemical an mineralogical composition of hardened secondary carbonates. Therefore, they m provide additional information about pedogenesis of this soil order. For exampl the presence of kankers in the upper part of the profile supports the concept a mixing process (Mermut and Dasog, 1986). Yaalon and Kalmar (1978) suggeste that turbation is a moderate rate process in Vertisols.

2.3.6. *Dark coloration*

Worthy of consideration is that the dark color associated with many of th Vertisols is not due to high organic matter contents but rather due to the inten mixing of organic matter with clay. The organic matter content in these soils generally less than 1.5% (Buringh, 1968). Adsorption of organic matter onto th surfaces of clay particles appears to be mediated through anaerobic conditions that are known to exist temporarily during wet seasons. For similar reasons, very dark Vertisols generally occur in depressions or in soils which experience waterlogged conditions for a part of the wet season. Dudal (1965) suggested that color of the soil may be controlled by the drainage condition, duration and severity of the dry season, composition and age of the parent material.

2.3.7. *Formation of smectitic clays*

The development of Vertisols requires conditions that ensure the formation and preservation of smectites. Such conditions have been identified as high pH with sufficient Ca^{2+} and Mg^{2+} in the soil system. An additonal requirement would be the presence of a relatively impermeable layer at some depth within the soil, which would prevent the leaching of the various components needed to form such clays (see Chapter 5, Mineralogy and Chemistry of Vertisols).

2.3.8. *Clay translocation*

It is recognized that the fine-sized smectitic clay which has a high specific surface area, has all the conditions necessary for dispersion, translocation and accumulation in subsurface horizons in Vertisols (Dudal and Eswaran, 1988). However, in Vertisols, clay translocation is not phenomenal, since pedoturbation processes tend to obliterate all evidence of the illuviation process. Nevertheless, illuviated clay has been identified in zone 5 (Fig. 2.1), which is subjected to the least amount of pedoturbation. With advancement of leaching and the formation of an argillic horizon, the soils would evolve into Alfisols (example Vertic Hapludalfs). Leaching also promotes the destruction of smectites. Therefore, in the transformation of Vertisols to Alfisols, the destruction of vertic properties is evident.

2.4. FORMATION OF VERTISOLS

Vertisols can form either *in situ*, through the weathering and development of a solum (autochtonous Vertisols) or from a sediment which is composed of materials that can produce vertic properties (allochtonous Vertisols). The latter is geographically more extensive and occupies the lower parts of the landscape. These two groups may form a catenary sequence in a landscape.

Nettleton et al. (1969) and Buol et al. (1980) have suggested that soils with an argillic horizon might become so enriched in clay through illuviation and *in situ* clay formation that they could develop sufficient shrink-swell activity and evolve into Vertisols. They consider that the surface horizons are incorporated in the soil and as a result, argillic features are obliterated. This line of Vertisol formation appears to be very minor.

Sufficient shrink-swell activity that would produce sphenoids and slickensides is critical for the formation of Vertisols. As discussed earlier, this can only take place if there is a flux in the moisture status of the soil. The shrinking on drying

results in cracks and compaction of the soil material, however, the swelling on moistening is responsible for the shearing which results in the formation of slickensides and/or the smoothed surfaces of the sphenoids. Most Vertisols have a combination of sphenoids and slickensides, but some have only one or the other. Formation of both features require the soil material to be in a plastic state.

According to Komornik and Zeitlin (1970) the lateral swelling pressures in soils are generally much larger than vertical swelling pressures, as the latter is substantially reduced by the overburden pressure. Swelling pressures of 400–1000 kPa and shear strengths of 20–40 kPa were measured in a cold Vertisol from Regina (Saskatchewan, Canada) by Fredlund (1975). The ratio for the Regina soil is about 20 times in favor of shearing. This shows that shearing is inevitable in Vertisols. It should be remembered that net pressure is a function of moisture content of the soil.

The site of maximum slickensides is between 50 and 125 cm depth, in most Vertisols (Fig. 2.2). However, fewer slickensides are found at depths between 25 and 50 cm. At such depths both vertical and horizontal pressures are small (Yaalon and Kalmar, 1978). As the moisture changes become limited at 125 cm depth, slickensides become rare below this depth. If the soil is shallow slickensides are not likely to form. Several such soils were observed by the authors in India.

Evolutionary stages of Vertisol formation are shown in Fig. 2.3. Sphenoid development is related to the vertical and horizontal pressures, both of which are controlled by moisture conditions of the soil. In some Vertisols swelling pressure does not result in slickensides, but may be only sphenoids. Figure 2.3 depicts the distribution of the sphenoids and slickensides in four soil pedons (Eswaran and Cook, 1988). This figure also illustrates the conceptual evolution of Vertisols from an unripe soil to maturity. As the long axis of the sphenoids and plane of slickensides are tilted, the net effect is a synclinorium of the two elements as shown in Fig. 2.1.

As discussed earlier, the stability of Vertisols depends on the pH and supply of Si and bases. Soil acidification causes clay dispersion and its downward translocation within the profile. Leaching also promotes destruction of the smectites. If this happens the Vertisol properties diminish, resulting in the formation of other soils.

2.5. PEDOGENIC MODELS

While Blokhuis (1982) suggested that the genesis of Vertisols is remarkably homogeneous and conforms to a single and simple process, Wilding and Tessier (1988) consider that Vertisols form through multiple genetic pathways, considerably more complex than that suggested by Blokhuis (1982). In either case, the genesis of Vertisols is strongly influenced by soil movement. Wilding and Tessier (1988) recognize three pedogenic models under which prevailing theories of Vertisol genesis can be grouped: (1) pedoturbation model, (2) soil mechanics model, (3) differential loading model.

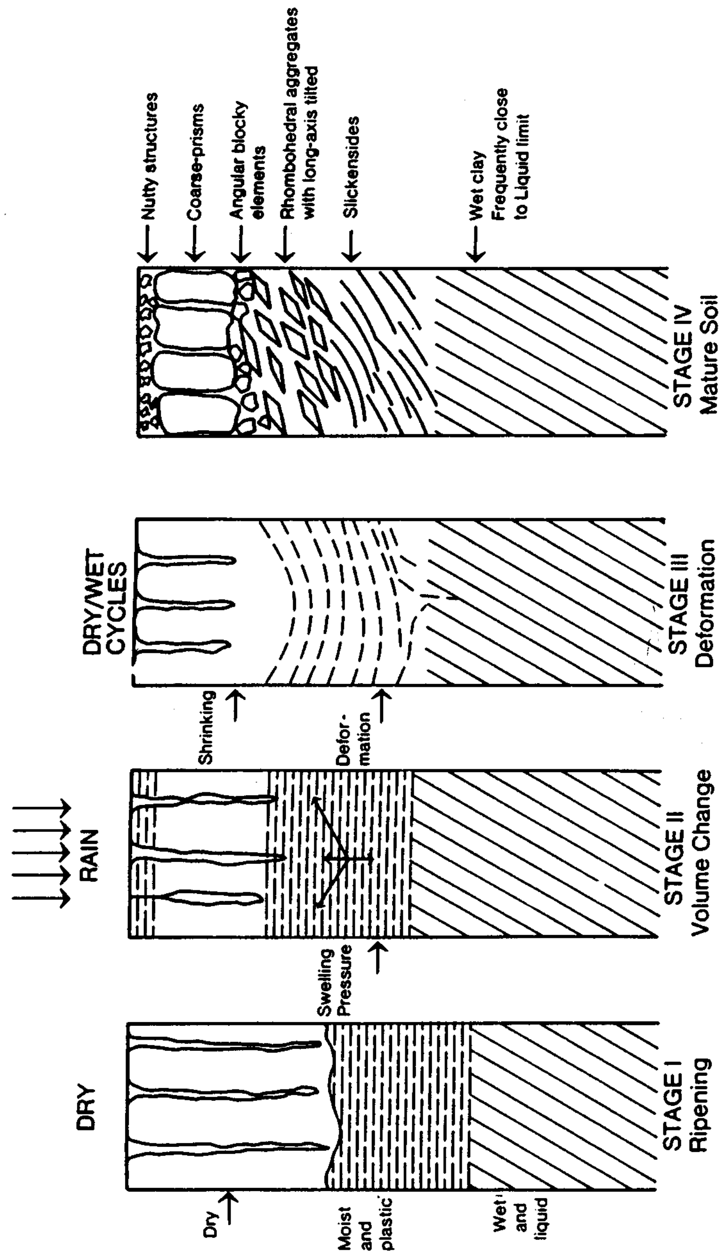


Fig. 2.2. Morphological changes associated with the formation of Vertisol.

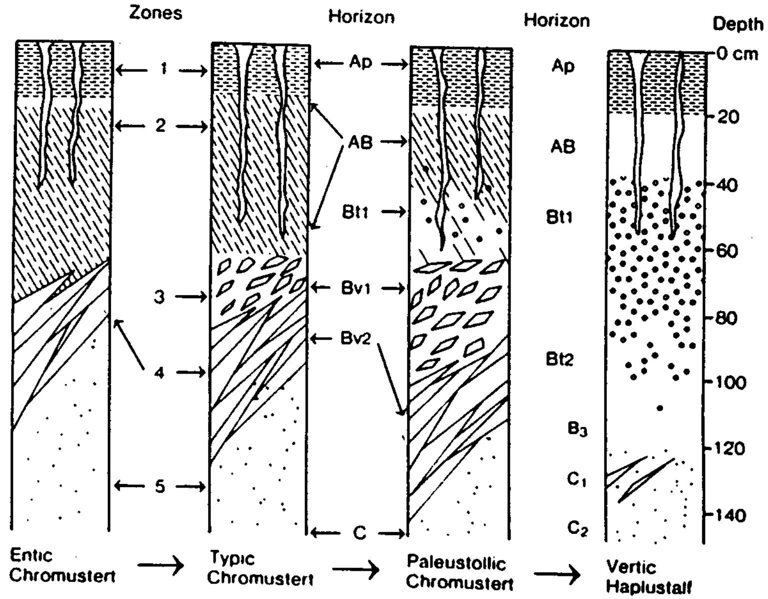


Fig. 2.3. The conceptualized successive stages of pedogenic evolution in Vertisols (Eswaran and Cook, 1988).

2.5.1. Pedoturbation model

This is described in detail by Buol et al. (1980) (Fig. 2.4). According to this classical model, when soil cracks during the dry season, the surface soil sloughs into the cracks. Rewetting induces high swelling pressures that are reviled by displacement of the soil upward and sideways, resulting in slickensides, gilgai, etc.

2.5.2. Soil mechanics model

This appears to be the most popular model among pedologists. According to this model, slickensides form when swelling pressure exceeds the internal forces resisting the shear, resulting in soil displacement; sloughing of soil is not a prerequisite to generate high swelling pressures. Ritchie et al. (1972) and Blake et al. (1973) demonstrated unequal wetting in clay soils whereby soil starts wetting from above and the bottom of the cracks. As mentioned earlier, when swelling occurs under confinement, lateral pressure exceeds vertical pressure, as mentioned above. The large resultant unequal stresses that act on small blocky units give rise to inclined shear stresses that exceed the shear strength of the soil, resulting in failure planes called slickensides.

The shear strength of a soil is a function of cohesion plus the angle of internal friction. The angle of internal friction in clay soils is low and the cohesion is a function of bulk density, clay content, clay mineralogy, and is inversly related to

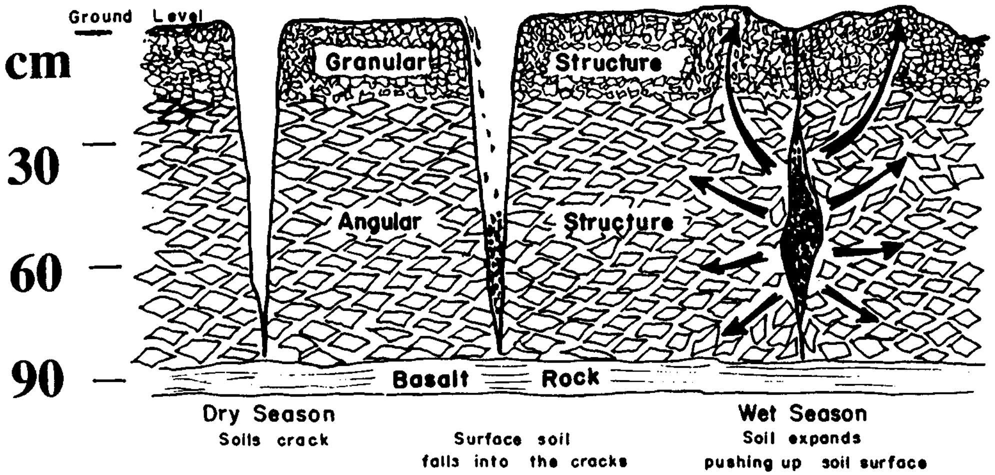


Fig. 2.4. Sketch illustrating the effect of wetting and drying cycles on self-swallowing and gilgai formation (Buol et al., 1980).

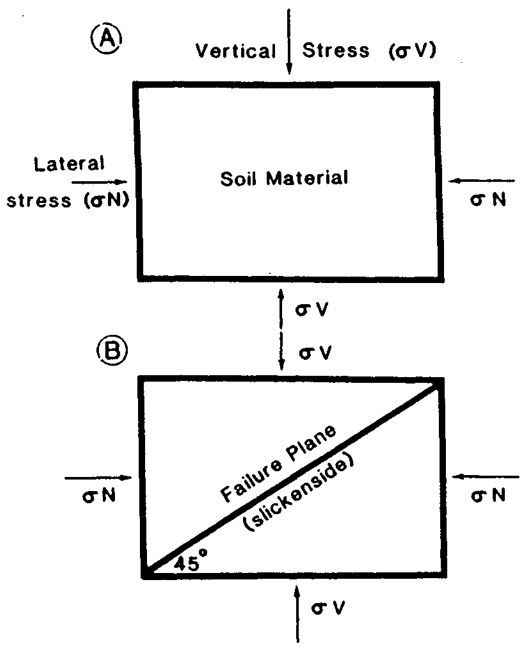


Fig. 2.5. Soil mechanics model of slickenside formation. (A) Vertical and horizontal stress acting on a soil ped; (B) orientation of shar plane at 45° to the principal stress (Wilding and Tessier, 1988).

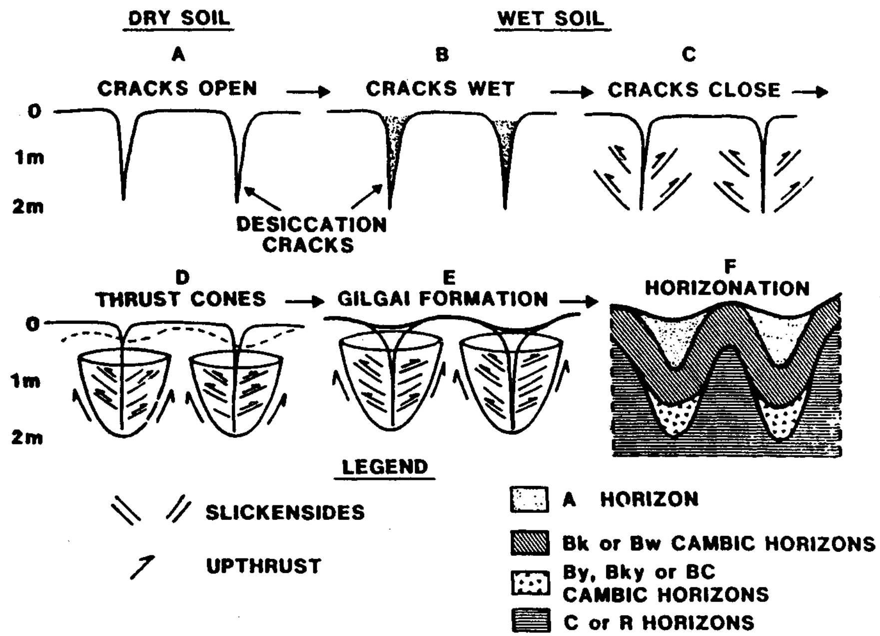


Fig. 2.6. Schematic illustration of possible stages (A–F) in the formation of slickensides, gilgai and cyclic horizonation (Wilding and Tessier, 1988).

moisture content (Petry and Armstrong, 1980; Wilding and Tessier, 1988). During swelling, a soil is acted upon by vertical and lateral stresses (Fig. 2.5). The soil material shears along slickenside glide planes. When observed in three dimensions, slickenside patterns often form cones of revolution (or bowls) with the vertex centered in the micro-low (Wilding and Tessier, 1988; Eswaran et al., 1988) (Fig.2.6).

The theory underlying the soil mechanics model is presented independently by Yaalon and Kalmar (1978) and Knight (1980). The slickensides form at a particular depth range as illustrated in Fig. 2.7 by Yaalon and Kalmar (1978) and are dependent on many factors. The distribution of slickensides with depth is also based on soil mechanics principles. As suggested by White (1966) , the weight of the overlying soil increases the packing density so that resistance to shearing will be greater at depth. As a result, slickensides will be less numerous. Deep-seated slickensides are likely formed during the ripening of sediments, large scale land slipping or mass wasting rather than the soil formation (Blokhuis, 1982; Ahmad, 1983).

2.5.3 *Differential loading model*

Paton (1974) explains the genesis of gilgai by a process of differential loading, whereby clays move from areas of high to areas of low confining pressure.

Kittrick, J.A., 1971. Montmorillonite equilibria and the weathering environment. Soil Sci. Soc. Amer. Proc., 35: 815–820.
Knight, M.J., 1980. Structural analysis and mechanical origins of gilgai at Boorook, Victoria, Australia. Geoderma, 23: 245–283.
Komornik, A. and Zeitlin, J.G., 1970. Laboratory determination of lateral and vertical stress in compacted swelling clay. J. Materials, 5: 108–128.
Kunze, G.W., Oakes, H. and Bloodworth, M.E., 1963. Grumusols of the coastal Prairie of Texas. Soil Sci. Soc. Amer. Proc., 27: 412–421.
Mermut, A.R. and Acton, D.F., 1985. Surficial rearrangement and cracking in swelling clay soils of the glacial lake Regina basin in Saskatchewan. Can. J. Soil Sci., 66: 317–327.
Mermut, A.R. and St Arnaud, R. J., 1983. Micromorphology of some Chernozemic soils with grumic properties in Saskatchewan soils, Canada. Soil Sci. Soc. Amer. J., 47: 536–541.
Mermut, A.R. and Dasog, G.S., 1986. Nature and micromorphology of carbonate glaebules in some Vertisols of India. Soil Sci. Soc. Amer. J., 50: 382–391.
Mermut, A.R., Acton, D.F. and Tarnocai, C., 1990. A review of recent reserach on swelling clay soils in Canada. In: J.M. Kimble (Editor), Proceeding of the Sixth International Soil Correlation Meeting, Characterization Utilization of Cold Aridisols and Vertisols. USDA, SCS, National Soil Survey Center, Lincoln, NE, pp. 112–121.
Mohr, E.C.J., Van Baren, F.A. and Van Schuylenborgh, J. 1972. Tropical Soils. Mouton, The Hague, 481 pp.
Murthy, R.S., Bhattacharjee, J.C., Landey, R.J. and Pofali, R.M., 1982. Distribution, characteristics and classification of Vertisols. In: Vertisols and Rice Soils Symposium II, 12th Int. Soil Sci. Congr., New Delhi, India, pp. 1–22.
Nettleton, W.D., Flach, K.W. and Brasher, B.R., 1969. Argillic horizons without clay skins. Soil Sci. Soc. Amer. Proc., 33: 121–125.
Oakes, H. and Thorp, J., 1951. Dark clay soils of warm regions variously called Rendzina, Black Cotton soils, Regurs and Tirs. Soil Sci. Soc. Amer. Proc., 15: 347–354.
Parsons, R.B., Moncharoan, L. and Knox, E.G., 1973. Geomorphic occurrence of Peloxererts, Willamette Valley, Oregon. Soil Sci. Soc. Amer. Proc., 37: 924–927.
Paton, T.R., 1974. Origin and terminology for gilgai in Australia. Geoderma, 5: 221–242.
Petry, T.M. and Armstrong, J.C., 1980. Relationships and variations of properties of clays. In: D. Snethen (Editor), Proc. Fourth Int. Conf. on Expansive Soils, Vol. 1, American Society of Civil Engineers, New York.
Probert, M.E., Fergus, I.F. Bridge, B.J., McGary, D., Thompson, C.H. and Russell, J.S., 1987. The properties and management of Vertisols. C.A.B. International, Oxon, U.K., 20 pp.
Ritchie, J.T., Kissel, D.E. and Burnett, E., 1972. Water movement in an undisturbed swelling clay soil. Soil Sci. Soc. Amer. Proc., 36: 874–879.
Rodrigues, G. and Hardy, F., 1947. Soil genesis from a sedimentary clay in Trinidad. Soil Sci., 64: 127–142.
Simonson, R.W., 1954. Morphology and classification of the Regur soils of India. J. Soil Sci., 5: 275–288.
Soil Survey Staff, 1975. Soil Taxonomy. Soil Conservation Service, US Department of Agriculture Handbook 436, 754 pp.
Soil Survey Staff., 1994. Keys to Soil Taxonomy. 6th ed., USDA SCS, Washington, D.C., 306 pp.

Weaver, R.M., Jackson, M.L. and Syers, J.K., 1971. Magnesium and silicon activities in matrix solutions of montmorillonite-containing soils in relation to clay mineral stability. Soil Sci. Soc. Amer. Proc., 35: 823–830.
White, E.M., 1966. Subsoil structure genesis: theoretical consideration. Soil Sci., 101: 135–141.
White, E.M., 1967. Soil age and texture factors in subsoil structure genesis. Soil Sci., 103: 288–298.
White, E.M. and Bonestell, R.G., 1960. Some gilgaied soils in South Dakota. Soil Sci. Soc. Amer. Proc., 24: 305–309.
Wilding, L.P., 1985. Genesis of Vertisols. In: Proc. 5th. Int. Soil Class Workshop, Sudan, 1982. Soil Survey Administration, Khartoum, Sudan, pp. 47–62.
Wilding, L.P. and Tessier, D., 1988. Genesis of Vertisols: Shrink-swell phenomena. In: L.P. Wilding and R. Puentes (Editors), Vertisols: their Distribution, Properties, Classification and Management. SMSS-Texas A&M University, pp. 55–79.
Yaalon, D.H., 1971. Soil-forming processes in time and space. In: D.H. Yaalon (Editor), Paleopedology. International Society of Soil Science and Israel University Press, Jerusalem, pp. 29–39.
Yaalon, D.H. and Kalmar, D., 1978. Dynamics of cracking and swelling clay soils: Displacement of skeletal grains, optimum depth of slickensides and rate of intra-pedonic turbation. Earth Surface Processes, 3: 31–42.
Yaalon, D.H. and Scharpenseel, H.W., 1972. Radio-carbon dating of soil organic matter in Israel soils. Isr. J. Agric. Res., 22: 154–155.
Young, A., 1976. Tropical Soils and Soil Survey. Cambridge University Press, London, 468 pp.

Spaargaren, O., 1994a. Introduction to the World Reference Base for Soil Resources. Trans.15th World Congress of Soil Science, Vol. 6a, pp. 804–818.

Spaargaren, O., (Editor) 1994b. World Reference Base for Soil Resources. Draft. ISSS/ISRIC/FAO, 161 pp.

Stace, H.C.T., Hubble, G.D., Brewer, R., Northcote, K.H., Sleeman, J.R., Mulcahy, M.J. and Hallsworth, E.G., 1968. A Handbook of Australian Soils. Rellim Techn. Publ., Glenside, South Australia, 435 pp.

Tejedor Salguero, M.L., Quantin, P. and Fernandez Caldas, E., 1978. Climatoséquence de la région méridionale de l'Ile de Ténérife (Iles Canaries). 2ème partie: caractéristiques minéralogiques; interprétation et classification. Cahiers ORSTOM, Sér. Pédologie 16: 83–106.

Thorp, J. and Smith, Guy D., 1949. Higher categories of soil classification. Soil Sci., 67: 117–126.

Van Baren, H. and Sombroek, W.G., 1985. Vertisols in the collection of the International Soil Museum and some suggestions on classification. Proc. 5th Int. Soil Classification Workshop: Taxonomy and Management of Vertisols and Aridisols, Sudan, 2–11 November 1982. Soil Survey Administration, Khartoum, pp. 63–69.

Chapter 4

SOIL MORPHOLOGY

A.R. MERMUT, G.S. DASOG and G.N. DOWUONA

"They are ill discoverers that think there is no land, when they can see nothing but sea."
Francis Bacon

4.1. INTRODUCTION

Morphology is the most important feature used to differentiate Vertisols from other soil orders. Indeed, the concept of Vertisols is derived from their morphological manifestations (Simonson, 1954; Eswaran et al., 1988). The major morphological markers of Vertisols are linear and normal gilgai (micro-relief), cyclic horizons, surface cracking upon desiccation, and slickensides (Soil Survey Staff, 1994). Soil structure (Chapter 7) is also peculiar, especially the occurrence of wedge-shaped aggregates. Slickensides between aggregates are the most characteristic feature of Vertisols (Blokhuis, 1982). Other morphological characteristics described in the literature include color, surface mulching (granular structure), thickness, carbonate, Fe and Mn segregations (glaebules), distribution of clay, bulk density (porosity), and characteristics observed under the petrographic microscope, such as basic related distribution pattern, plasma separation, voids, micro-structure, pedo-features, and relict features.

The most striking morphological characteristics are associated with swelling and shrinking on alternate wetting and drying cycles. Vertisols were believed to be remarkably homogeneous, but recent studies (Wilding et al., 1990) showed that morphological and other characteristics are very complex. The differences in degree of expression of any morphological characteristics reflect the differences in chemical, physical, mineralogical and environmental conditions. In this chapter, the morphological characteristics of Vertisols, without giving any reference to their genesis, are discussed. Because of its importance, soil structure is treated in a separate chapter (Chapter 7) and, therefore, excluded from this chapter. For convenience, this chapter is divided into two sections: macromorphology and micromorphology.

4.2. MACROMORPHOLOGY

The characteristics that are discussed here are: gilgai, nature of cyclic horizons, cracking, surface granular structure, slickensides and sphenoids, color, depth, carbonates and Fe and Mn segregations, clay distribution, and bulk density. These characteristics may not be all present in Vertisols, except for slickensides.

4.2.1. Gilgai

One of the features of Vertisols is the surface configuration (micro-relief) called gilgai (Prescott, 1931; Oakes and Thorp, 1950; Hallsworth et al., 1955); this is an aboriginal Australian term (meaning small waterhole) for the ground surface characterized by knolls and depressions (Hubble et al., 1983). This term is now firmly established in world literature (Soil Survey Staff, 1994). This micro-relief has a repetitive pattern of mounds and depressions, and is a common feature of many clay soils in sub-tropical and tropical areas of the world. The term gilgai is not applicable to micro-topography apparently resulting from freeze–thaw cycles, solifluction, or faunal activity (see Chapter 2). Six gilgai types were recorded by Hallsworth et al. (1955); normal (round) lattice, wavy, tank, stony and melonhole. In addition to these, Thompson and Beckmann (1982) also recognize linear gilgai in Australian soils. The genesis of gilgai is discussed in Chapter 2.

The gilgai feature is not found on all Vertisols. Some Vertisols do not have them, and with others they have been removed by man. Several years, without disturbance, are required for gilgai to develop; man's influence would not be conducive to their formation. It is probable that on the frequently and long term cultivated soils of Europe and much of Asia, their re-emergence has been prevented (Hallsworth and Beckmann, 1969). Gilgai features are reported in Vertisols from U.S.A, Sudan, India, Australia and several other countries. In northeastern Turkey, the Vertisols (Grumusols) on native pastures under about 600 mm of annual rainfall have a distinct gilgai of low humps and knolls or pits of several meters across, with distinct color and depth differences between the two micro-relief elements (Oakes, 1954; Akalan, 1976). This kind of gilgai is found on gentle slopes in which knolls and pits are continuous and form regular lines more or less at right angles to the contour (Dudal, 1965). Spectacular gilgai features are reported in Vertisols from Texas (Wilding and Tessier, 1988).

Gilgai do not include features formed due to frost action or animal activity and their development seems to be dependent, at least in part, on the shrinking, cracking, swelling, and heaving characteristics of particular clay types (Probert et al., 1987). For example, gilgai are not prominent in the semiarid southern India, compared to central India with higher rainfall. Sehgal and Bhattacharjee (1988) reported that gilgai were prominent in Linga, Aroli and Sarol series which occur in areas receiving rainfall >1000 mm per year. They also reported gilgai in Torrerts from the shallow basin of the Mesopotamian plain in Iraq, with a shallow ground water table. Wilding and Tessier (1988) suggest that the degree and frequency of desiccation-rewetting cycles enhances the expression of slickensides and gilgai. The gilgai development in Sudan was localized and infrequent and the incidence of micro-relief was only in the wettest areas (de Vos and Virgo, 1969). No gilgai were so far observed in cold Vertisols (Cryert) from Canada (Mermut et al., 1990).

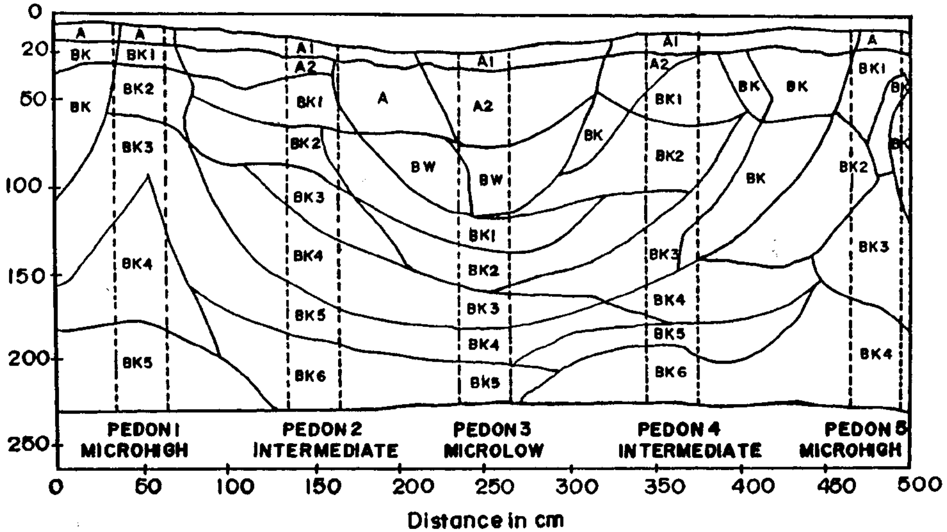


Fig. 4.1. Cross-sectional profile showing the microvariability of Lake Charles clay in Victoria County, Texas (Wilding et al., 1990).

4.2.2. Nature of cyclic horizons

Morphological variations with and without gilgai microtopography are well documented by Beckmann et al. (1971), Spotts (1974), and more recently by Wilding et al. (1990). One of the problems is that horizons are discontinuous. The gilgai complex affects both the soil's physical behavior and the ability to grow crops (Thompson and Beckmann, 1982).

Wilding et al. (1990) have provided an excellent cross sectional profile illustrating the complexity of soil morphology in Vertisols (Fig. 4.1). This presents problems in characterization, sampling and classification of these soils. Striking differences were found in three gilgai positions, namely micro-high, micro-low, and intermediate positions. In the micro-highs, narrow (30–70 cm) "tepee-shaped" structures or diapirs "chimneys" of grayish, calcareous clays extending from the lower Bk horizon to the surface were observed. This portion of soil appears to have been pushed or squeezed up along slickenside planes.

A dark A horizon may be only a few cm thick or even absent on the micro-highs and more than 100 cm thick in the micro-depressions. Organic matter content and depth to carbonates or to a Bk horizon can be equally variable. The micro-lows have little or no carbonate either as nodules, soft segregations or disseminated carbonates and have an A, Bw and Bk horizon sequence. The zone between the micro-lows and micro-highs is generally darker in color than the micro-high, but lighter than the micro-low. The micro-highs have no Bw but a thin A horizon and Bk horizons. Large slickenside planes tend to outline the three gilgai positions

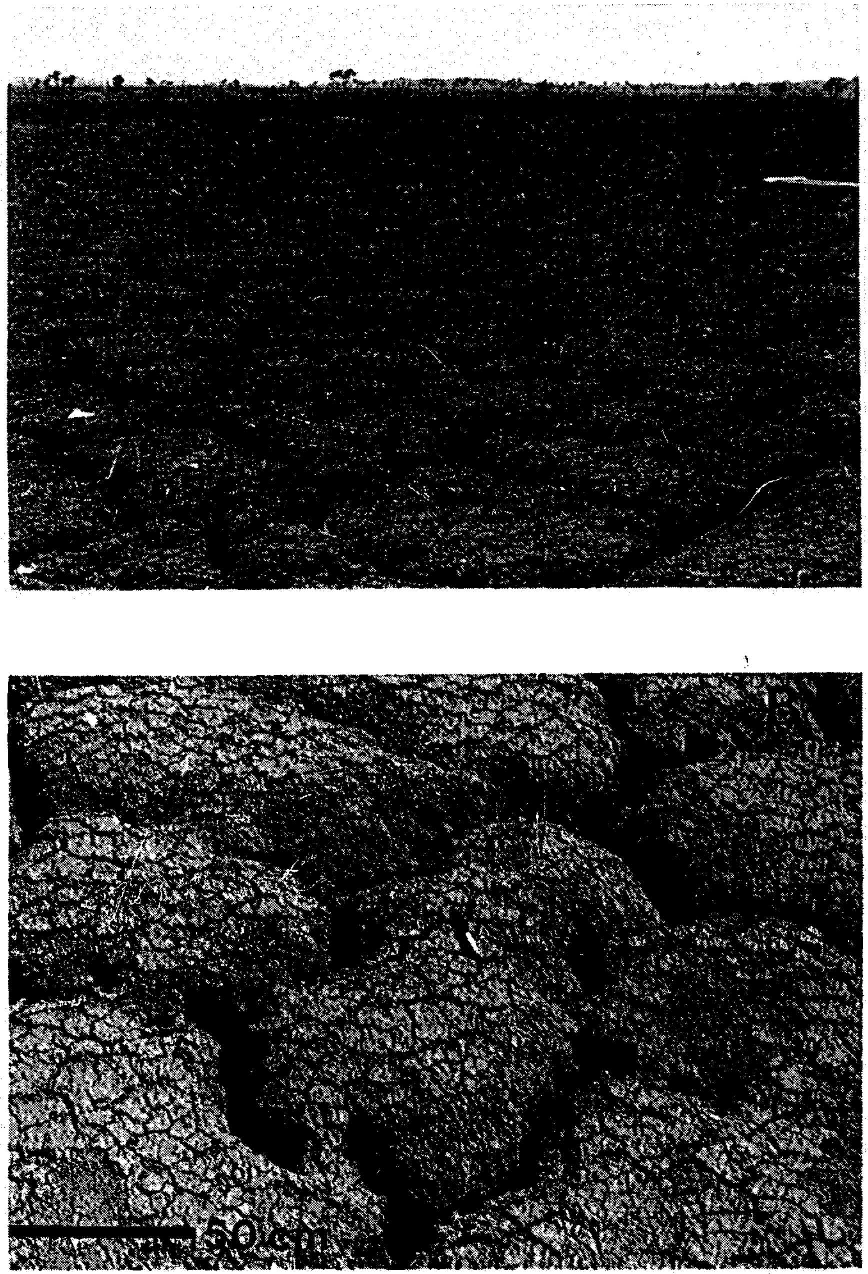


Fig. 4.2. Open cracks: (A) in a Vertisol from Dharwad, India; (B) close up view of cracks, about 5–12 cm wide, showing partial in-filling and formation of thin crust.

as well as the horizons. Wilding et al. (1990) suggested that it is paramount to study close-interval spatial variability so that sampling schemes can be developed for better representation, understanding, and use and management of Vertisols.

4.2.3. Cracking

Cracks are another striking morphological feature of Vertisols (Fig. 4.2) which are used to define this soil order. Partial drying of soil causes the formation of cracks. The open cracks are tortuous and may be 1 cm or more wide at a depth of 50 cm and may extend to a depth of 1 m or more in many Vertisols. Due to cultivation and wetting of the surface (Fig. 4.3), cracks may not be observed in the plow zone, but those below this zone continue to exist. The depth, frequency, size and shape of the cracks are related to the differential moisture status of the cracking zone (El Abidine and Robinson, 1971; Dudal and Eswaran, 1988), and moisture content at the surface and deep in the profile. For example, White (1972) demonstrated that crack outlined polygons were 0.6–0.9 m across when the soil was dried to a depth of 1.0–1.3 m and 1.5–1.8 m across when the subsoil was dried to a depth of 2.5–2.8 m. These crack spacings defined the width of prisms that are described in a vertical section.

The nature of cation-saturation of 2:1 silicate clays tends to affect the size and frequency of cracks. Vertisols in the Ca-saturated clays form wide cracks with

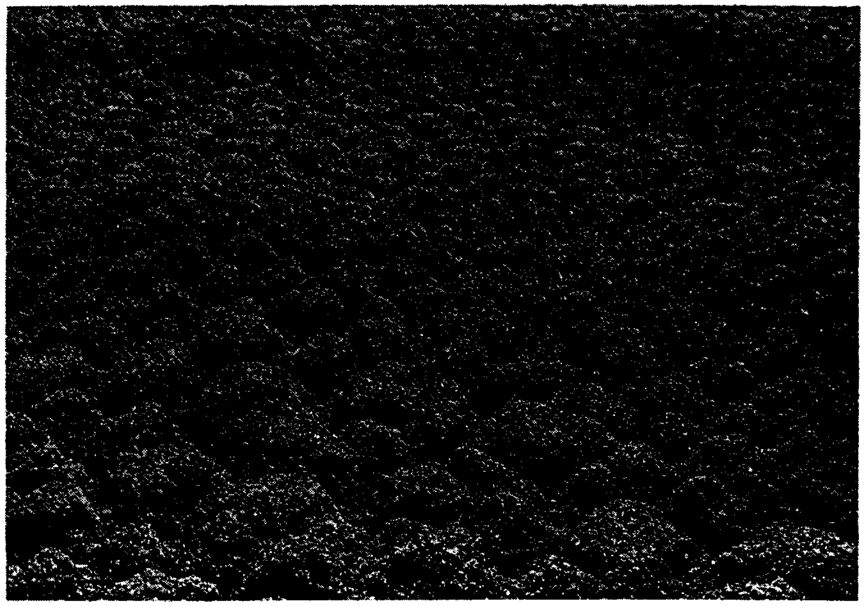


Fig. 4.3. Closed cracks after a short rain shower on a Vertisol, Dharwad, India showing a very interesting pattern of microtopography.

TABLE 4.1

Data on soil cracks in Vertisols from different regions of the world

Sl. No.	Soil/Site	Crop	Crack parameters				Source
			Width	Depth	Spacing	Volume (m^3/ha)	
	Sudan						
1.	GRF	Fallow	4.2	51	28	867	El Abedine and Robinson (1971)
2.	OUH	Natural forest	5.1	42	51	450	
3.	GHAT5	Cotton	3.4	60	39	609	
4.	GHAT7	Fallow	3.7	60	62	392	
	Canada						
1.	Regina	Forage grasses	1.5	34.8	91	82	
2.	Sceptre	Natural grassland	0.9	28.2	70	43	Dasog et al. (1988a)
3.	Sceptre	Wheat	2.0	40.1	111	89	
4.	Tisdale	Wheat	1.6	36.5	154	39	
	Israel						
1.	Measured in 1973 and 1974 at maximum opening					325–287	Yaalon and Kalmar (1978)
	India						
1.	Dharwad	Different crops	1.3	26.9	—	234	Dasog and Shashidhara (1993)

rather low frequency on drying (Smith, 1959; Sleeman, 1963). On the other hand, a higher intensity of fine cracks develops when the clays are Na-saturated.

Characterization of cracks is done as part of the profile description and wider and deeper cracks are emphasized. Systematic measurements of crack width, depth and other parameters for an area greater than that of the pedon is very limited and the techniques used are inconsistent. Soil Survey Staff (1994) requires information on duration of opening of cracks at a specified depth for soil taxonomy, but does not specify their detailed description.

Cracks are formed due to drying of a moist soil when tensile strength exceeds the cohesive strength of the soil (Blokhuis, 1982). Studies on evolution of cracks have revealed two stages in the shrinkage process (Hallaire, 1984). At first, thin cracks (less than 5 mm wide) appear with about 3 cm spacings. Later in the drying process, some of these cracks open wider (more than 1 cm) with about 20 cm spacings while the remaining cracks are partially or even totally closed. On a field scale, it is only the latter types that are measurable. However, progress made in image analysis now provides an opportunity to rapidly quantify, in two dimensions, any type of cracks in Vertisols (Bui and Mermut, 1989; Moran et al., 1989).

Data available on cracks in Vertisols for different climatic regions of the world are limited (Table 4.1). It is obvious that no single parameter can adequately explain cracking intensity. The crack volume per unit area is the best index of the

intensity of cracking as it takes width, depth and spacing of cracks into account. Cracking intensity as measured by a direct sand filling technique utilized by Dasog and Shashidhara (1993) for a Vertisol from Dharwad, India was found closely related to moisture depletion due to evapotranspiration and crop removal. The cracking in fallowed areas was less than in cropped fields. Cracking intensity, in general, is the least in cool temperate areas compared to warm tropical regions (e.g. the Sudan) and intermediate in Mediterranean regions (e.g. Israel) (Table 4.1).

Thompson and Beckmann (1982) observed a network of cracks developed mainly in the depressions during a prolonged drying season in Australia. Cracks were less on the mounds and were apparently much finer and shallower than in depressions. Similar observations were made by Stirk (1954). As suggested by Wilding et al. (1990), cracking patterns, crack closure hysteresis and cracking depths as a function of seasonal soil moisture changes, especially in Vertisols with gilgai microtopography have been little explored, and require further attention.

4.2.4. *Surface granular structure*

In several Vertisols, the surface 2–10 cm layer is a loose mulch consisting of medium to fine granular aggregates (Blokhuis, 1982). It is commonly known that the surface structure of Vertisols ranges from fine granular to massive (Hubble, 1984). There are a few authors who suggest that all Vertisols have finely-aggregated surface layers. The surface mulch provides a fine seed bed and partly or fully fills the cracks. Chapter 7 fully and critically reviews the mechanism of formation of the surface mulch. It is reported that the thickness of the mulch decreases in passing through semiarid to humid areas of Sudan (de Vos and Virgo, 1969; Blokhuis, 1982). In the Sudan, the surface mulch is not well developed where rainfall exceeds 500 mm (Jewitt et al., 1979). Similarly, in India, Sehgal and Bhattacharjee (1988) noted that Vertisols of the semiarid regions had 20–30 mm thick pulverized granular surface mulch.

As defined by Wenke and Grant (1994) mulching is the ability of a dry remolded soil material to form aggregates (<5 mm size) after only a few cycles of drying and gentle wetting. When the soil structure is damaged the self mulching ability facilitates the amelioration of soil structure (Dexter, 1991). Attempts were made to determine self-mulching ability of the soils in different geographic regions by using an index. A laboratory-based numerical index of self-mulching (I_{sm}) developed by Grant and Blackmore (1991) is given below:

$$I_{sm} = (f_{ns} + S/P)^{-1} \ (T/35)$$

where P is the percentage clay released by puddling of natural aggregates, S is the percentage clay released from a <5 g sample by shaking after air drying (45°C, for 24 h) and slaking of the puddled soil, T is the total percentage clay and f_{ns} is the fraction of the puddled and dried soils that, after swelling and slacking for up to several hours, does not fall unaided through a 5 mm sieve in free water. Attempts were made to determine the self-mulching ability of the soils in different geographic regions (Grant and Blackmore, 1991; Grant et al., 1993).

Alfisols. In: L.P. Wilding and R. Puentes (Editors), Vertisols: their Distribution, Properties, Classification and Management. Tech. Mono. No. 18, Texas A&M Printing Center, College Station (TX), pp. 147–164.
Yousif, A.A., Mohamed, H.H.A. and Ericsson, T., 1988. Clay and iron minerals in soils of the clay plains of Central Sudan. J. Soil Sci., 39: 539–548.
Yule, D.F. and Ritchie, J.T., 1980. Soil shrinkage relationships of Texas Vertisols: I. Small cores. Soil Sci. Soc. Amer. J., 44: 1285–1291.

Chapter 6

WATER RELATIONS AND WATER MANAGEMENT OF VERTISOLS

MIROSLAV KUTÍLEK

The knowledge on water behavior in soils as well as successful water management practices are based upon a proper understanding of the soil hydraulic functions and characteristics. These are the soil water retention curve (SWRC), saturated hydraulic conductivity K_S, unsaturated hydraulic conductivity K as a function of the soil water content, or of the soil water potential, and soil water diffusivity as a function of the soil water content. For the solution of the transport of dissolved matter further characteristics are needed. Water at hydrostatic equilibrium, the flow of water and transport of chemical solutions, all of which are realized in the porous space of the soil. I shall start therefore with a discussion on the soil porous system in Vertisols, especially with regard to its dynamics as the reaction to volumetric changes of Vertisols related to the soil water content change. After that, sections on hydrostatics and hydrodynamics in Vertisols follow, especially in relation to dissimilarities from the rigid, non-swelling, non-shrinking soils. Specific features of water management practices are included.

6.1. POROUS SYSTEM OF VERTISOLS

The term pore will denote that part of the soil space which is not filled by the soil solid phase. The shape, size and origin of pores play a role in the detailed classification of the soil porous systems only.

Porosity P is generally defined as the ratio of the volume of pores V_P to the total volume of the soil V_T

$$P = V_P/V_T \tag{6.1}$$

Equation (6.1) is exact for inert soils which do not swell or shrink. In Vertisols it has a principal disadvantage in the non-constancy of the reference total volume of the soil V_T which is dependent upon the water content; V_T decreases with water loss, and increases with absorption of water. This is why the void ratio e is sometimes preferred, especially when we deal with the processes where the water content is changed in time. The void ratio is

$$e = V_P/V_S \tag{6.2}$$

where V_S is the volume of the soil solid phase. Similarly, instead of the volumetric soil water content θ of the rigid soil

$$\theta = V_W/V_T \tag{6.3}$$

the soil water ratio was introduced (Smiles and Rosenthal, 1968)

$$\varepsilon = V_W/V_S \tag{6.4}$$

In both, equations (6.3) and (6.4), V_W is the volume of the soil water. In order to avoid the use of the non-constant term V_T, some authors use for simplicity the gravimetric (mass) soil water content w

$$w = m_W/m_S \tag{6.5}$$

where m_W is the weight of the soil water, m_S is the weight of the oven dried soil. While ε in relation to e shows the degree of filling of the porous system by water, w has lost this advantage. The mutual conversion of various expressions is

$$P = e/(1+e) \quad e = P/(1-P) \quad \theta = w\mu_T \quad \varepsilon = (1+e)\theta \quad \varepsilon = w\mu_S \tag{6.6}$$

where μ_T is the bulk density of the oven dried soil ($= m_S/V_T$) and μ_S is the particle density ($= m_S/V_S$ [M L^{-3}]).

The extent of the change of V_T depends upon the change of the volume of soil water within the soil V_W and further upon the domain of the $V_T(V_W)$ relationship. When we follow the change of the volume of soil from wet to dry soil, we distinguish, see Fig. 6.1:

1. Structural domain, or phase (Yule and Ritchie, 1980).
2. Normal or shrinkage domain (Haines, 1923).
3. Residual domain (Haines, 1923).
4. Zero domain (Haines, 1923).

In the structural domain, the water is lost from very wet soil without a macroscopic change of the volume of the soil. Water is released from the stable macropores, which are stable in shape at this stage. It is assumed that the drained pores are mainly interpedal macropores. In addition to these, there are voids formed by the soil macroedaphon. Some of the big interpedal micropores can participate as well. When we deal with clods or remolded soil material, this phase may not be detected. The structural phase was found on unconfined core samples (McGarry and Malafant, 1987) and in field soils (McIntyre et al., 1982). No cracks are formed owing to the drainage at this phase, see Fig. 6.2 (Novák, 1976).

In the normal (or shrinkage) domain, the change of the soil volume V_T is equal to the change of the water content V_W and $de/d\varepsilon = 1$ and $e = \varepsilon$, if the structural macropores are not considered. All water loss is from the variable volume of pores. The porous system undergoes a gradual change in shape and in distribution patterns of the pores. The lost water is not replaced by air. Cracks are formed. When the soil is wetted, the pores increase in volume. Since there is no air in the pores of the matrix, the absorbed water does not replace air. Air which was

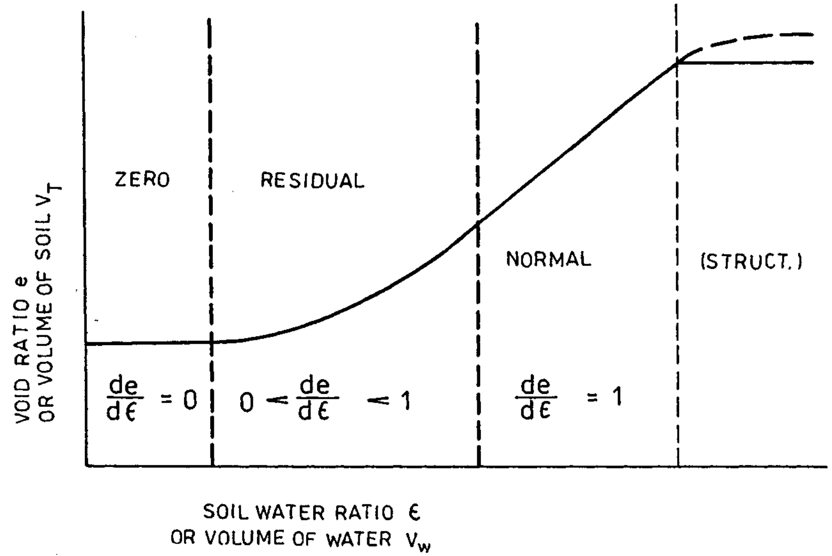


Fig. 6.1. Volumetric change of the soil with the change of the soil water content, drying curve.

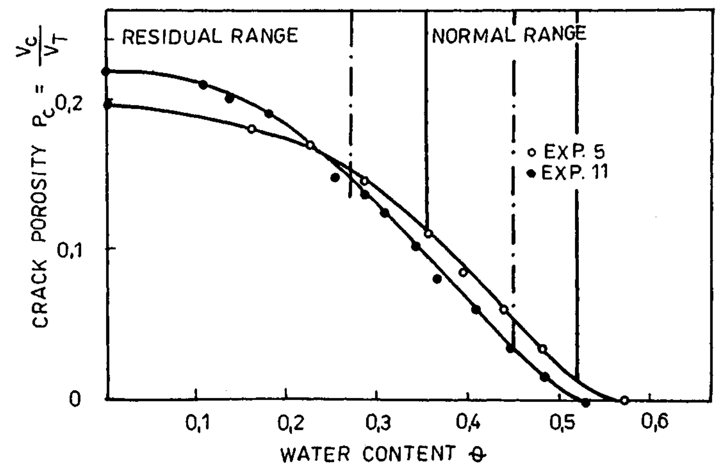


Fig. 6.2. The increase of the crack porosity with the decrease of the water content. Vertisol of the mild zone, Slovakia (Novák, 1976).

contained in the stable macropores of the first phase is not influenced by the volume changes in the matrix during the normal stage.

In the residual domain, the change in the volume of soil V_T is smaller than the change in the water content V_W, therefore $0 < de/d\varepsilon < 1$ and $e > \varepsilon$. When the soil is drained, the porous system is changed but to a lesser extent than in the normal domain. A portion of the water removed is replaced by air. When the soil is wetted, the soil volume increases but still a portion of pores of the matrix is filled by air. The boundary between normal and residual shrinkage is formed by the air entry value of the matrix. Since this term is used in non-shrinking soils at the first penetration of the air into the total volume of the soil (in the representative elementary volume REV (Bear, 1972), the misinterpretation in Vertisols should be avoided. In Vertisols air has penetrated into the structural pores even at the start of the drainage, i.e. before the normal phase has started.

The zero domain is sometimes not included in the classification of the volumetric changes of Vertisols. This domain is characterized by the constant volume of the soil, $V_T = \text{const}$. Therefore $de/d\varepsilon = 0$, $e = \text{const.}$, $e > \varepsilon$. The replacement of air by water is not accompanied by macroscopic volume changes of the soil. Although the porous system does not change on the detectable scale, it does not mean that the porous system is fully unaltered. The swelling or shrinkage has a tendency to continue in the ultramicroscopic scale where it is the shift in the *c*-axis of montmorillonites with the consequence of some reorientation of platty particles. But both processes are probably very restricted due to the envelope pressure. If the zero domain occurs, then it is at a very low soil water content in the range of hygroscopicity.

The volumetric change is described by empirical equations, see e.g. Giraldez et al. (1983) and Tariq and Durnford (1993). When the volumetric change runs the opposite way, i.e. the soil is wetted and swells, the nature of the domains remains generally the same but the actual curve will differ from the shrinkage curve due to hysteresis and thixotropy.

McIntyre et al. (1982) confirm by field experiments the existence of the first three phases: structural, normal and residual in the natural profile of the Vertisol and the sensitivity of the boundaries of those phases upon the value of ESP and upon the ameliorative effect of gypsum.

Normal and residual swelling or shrinkage in the field soil cause rising or sinking of the topographical surface. The upward or downward shift of the soil has been observed at depths greater than 2 m when the surface was ponded by water for a very long period (McIntyre et al., 1982). Bronswijk (1991) converts the vertical soil movement to estimates on volume and water change in field soils.

For a description of the dynamics of the porous systems of Vertisols I shall use the classification of pores based on the hydrodynamic formulation of the transport phenomena. The efficiency as, for example, the accelerated conductance of the pore, is considered as a secondary characteristic. The original system of Corey (1977) is completed as follows:

6.1.1. Submicroscopic pores

Their size is smaller than the mean path of the water molecules. The laws of fluid mechanics are not applicable here. Their volume in soils is negligibly small, usually well below 1 percent of porosity. These pores are important in processes of binding of soil particles.

6.1.2. Capillary pores

The shape of the interface between air and water is determined by the configuration of pores and by forces on the interface water–air, i.e. on the capillary meniscus. The type of flow is taken as laminar. The flow in porous systems containing capillary pores is described by the Darcy equation when all pores are saturated with water, or by the Darcy-Buckingham equation when a part of the pores is not saturated with water.

The Darcian linear relationship between the flux density q and the hydraulic slope I (or potential gradient, respectively) was not confirmed in some instances for clays, when q increased more than proportionally with I. This deviation from Darcy's equation has been explained by the probable action of three factors (Kutílek, 1972):

(1) The viscosity of water close to clay surfaces is assumed to be different from that of bulk water. According to Eyring's molecular model where the viscosity depends upon the activated Gibb's free energy, the first two to five molecular layers have a distinct increased viscosity. Owing to the large value of the specific surface in montmorillonitic clays, the contribution of the first molecular layers to the alteration of the average viscosity may not be negligible. We suppose that with the decrease in the water content the role of water having a different viscosity increases, i.e. the non-Darcian flow may be more distinct (Swartzendruber, 1962, 1963, Kutílek, 1964b).

(2) The coupling of the water, heat, and solute flow may contribute to the non-Darcian effects. The role of the streaming potential in clays is mentioned, too (Gairon and Swartzendruber, 1973).

(3) Clay particles shift and the clay consolidates owing to the imposed potential gradient and flow of water (Kutílek, 1972).

However, deviations from Darcian flow are not frequently observed and Darcy's equation is at least a very good approximation for the solution of practical tasks.

For solution of the unsteady soil hydrological processes, the Richards' equation (RE) is used. For solution of the transport of chemicals, the advection–dispersion equation (ADE) is applied. When the frequency of pores of various equivalent radii is plotted, we obtain a curve similar to the skewed probability density function. Often, the secondary peak is observable (Durner, 1991; Othmer et al., 1991) and porosity is described as bi-modal or even n-modal. We differentiate:

(1) Intrapedal pores inside of peds (aggregates)

They are also described as matrix pores. Their percentage of the total porosity is very high but they are hydraulically less efficient due to their small equivalent radii.

(2) Interpedal pores between the peds

These occupy a small portion of the whole porosity, but they are hydraulically much more efficient than intrapedal pores owing to their large radii. Thus the unsaturated hydraulic conductivity K is by orders higher when a part of the interpedal pores is filled with water than K for the same soil, when the interpedal pores are drained. Interpedal pores form a part of the "preferential ways", also called "by-pass pores". They are responsible for the accelerated transport of dissolved matter and of water as well even under unsaturated conditions. The original cracks, when they are closed after rewetting of the soil, belong often to the category of the interpedal pores (Kutílek,1983). The shape of interpedal pores is frequently planar.

The mutual arrangement of those two categories is hierarchical. In the case of multiple aggregation from primary to tertiary peds, the number of groups of interpedal pores can increase up to three. The hierarchical arrangement is still kept and reflected by the unsaturated conductivity–water content function, see Durner (1991). The soil water retention curve as well as the whole unsaturated conductivity is constructed on the principle of superposition (Durner et al., 1991: Othmer et al., 1991).

The system of micropores depends predominantly upon the development of the soil structure. Peds are separated from the adjoining peds by natural surfaces of weakness. They include simple or compound concentration coatings named cutans. The plasma arrangement in cutans shows different internal architecture compared to the matrix of peds and the bulk density in cutans is higher than that of matrix of peds. The hydraulic conductivity and the diffusion coefficient of cutans are lower than in the internal matrix of peds (Gerber et al., 1974). The size of peds increases with the depth while the strength of the structural development decreases with the depth. In the whole profile of Vertisols we find therefore interpedal pores along peds. They form a three-dimensional net of by-pass pores and the porous system cannot be taken as homogeneous, see the review of Wilding and Hallmark (1984).

6.1.3. *Macropores (non-capillary pores)*

They have such a size that the formation of capillary menisci is excluded accross the pore. The air/water interface is not influenced by capillary forces. The flow of water inside of such a pore can be either in the form of a film on the walls with all irregularities induced by the roughness and by the shape of the wall. Or, the flow is realized throughout the whole cross-sectional profile of the pore. The macroscopic flow (flux density) is not described by the Richards' equation. Macropores are either tubular or planar. The tubular shape is mainly due to the action of soil macroedaphon and the walls are then fixed by the organic cutans. Consequently, such pores are relatively stable when the soil water content is changed. Their drainage usually does not cause a volumetric change in the soil. Planar pores have predominantly the form of cracks and fissures. They originate due to volume changes in the soil when the soil dries out. When the soil is wetted, the great majority of cracks cease to exist and they are transformed into micropores of similar character as the interpedal pores. A more detailed review on the description of macropores can be found in Bouma et al. (1977), or in Beven and

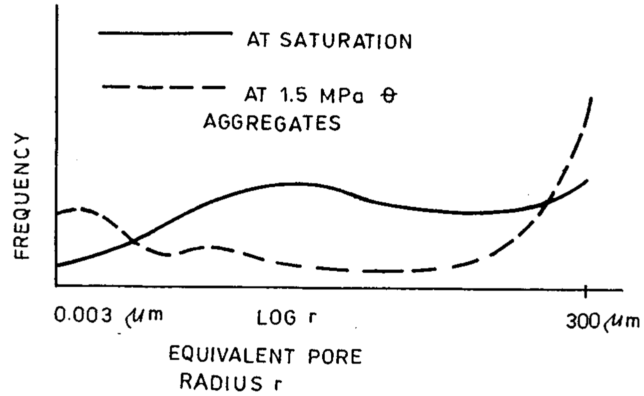


Fig. 6.3. Pore size (equivalent radius) distribution in Vertisol aggregates at the saturation with water and at water content corresponding to 1.5 MPa (wilting point). Derived from Schweikle's (1982) data.

Germann (1981). The boundary between the micropores and macropores is arbitrarily chosen in the literature with vague criteria. For practical engineering use as well as for theoretical purposes, the division is where the capillarity ends. The proposal of Beven and Germann (1981) is very acceptable. The macropores start at the pressure head $h > -1$ cm which corresponds to a width of 3 mm.

The cracks are not necessarily only vertical. Bui and Mermut (1988) have reviewed observations on substantial deviations of orientation of cracks from the vertical. The orientation of cracks is strongly anisotropic in all soils studied, originating in a broad zone from the semiarid to mild climatic regions.

The patterns of the cracks are disturbed by cultivation and so is the original soil structure. If the soils are cultivated when wet, the original structure is strongly damaged. The hydraulic characteristics of the matrix are substantially altered and the patterns of cracks are also different from the original ones. The hydraulic characteristics obtained on virgin soils can not be reliably applied on soils after they have been included in the irrigation and cultivation.

One of the most striking differences between the Vertisols and rigid soils is a distinct non-constancy of the porous systems of Vertisols. Their porous system undergoes quantitative changes due to drying or wetting. First, there is the change in orientation of platty soil particles and the collapse of the space arrangement of those particles, when the soil is drained (Schweikle, 1982). The original isotropic character of the internal architecture is gradually transformed to an anisotropic one. Due to the origin of local stresses either by swelling or by shrinkage, there is a typical reduction of some subcategories of pores, while the frequency of other subcategories increases.

The distinct difference in the pore size distribution in wet and dry Vertisol is demonstrated in Fig. 6.3. The graph was derived from the summation curves of pores measured by mercury porometry on wet aggregates and on 1.5 MPa dry

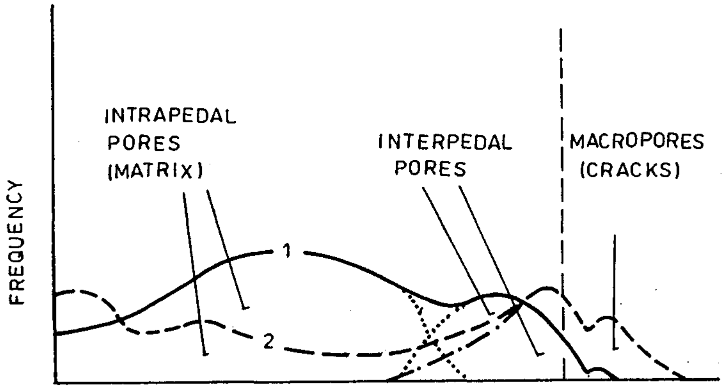


Fig. 6.4. Schematic pore size distribution in REV (representative elementary volume) of the Vertisol at the full saturation with water (1) and at 1.5 MPa water content (2). REV includes the cracks.

aggregates by Schweikle (1982). When the soil is saturated with water, the pore size distribution is near to uniform with slight increase of frequency of coarse pores and a decrease of content of ultrafine pores. The only one weak peak is at medium size pores. When the soil is drained up to 1.5 MPa, the pore size distribution changes substantially: the frequency of medium pores is at a minimum. The shrinkage has reduced this category of pores. Coarse pores have very high freqency, a majority of medium pores was transformed to coarse pores. The continuous increase of frequency with increase of the size of pores indicates a continuous transition of the microporous system to macropores. In the ranges of fine and ultrafine pores, two secondary peaks develop, both due to a reduction of medium pores.

When the principle of the representative elementary volume (REV) is applied to the studied volume of soil, we shall obtain distinct multimodal porous systems, schematically presented in Fig. 6.4. The cutans on the aggregates of Vertisols are a result of the permanent existence of the separation walls of aggregates and interpedal pores are distinct. The cracks appear even at the same position (Virgo, 1981, Kutílek, 1983). When the soil is wetted, the cracks disappear from the category of macropores either totally or at least in majority of instances. They are transformed mainly into pores of a similar hydraulic character as the interpedal pores. After drying, they appear again as cracks-macropores. The original interpedal pores remain partly in the category of capillary pores, partly they are enlarged so that they now belong to the category of macropores, too. It follows that the porous system is highly variable, dependent upon the water content and that it can not be simply derived just from the soil water retention curves, as is usually done in rigid soils. The alterations in the porous system are more pronounced in Vertisols affected by alkalization processes, when ESP is high and the concentration of the soil solution is kept low.

Up to now we have discussed the porous systems of unconfined soils. In confinement, practically the overburden pressure reduces the changes in the porous system during either wetting or drying of Vertisols. With the increase in depth, the overburden pressure increases, the volumetric changes decrease and they disappear practically at a certain depth. When the soil is dry, the density and size of cracks decrease with the depth. At a certain depth the cracks cease to exist. We can assume that the width of the cracks decreases exponentially with depth. When we measure the area of cracks A_{co} in the representative area A_T on the surface, we can express the volume of cracks V_c (Kutílek and Novák,1976)

$$V_c = \int_0^\infty A_{co} \exp(-\alpha z)\, dz = A_{co}/\alpha \tag{6.7}$$

The empirical coefficient α is in the range between 0.01 and 0.1 cm^{-1}. Its value is obtained by two replicas of direct measurements of the width of cracks or area of cracks, respectively, once on the topographical surface, once at the defined depth, and α is obtained from the exponential decay either of the width, or of A_{co}. The crack porosity P_c is then computed as the ratio of V_c to the REV. The depth of REV is taken as the depth of the macroscopically observable decay of cracks.

In the majority of reports, however, the decrease in the width of cracks with depth is not considered and the data on the volume of cracks are computed according to the Delesse–Rosiwal principle which states in modification for crack porosity $P_c = V_c/V_T = A_{co}/A_T$ where A_T is the total area on which the cracks are mapped. The layer thickness for which the estimate is computed is rarely defined. For mapping of the net of cracks and for the determination of their area and of the width of cracks a computerized image analysis is used (e.g. Ringrose-Voase and Bullock, 1984).

The crack porosity increases not linearly with the decrease in the water content and Novák (1976) shows its sigmoidal functional shape for two locations of Vertisols, see Fig. 6.2. The close relationship to the volume change in Fig. 6.1 is distinct. The relationship can be described by an equation similar to the soil water retention curve.

The dynamics of the formation of the system of cracks is not yet fully clear. In semiarid conditions, cracking was described as a gradual decrease in spacing of cracks in time, i.e. as the gradual increase in the density of the cracks together with increase in their width and depth (Yalon and Kalmar, 1984). In the mild zone, Hallaire (1984) found that the cracking process proceeds under the vegetation in two stages: in the first relatively wet stage in the late dry spring, thin cracks of width <2 mm develop in a dense net, the width of cracks is nearly constant and the crack porosity is <1 percent. In the second stage, when the dry period continues in summer, some of the cracks develop further on up to a width >1 cm. The system of cracks on the surface consists of two nets: (1) big cracks of width about 1 cm, (2) fine cracks of width <0.2 cm. The total volume of cracks is up to 4% but the density of all cracks has decreased when compared to the first stage.

TABLE 6.1

	(A)	(M)
Maximum width of cracks	9.5 cm (8)	3 cm (6)
Mean width of cracks	3.5 cm (8)	0.3 cm dry (5) 0.1 cm wet (5) 0.1 cm dry (7)
Distance between cracks at initial stage	50 cm (8)	
at final stage	<20 cm (8)	10 cm (5) 20–100 cm (4)
Length of cracks per area	2.8 cm cm^{-2} (8)	
Mean density of cracks	3.3 cracks m^{-1} (8)	
Surface of cracks related to the topographical surface		8.3 (4)
Depth of cracks max.	80 cm (8)	100 cm (4), (5) 40 cm (6)
mean	<40 cm (8) 80 cm (2)	100 cm (2)
Crack porosity	3 per cent (8) 2.5 per cent–5 per cent (9)	1.4 per cent–2.4 per cent (1) 1 per cent–4 per cent (3) 9 per cent dry, 3 per cent wet (5) 9 per cent dry, 4 per cent wet (7) <4 per cent (4)

Authors: (1) Bouma and Wösten (1979), (2) Bui and Mermut (1988), (3) Germann and Beven (1981), (4) Hallaire (1984), (5) Jarvis and Leeds-Harrison (1987), (6) Novák (1976), (7) Ringrose-Voase and Bullock (1984), (8) Yalon and Kalmar (1984), (9) Zrubec (1976).

Opposite to this observation are the data of Ringrose-Voase and Bullock (1984) where the width of cracks has a tendency to log-normal distribution. Raats (1984) described the crack formation by means of mechanics and his paper offers a review of earlier theoretical studies.

Some observed data on crack development in Vertisols by various authors are summarized in Table 6.1. (A) is for arid and semiarid zones, (M) for the mild zone.

6.2. SOIL WATER RETENTION CURVES

When we describe and compute the flow of water in unsaturated soils, we describe the driving force of the process as the gradient of the total potential. For practical reasons the potential is expressed in units of energy [J] per unit weight

of water [N] and the dimension is [L], the convential unit is cm. The moisture, or matrix potential is then the pressure head h of water and in the unsaturated soil $h < 0$. A good working approximation of the total potential H for the rigid non-saline soils is $H = h + z$, where z is the vertical coordinate positive in the upward direction, or z is the gravitational potential expressed in [L]. The soil air pressure is assumed equal to the atmospheric pressure. In rigid soils, the relation between h and the volumetric soil water content is unique for the given soil and for the direction, how the equilibrium has been reached. This relation is called either the soil water retention curve (SWRC) or, sometimes, soil water characteristic. When h is plotted on a logarithmic scale, the term $pF = \log h$ is introduced and this graphical presentation of SWRC is called the pF curve as well.

In Vertisols the swelling pressure develops when the soil is wetted. It can be demonstrated as the pressure which should be applied in order to prevent the soil from swelling. In terms of the potential, it is part of the envelope potential. In addition to swelling pressure, the envelope potential includes all acting external pressures such as, for example, the load on the soil surface.

In the following discussion, I shall consider the confinement and unconfinement of soil in one direction of the vertical z axis only and swelling–shrinkage will be restricted to one direction only instead of describing them as a three-dimensional phenomenon. In spite of the approximate approach, the model provides evidence on specific features of the SWRC of Vertisols.

Let us suppose that two soil core samples are provided with tensiometers, one sample is firmly covered by perforated lids or semipermeable plates, the other one is left free on the top. The first one is confined, the second one is unconfined. When both samples are wetted from the bottom in the same way, the tensiometer reading will be different at equilibrium. The difference is the envelope (or overburden) potential, here read as the envelope pressure head h_e. The tensiometer reading is $h_t = h + h_e$. The importance of h_e increases with increase of the content of clay particles in the soil. In sandy soils the contribution of h_e is usually negligible since the external pressure is predominantly transmitted on direct contact of the grains, and the pore water is intact by the external pressure. When water forms films around the particles and the pressure is transmitted from particle to particle via the water films, the pore water pressure is greatly influenced by the external pressure and the value of h_e is then significant. In the field Vertisols, h_e is dependent upon the weight of the soil column above the point of measurement. If there was some additional load on the surface as for example, the pressure of wheels of machinery, it is added to the overburden pressure of the upper soil layers. The envelope pressure head is then

$$h_e = \alpha/\mu_w g \ (p_s + \int_0^z \mu g \, dz \qquad (6.8)$$

where $\mu_v = (m_S + m_W)/V_T$, p_s is the additional pressure on the surface, m_S is the mass of dry soil, m_w is the mass of soil water, g is the gravitational acceleration, μ_w is the density of water and α is the consolidation characteristic. In inert soils, such as consolidated sands, all load is transferred on contact of the solid particles and $\alpha = 0$. In unconsolidated fully swelling unconfined soils just on the surface, the change in the water content results fully in a change in the volume, i.e. in the uplift of the surface, provided that the soil below is consolidated and not compressed. Then, $\alpha = 1$. Between these two extremes the transitional situation exists with (Groenevelt and Bolt, 1972)

$$\alpha = 1/h_e \int_0^{h_e} (\partial e/\partial \varepsilon)_{he} \, dh_e \tag{6.9}$$

If the influence of h_e upon α is neglected we obtain equation (6.8) with $\alpha = de/d\varepsilon$, i.e. with the slope of the curve in Fig. 6.1 (Philip, 1969)

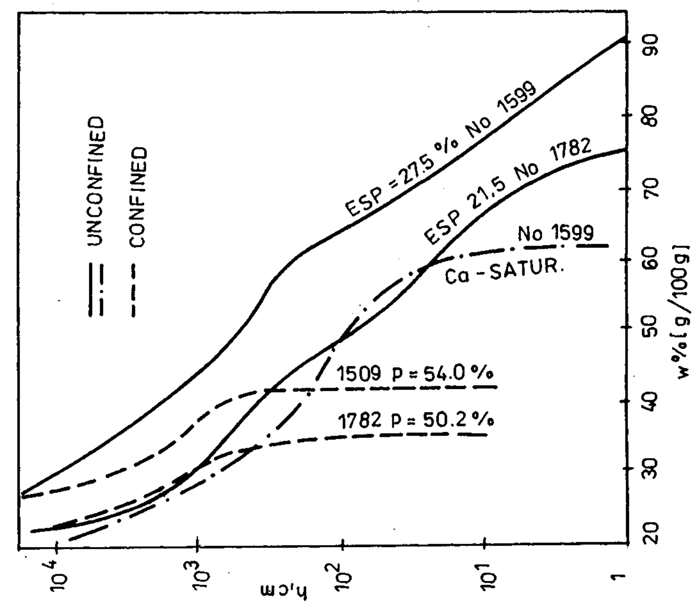
$$h_e = de/d\varepsilon[p_s + \int_0^z \mu_v g \, dz] 1/\mu_w g \tag{6.10}$$

In the majority of practical instances in preconsolidated Vertisols $\alpha = 0.2$–0.4. With the increase in the ESP the value of α also increases, and the increase is more pronounced when the concentration of soluble salts is kept low. The alkali Vertisols with a high ESP are a good material for modeling of the extreme action of swelling upon the hydraulic properties. The SWRC on unconfined and confined samples of Vertisols of different ESP and with EC<1 mmho/cm are shown in Fig. 6.5. The curves were obtained on the pressure plate apparatus, the applied pressure head h is in equilibrium with the gravimetric water content w. In confined samples we should write h_t since the confinement of the sample causes the envelope pressure which is part of h_t. The envelope pressure head was not measured. In the unconfined samples, the increase of ESP results in a more monotonous curve close to a straight line on the semi-log paper. Thus for unconfined Vertisols with a high value of ESP, the retention curve is expressed by

$$h = [\exp(-aw)]b \tag{6.11}$$

with the empirical parameters a and b. If the ESP is low and for all confined samples, the van Genuchten (1980) equation is applicable when the effective volumetric soil water content is replaced by w/w_S, the gravimetric water content is in parametric form where w_s is gravimetric water content at $h_t = 0$. Contrary to general expectations, the confined SWRC keeps the values below the unconfined SWRC if ESP is high. When the soil is saturated by calcium, the unconfined SWRC is roughly identical with the confined soil after the porosities of both start to be identical in the drainage. The set of unconfined SWRC represents the SWRC at the topographic surface. The confined SWRC are those where the overburden pressure of the topsoil reduces both the swelling and the maximum attainable porosity (Kutílek and Semotán, 1975).

When the envelope pressure head h_e is measured together with the tensiometer head h_t, we obtain a set of SWRC in the three-dimensional plane of $\varepsilon(h_t, h_e)$,



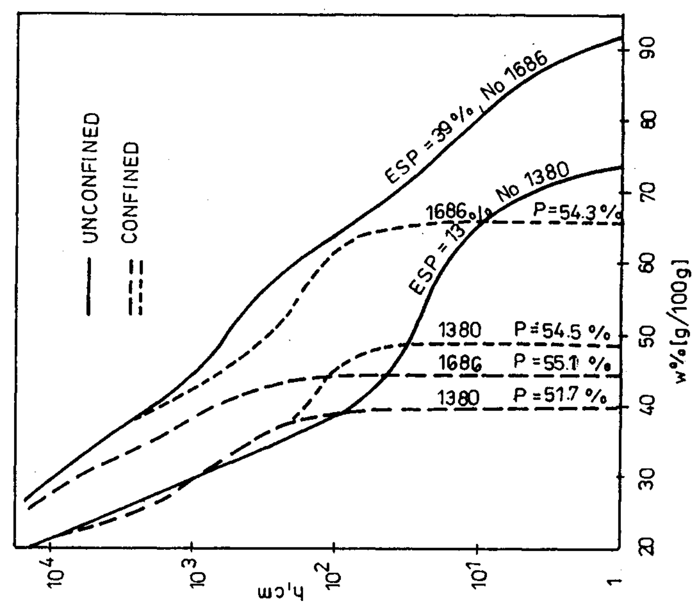


Fig. 6.5. Soil water retention curves of unconfined (full line) and confined (dashed line) samples of Vertisols with increasing ESP (exchangeable sodium percentage) and of one sample originally alkali and then saturated with Ca (dot and dash line) according to Kutílek and Semotán (1975). Vertisols from Gezira, Sudan. P is porosity of core sample, w water content by weight, h is potential, here the applied pressure in pressure plate apparatus.

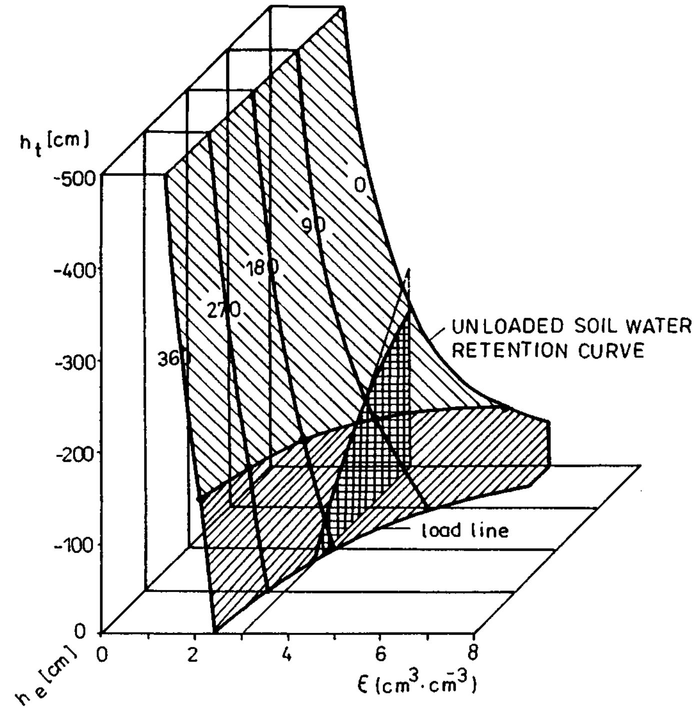


Fig. 6.6. Three-dimensional representation of a hypothetical family of soil water retention curves with gradual increase of the envelope potential h_e, the tensiometer potential is h_t (Stroosnider and Bolt, 1984).

see Fig. 6.6 (Stroosnijder and Bolt, 1984). A special case is $\varepsilon(h_e)$ when $h_t = 0$, the curve is called the load line and it is frequently determined in civil engineering as a consolidation curve. It shows the change of ε when the load is changed and the water is free to flow out to the pool with free water, i.e. when $h_t = 0$.

To the less important differences between the SWRC in rigid soils and in Vertisols, especially in Na-Vertisols, belongs the thixotropic behavior of soil particles which is then reflected by the thixotropic behavior of the SWRC. The tensiometer potential h_t depends upon the time which has elapsed from the last mixing or disturbance of the soil. Due to the disturbance, the internal arrangement of particles is destroyed and the tensiometer potential rises. Then, with time, the structural reorientation of particles is gradually gained again and the tensiometer potential drops. The disturbance is also due to drying. When the soil is then wetted, the swelling process hides the thixotropic effects. The family of thixotropic SWRC is demonstrated schematically in Fig. 6.7. A practically observable thixotropic effect could be found just on the soil surface. The overburden pressure restricts the possibilities for reorientation of particles and thus a couple of centimeters below the surface the thixotropy is restricted.

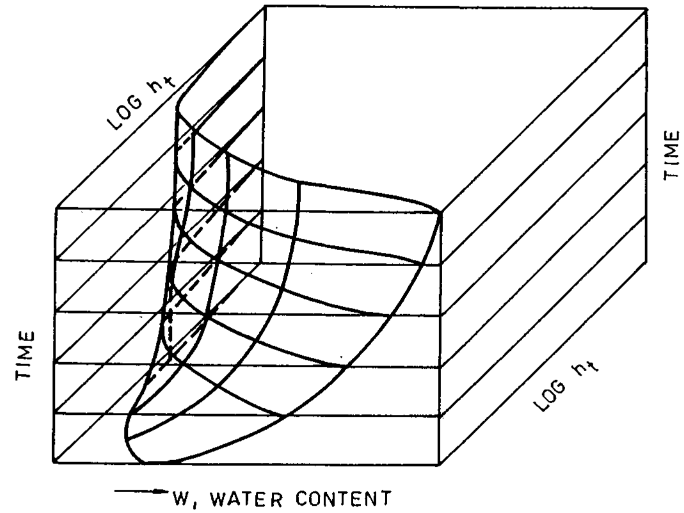


Fig. 6.7. Schematic presentation of the thixotropic behavior of the soil water retention curve in swelling clay (Kutílek, 1964a). The tensiometer potential h_t is read as absolute values.

The theoretical treatment of the soil swelling/shrinkage on a microscopic scale is given by Iwata et al. (1988). Such discussion is excellent for proper physical interpretation of our macroscopic observations.

6.3. HYDRODYNAMICS IN VERTISOLS

Vertisols are soils where two main porous systems exist simultaneously: the system of macropores and the system of micropores. The pore size distribution in both is variable, dependent upon the soil water content and the same is for the mutual ratio of the total volumes of pores. In the microporous system, the modified Darcy–Buckingham and Richards equations are applicable. In the system of macropores, the wetting process is described in a different way and the vapor flux in macropores in the drying process is diffusion. For the sake of lucidity, we shall first discuss the flow of water in each system separately and, mainly, one-dimensionally. The mutual interdependence including the opening and closing of the cracks has not yet been satisfactorily formulated in quantitative terms and as a physical description.

6.3.1. Flow in the microporous system of Vertisols

The term saturated conductivity K_S has not an exact meaning with regard to the normal range of shrinkage/swelling, see Fig. 6.1. The soil matrix is actually saturated in a wide range of water content. The theoretical development of the functional relationship $K(\varepsilon)$ will be shown later. Now, let us assume that K_S is related to $h = 0$ in unconfined and to $h_t = 0$ in confined samples. The dependence

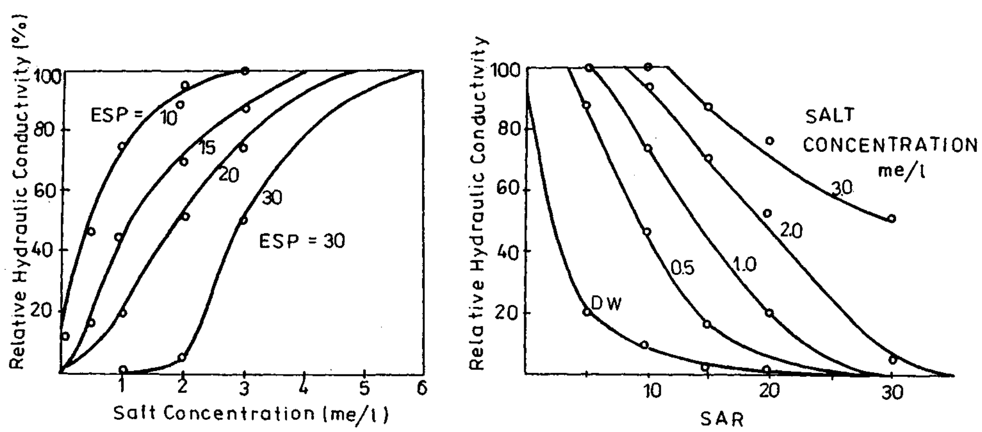


Fig. 6.8 The effect of the variation of ESP and of the concentration of the soil solution upon the relative hydraulic conductivity, DW is for distilled water, values after reaching steady conditions are plotted (Shainberg et al., 1981).

of K_S upon the ESP and upon the salt concentration expressed by EC was studied in detail on soil–sand mixtures by Shainberg et al. (1981), see Figs 6.8 (left) and 6.8 (right). Their conclusions were confirmed qualitatively on Vertisols by many authors, but the sets of data in Vertisols were less systematic. The self-mulched soil from the surface of some Vertisols did not follow the scheme of Shainberg. There, the tendencies exist, but the change in the conductivity is much less pronounced (Kutílek, 1983).

With the increase in the ESP or with increase in the sodium adsorption ratio (SAR) and with the decrease in the electrical conductivity EC, the double layer is extended and the value of the zeta potential increases. The induced swelling results in reduction of both medium and large micropores. In addition to this, the dispersion of microaggregates and of floccules leads to microtransport of individual clay particles and to clogging of some pores. The hydraulic conductivity is decreased by a combination of the two effects. Just the increase of EC when other characteristics are kept constant leads to a compression of the double layer, the floccules are more stable and K_S is increased. The physical interpretation has been confirmed by model experiments in laboratory conditions (van Olphen, 1977) and by theory (Iwata et al., 1988). Theoretically, the effects should be reversible.

Shainberg et al. (1981) show that the K_S changes can have a seasonal character in semiarid conditions when irrigation is practised. The irrigation water has usually a relatively high concentration of salts and K_S is kept relatively high. After the irrigation season when rains start, the rain water dilutes the soil water in the upper part of the profile. If ESP is high, K_S drops significantly. This explains the extensive flooding of the surface of Vertisols in the Sahel region after heavy rain.

The effects of salinization and alkalization are further modified by the overburden pressure. When K_S was measured on unconfined and confined core samples, the confinement to bulk density of about 1.3 g cm^{-3} leads to a decrease

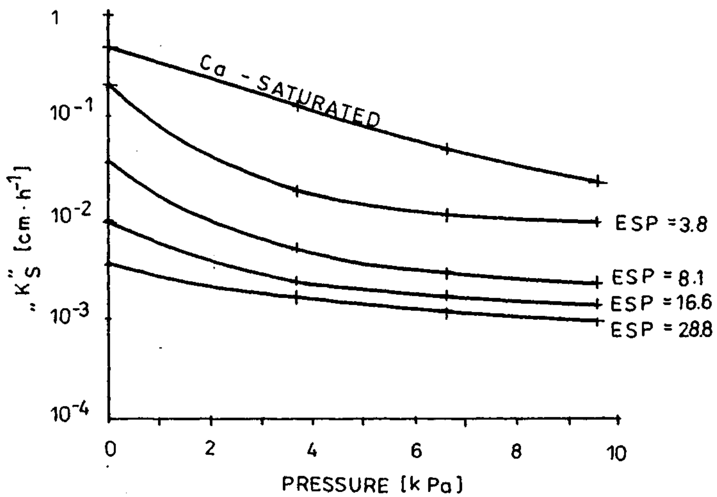


Fig. 6.9. Abdine's (1971) data on K_S as affected by ESP and by the overburden pressure, Vertisols from semiarid zone (quotation in Farbrother, 1987).

in the conductivity of one order (Kutílek and Semotán, 1975). Farbrother (1987) quotes the Abdine (1971) data on K_S measured for Gezira soil samples when external pressure was applied to the soil. The values of pressure used were equivalent to the overburden pressure at depths of 20 cm, 35 cm and 50 cm, respectively. The data are plotted in Fig. 6.9. The lower the ESP, the higher is the difference in K_S between unloaded and loaded samples. When ESP is high, the large micropores are strongly restricted even in unloaded soil. The external pressure has either a limited or no chance to liquidate them. In Vertisols of low ESP, the increase in the external pressure may decrease the volume of large micropores and the decrease in conductivity is more pronounced.

The self-mulched soil from the surface has K_S of one to two orders higher than the underlying soil in the Sudanese Gezira. In ESP as high as 13 percent does not reduce K_S. However, the external pressure and confinement reduces K_S more than in the underlying unmulched soil (Kutílek, 1983). The confinement can be realized only on that portion of self-mulched material which falls in the cracks. Even there, its conductivity is substantially higher than in the neighboring confined soil.

The existence of n-modal porosity has already been discussed. However, its influence upon the character of the $K(w)$ or $K(\varepsilon)$ function in Vertisols has not yet been treated.

Due to volumetric changes in Vertisols accompanying the unsteady flow of water and owing to the consequent alteration of the hydraulic characteristics of the matrix, the equations developed originally for inert soils should be modified. The theoretical treatment I am presenting here is just the starting point to the solutions. As for the introduction, the one-dimensional formulations will do. The complete theory of three-dimensional unsteady flow processes together with soil volume change and dynamics of the cracks has not yet been developed.

When the soil is not stable during unsteady flow, Darcy's law is modified according to Gersevanov (1937) and the flow rate of water is related to the solid phase and a material coordinate system is introduced: Euler's coordinate system is replaced by the Lagrange system. For one-dimensional flow and swelling, the new material coordinate is defined as follows (Smiles and Rosenthal, 1968): the ratio of the material coordinate m to the Eulerian coordinate z equals the ratio of the volume of the solid phase to the total volume of the soil

$$\mathrm{d}m/\mathrm{d}z = 1/(1+e) \tag{6.12}$$

After integration

$$m = \int_{-\infty}^{z} \mathrm{d}z/(1+e) \tag{6.13}$$

or, with the porosity P

$$m = \int_{-\infty}^{z} (1-P)\,\mathrm{d}z \tag{6.14}$$

The physical meaning of Equation (6.14) can be demonstrated by the infiltration experiment into the dry soil column provided by walls on the sides. The swelling can occur only in the vertical direction. At time $t = 0$, $z = 0$, identical with the soil surface, the z coordinate is positive downward. When water infiltrates, the soil surface rises since the soil swells. The wet swollen surface is at $z < 0$ and according to Equation (6.14) it is at $m = 0$. Between $m = 0$ and the depth of the wetting front is $m > z$ and the mutual relationship is non-linear. Below the zone of the wetting front, there is no change of porosity due to the infiltration and the relationship between m and z is linear. Darcy's equation in the Lagrangian coordinate system is

$$q_\mathrm{m} = -K_\mathrm{m}(\varepsilon)\ \mathrm{d}H/\mathrm{d}m \tag{6.15}$$

where index m denotes the parameters in the Lagrangian coordinate system, q_m is the flux density (flow rate of water) H is the total potential expressed as the pressure head, $K_\mathrm{m}(\varepsilon)$ is hydraulic conductivity. Assuming $K_\mathrm{m}/\mathrm{K} = \mathrm{d}m/\mathrm{d}z$ and with Equation (6.12) is for normal range of swelling or shrinkage (Smiles and Rosenthal, 1968)

$$K_\mathrm{m}(\varepsilon) = K(\varepsilon)/(1+e) \tag{6.16}$$

The equation of continuity is

$$-\,\partial\varepsilon/\partial t = \partial q_\mathrm{m}/\partial m \tag{6.17}$$

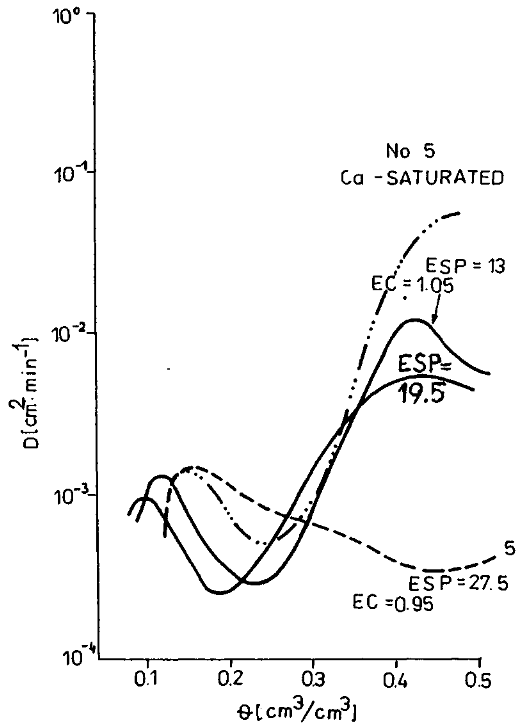


Fig. 6.10. The dependence of the soil water diffusivity D upon the water content θ in confined Vertisols from Gezira, Sudan. Soils at various ESP (Kutílek, 1973, 1983).

and by combination of Equations (6.15) and (6.17) is

$$\partial\varepsilon/\partial t = \partial/\partial m\ [K_\mathrm{m}(\varepsilon)\ \partial H/\partial m] \tag{6.18}$$

In a similar way

$$D(\varepsilon) = K_\mathrm{m}(\varepsilon)\ \mathrm{d}h_t/\mathrm{d}\varepsilon \tag{6.19}$$

If we measure $K(w)$ or $D(w)$, the transformation to $K(\varepsilon)$ is through a constant according to Equation (6.6). When the consequence of unsteady flow is the transition from the normal to the residual range of the volumetric changes, the transformation of the equations will be more complicated.

Sposito and Giraldez (1976) developed for infiltration into an unconfined matrix an equation formally similar to the Richards equation. For its solution, they proposed a new material coordinate transformation.

When the soil is confined, which is an approximate case of the subsoil, we can neglect the slight volumetric change. In alkali Vertisols, the diffusivity dependence upon the soil water content decreases with increase in ESP and it was near to constant when ESP > 20 percent and for ESP = 27.5 percent it was found even decreasing with increase of the soil water content, see Fig. 6.10. In the last case,

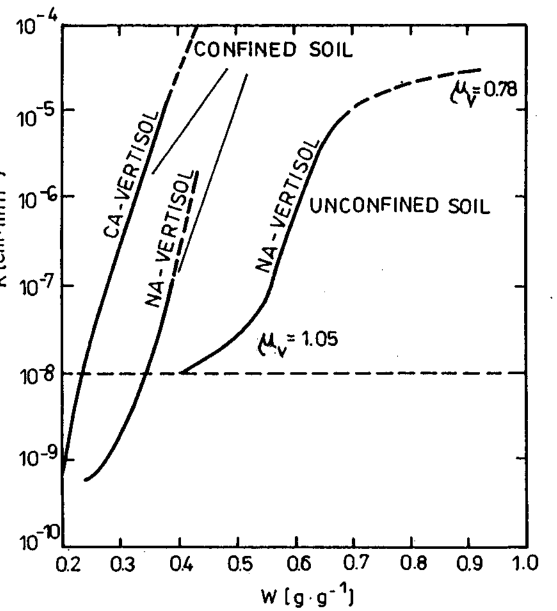


Fig. 6.11. Effect of the cation saturation upon the unsaturated hydraulic conductivity function (Kutílek, 1973).

the value of D was close to the maximum water vapor diffusivity in a dry matrix. The decrease of D with increase of ε was found also for swelling materials by Smiles (1976). Detailed analysis shows that D = const. is a good working approximation for all Vertisols under salinization hazard (Kutílek, 1984).

The substantial difference in K between Ca-saturated Vertisol and Vertisol of high ESP is maintained when the water content decreases, see Fig. 6.11. The character of the functional relationship $K(w)$ is different for the two different ionic saturations. The increase in salinity of the pore water partly reduces the discussed extremely low values of K due to ESP in a similar way as was found for the K_S.

6.3.2. Flow in cracks of Vertisols

It was recognized early that the cracks are the predominant and frequently the only channels available for deep moistening of Vertisols. Figure 6.12 illustrates well the deep moistening of the Vertisol in the Sudanese Gezira due to the rapid flux of water through the system of cracks. Sometimes a secondary peak of the soil water content at a depth of about 50–60 cm is reported in addition to the full wetting of the topsoil. The maximum depth of moistening is identical with the depth of the system of cracks.

When the ponding time is estimated from the rain infiltration, the first approximation is the equality of the cumulative rain and of the volume of cracks.

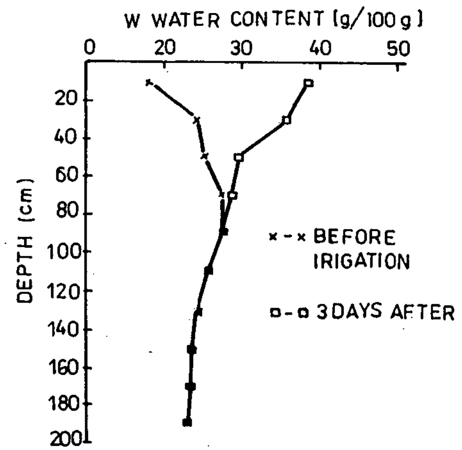


Fig. 6.12. Soil water content in cracked Vertisol before and after irrigation (Farbrother, 1987).

When we divide V_c in Equation (6.7) by the representative elementary area A, we obtain the cumulative rain at the ponding time. The rate of inflow into the matrix is here neglected and the rate of inflow into the cracks is taken as infinite. The ponding time is obtained from

$$V_c/A = \int_0^{t_p} v_r(t)\,dt \tag{6.20}$$

where v_r is the rain intensity. When absorption through the walls of cracks is considered, too, we estimate the area of walls of cracks similarly to Equation (6.7):

$$\int_0^{\infty} 2\,l_{co}\exp(-\alpha z)\,dz = 2\,l_{co}/\alpha \tag{6.21}$$

where l_{co} is the length of the net of cracks on the representative elementary area and α is identical with α from Equation (6.7). When we neglect the gravity in vertical infiltration into the matrix and when the infiltration through the walls is taken as horizontal only, we obtain the ponding time in ranges of inequality

$$1/2v_r^2(S^2 + 2v_rA_{co}/A\alpha) < t_p < 1/2v_r^2[S^2(n+1)^2 + 2v_rA_{co}/A\alpha] \tag{6.22}$$

with increase of area of infiltration into the matrix from the walls of the cracks $n = 2l_{co}/A\alpha$, and S is the sorptivity (Philip, 1957) of the matrix. The first term is for instantaneous filling of cracks at the start of the rain, the last term is for the start of the filling of the cracks at the ponding time of the matrix in vertical infiltration through the topographical surface (Kutílek and Novák, 1976). The ponding time is increased by $A_{co}/A\alpha$, practically by 10–100 min.

When the conduction of water in cracks is combined with the absorption, the kinematic wave approximation (Germann and Beven, 1985) can be used. In this theory, the volume flux density q [L T^{-1}] is restricted just to macropores. Further on, the kinematic wave velocity c [L T^{-1}] is dependent upon the depth and the flux in macropores is gradually reduced by the absorption r [T^{-1}]. The macropore content is w. The starting relations between the variables are:

$$\partial q/\partial t + c\ \partial q/\partial z + crw = 0 \qquad (6.23)$$

$$q = bw^{a} \qquad (6.24)$$

where b is the macropore conductance. Since $c = \mathrm{d}q/\mathrm{d}w$, we get from Equation (6.24)

$$c = abw^{(a-1)} \qquad (6.25)$$

Absorption is

$$r = -[1/w(t)]\ \mathrm{d}w/\mathrm{d}t \qquad (6.26)$$

or

$$r = -[1/aq]\ \mathrm{d}q/\mathrm{d}t \qquad (6.27)$$

The coefficient $a = 3$ follows from the assumption of laminar flow. Further on, Equation (6.26) says that the absorption of water from macropore to matrix is analogous to the linear adsorption isotherm. Variable r, Equation (6.23) is dependent upon w, the water content in the macropores and not upon the soil water content (initial) in the matrix as is the usual formulation of the absorption. It is assumed that the macropore volume V_m does not change with depth and the portion of water absorbed from w is independent of the depth. The transformation of a simple square pulse with time and depth is demonstrated in Fig. 6.13: q_s denotes the surface flux (puls) of duration t_s. The curve q_s(B) shows the decreasing flux density at the wetting front. The projection of this curve to the plane t, z, is the wetting front depth. AC is the volume flux density at the draining front and z^* is the maximum depth reached by the propagation of the pulse. It is worth noting the maximum shift of t_s, the final time of the pulse at the depth $z(t_s)$. The procedure looks adequate for approximately the same density and volume of macropores with the depth. Then the method is very applicable, especially for the solution of the by-pass transport of dissolved matter. In cracking Vertisols, the gradual closing of the cracks with wetting can reduce both w and q with time. The speed of closing depends upon the nature of the soil. In some alkali Vertisols the closing is, due to high ESP, fast, from hours till days. In Pakistani black cotton soils, the closing of cracks lasted even more than 1 month (Kamphorst, 1988).

The flow in the system of cracks combined with absorption into the soil matrix is also solved numerically (e.g. Towner, 1987) but without appropriate generalization of the individual solutions.

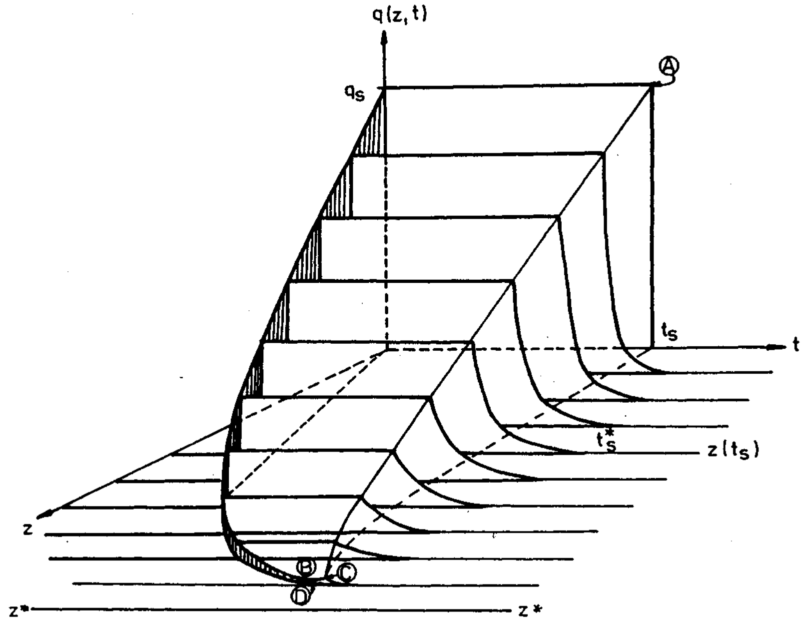


Fig. 6.13. Transformation of the square flux q_s pulse with time t and depth z when the absorption into matrix is considered in kinematic wave approximation (Germann and Beven, 1985).

6.4. CONSEQUENCES FOR WATER MANAGEMENT PRACTICES

6.4.1. Variability of hydraulic functions

Random determination of soil hydraulic functions cannot contribute properly to water management practices without appropriate study of the spatial variability of those functions. When the variability of factors influencing the soil hydraulic functions is compared, we find that ESP and EC of the pore water are highly variable (see, e.g., Folorunso et al., 1988) when compared to the relatively low variability of other factors influencing the hydraulic functions, as, for example, clay content, bulk density, organic matter content. Therefore, the variability of SWRC and of both K_S and $K(w)$ are much higher in Vertisols under the salinity hazard than in Vertisols of the mild zone. When this variability is neglected the comparison of yields with water management practices can bring contradictory results just due to the different size of experimental plots.

If the procedure of pedotransfer functions (Wagenet et al., 1991) is applied in the areas where the density of directly measured soil hydraulic functions is low, the ESP and EC are important parameters for the estimates of hydraulic functions through regression analysis. Principally, the parameters of the pedotransfer functions should be derived for the same region by regression analysis as in the region where those functions are used, especially when we deal with Vertisols under the salinity hazard.

2. *Available water capacity*

Vhen the available water capacity (AWC) was taken as the difference between water contents at 330 hPa (0.33 bar) and 1.5 MPa (15 bar), Kutílek (1973) ıined from this static traditional method:

	Bulk density (g/cm^3)	AWC (g/100 g)
ertisol, unconfined	0.78	29.5
= 27.5), confined	1.17	21.7
confined	1.25	15.7
ertisol, unconfined	1.02	17.0
confined	1.16	20.5

ɔntrary to the statical traditional determination of AWC, the practitioners ı very well that an increase of ESP is accompanied by a decrease of AWC. statical concept of AWC should be critically reexamined.

ıe field capacity represents the soil water content when the rate of tribution after the initial full saturation is so slow that practically the soil water nt does not change in time. In the experiment, the soil surface should be cted against evaporation. For routine methods, the "1/3 bar soil water nt" is regularly used. When very low values of K or, eventually of D are lered, the infiltration in soil matrix is very slow and sometimes even not able in reasonable time. The redistribution process is much slower than the ation and the change of water content is not measurable even in 5-day als. If we reach the saturation of the top of the soil profile with water then :duction of water content is not detectable. This is why Farbrother (1987) ses to consider the saturated water content as the top boundary of AWC tisols of the semiarid zone. However, the question arises immediately, which ted water content, if there is a broad range of water content at saturation. classical method of Kopecky (1914) is used, his absolute capacity is more ı related to the field capacity and to the field conditions and the problem ı to "1/3 bar soil water content" will be avoided.

: lower boundary of AWC is formed by the wilting point. Let us discuss this l soil water content when very low values of water diffusivity are considered. ve already shown that in alkali Vertisols D is approximately constant over d range of water content and when ESP is very high, D can even decrease the soil water content increases. When we take $D_w = 2 \times 10^{-3}\,cm^2\,min^{-1}$ ritical value when the permanent wilting of plants occur under current ons, Kutílek (1973, 1983) shows that the soil water content related to D_w h higher than the 1.5 MPa (15 bar) water content in Vertisols with 10 t < ESP < 20 percent. For Vertisols with ESP > 20 percent when confined, ıe whole range of water content was lower than D_w and we can expect that ter flux into the roots is so insufficient that the plants either cannot develop

as they don't get enough water to cover the minimum evapotranspiration, or the flux into the roots is very low so that stomatal openings are so closed that the photosynthesis is at a minimum. Generally, the alkali processes induce a drastic increase of the lower boundary of the available water. The concept of 1.5 MPa wilting point should be then abandoned in Vertisols under the hazard of alkali processes. Instead of this method, the direct experiment on plant wilting should be employed. In deep horizons with ESP > 20 percent we can expect that those horizons form a barrier against the development of the root system to the depth. This barrier effect is due to the extremely low influx of water in the roots.

We arrive therefore at the conclusion that available water cannot be considered for Vertisols by procedures empirically derived in the mild zone.

6.4.3. *Evaporation from Vertisols*

It is generally assumed that the soil–atmospheric air interface increases due to cracks up to several times and evaporation is supposed to increase substantially as well. Ritchie and Adams (1974) have experimentally found that cracks are the main path of evaporation from the bare soil. When plants are present, we shall differentiate between the evaporation from the topographic soil surface, E_s, evaporation from the cracks, E_c and evapotranspiration, E_T. The ratio of evaporation and evapotranspiration $(E_s + E_c)/E_T$ decreases with increase of the leaf area index LAI and starting from approx. $LAI = 0.2$ to 0.5, the ratio is < 1 when the data of Novák (1981) are interpolated. In his report on experimentation the cracking system was not mentioned.

When we analyze the process in detail, we have to consider the principal difference between the air in cracks and the air above the topographical surface. Air in cracks has different regime from the atmosphere. Even for bare soil, the simple assumption that the area of walls of cracks increases the total evaporation area algebraically is false. Hatano et al. (1988) have derived theoretically and proved experimentally that in the cracks a zero vapor flux depth (ZVFD) exists well above the bottom of the cracks. Above ZVFD, the vapor diffuses upward into the free atmosphere. Below ZVFD the vapor flux is downward. The thermal effects play an important role in the cracks. The thermal conductivity of cracks filled with air is substantially lower than in soil which is a mixture of solid particles, water and air. Consequently, there was a rapid rise of temperature in the upper portion of cracks which caused a considerable increase of water vapor density and the gradient of the water vapor density was divided into upward and downward directions at ZVFD. In the wheat field with $LAI = 1.2$ Hatano et al., found that $(E_s + E_c)/E_T$ was 0.02–0.12 and the cracks contributed in ranges from $E_c = 0.1E_s$ to $E_c = E_s$. The whole area of crack walls was about $14\,m^2$ per $1\,m^2$ of topographical surface. The variation of the components of the evapotranspiration was due to the variation of meteorological conditions in the period June–July. The position of ZVFD was about 5 cm. The observation of Hatano et al., is valid for the mild zone, in the semiarid zone the position of ZVFD can be expected to be deeper and the contribution of cracks to evaporation will be slightly higher.

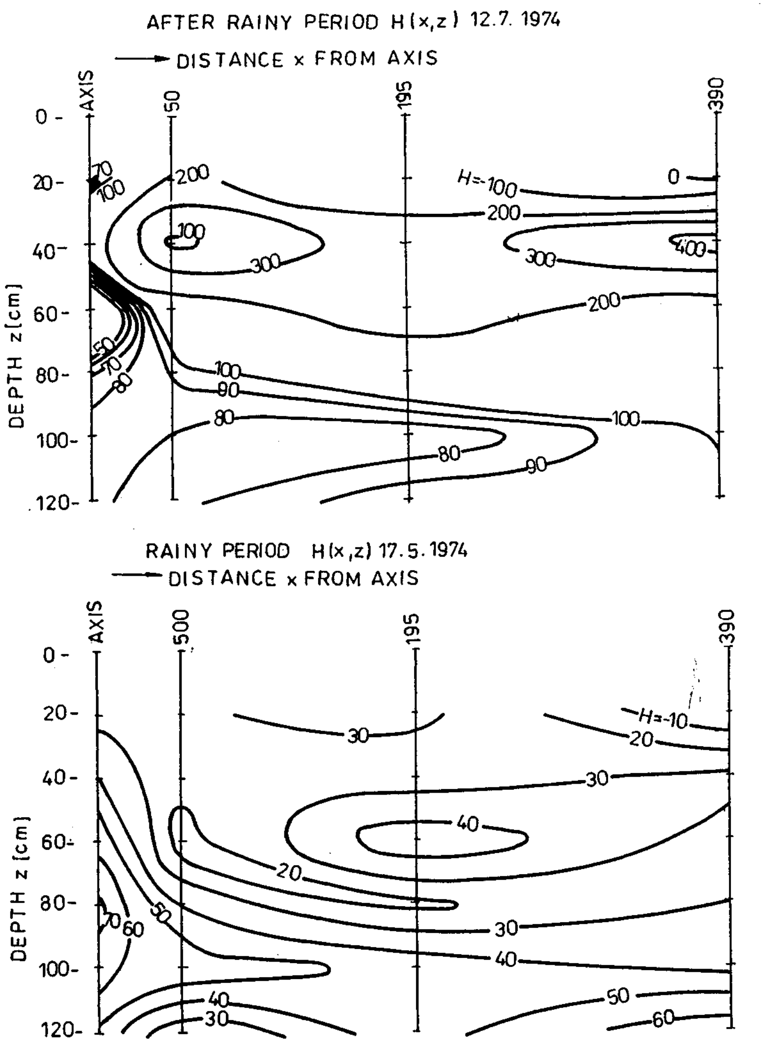


Fig. 6.14. Potential fields in the drained Vertisol of the mild zone. Tile drain is on axis position at the depth 80 cm, spacing is 7.8 m (Kutílek et al. 1976).

Generally, there is no simple addition of the crack wall area to the evaporation surface and we have to avoid the up to now overestimation of the role of cracks in the evaporation process. The assumption of substantial reduction of evaporation from the soil if the crack evaporation is prevented does not work.

6.4.4. Drainage of Vertisols

Theories on tile drainage in Vertisols can be based on more or less controversial approaches. Generally we can classify three groups:

(1) The classical potential field theory where the flow is analogous to the flow observed and theoretically treated in sandy and light loamy soils. The draining function is assumed to be the same all over the cross-sectional plain between the tiles and only the low values of hydraulic conductivity distinguish clays from silty and sandy soils. This model has been abandoned as non-realistic, the real discharge in the tiles is by orders higher than the flux produced under the field gradients of the potential and with the extremely low values of the conductivity.

(2) In drained Vertisols two distinct flow fields exist: The backfill of the trench and the undisturbed soil between the trenches and tiles, the interfield. While the soil between the trenches has no draining function, the backfill is drained. Its hydraulic conductivity is by orders higher and the preferential ways of macropores in backfill result in fast reaction of the drainage discharge to the drains. The earlier assumption that the ploughed subsoil conducts water to the backfill and contributes to the discharge in tiles has not been confirmed by experiments. In Fig. 6.14 the potential fields on the plane perpendicular to the tiles illustrate well how the two flow fields are separated in the wet period when the discharge in the tiles was measured. In the long term experiments, it has been found that the groundwater level during heavy rains and in 50 percent of the rainless periods was higher in the backfill than in the interfield. The drainage of the interfield by backfill and by the tiles can, in these periods, be excluded. It was found that probably the backfill collects a portion of rainwater from the interfield by surface runoff (Kutílek et al., 1976). This model looks appropriate when the macropores in the interfield are rare, the cracks are due to the higher water content nonexistent in the interfield.

(3) The preferential ways exist in both the backfill as well in the interfield. Their density in the backfill and the interfield may be different. The main flux between the water on the soil surface and the drain occurs through the interconnected net of macropores. The matrix of peds is wetted or drained by the macropores with a substantial delay. The water level in macropores reacts fast to rains and when regulated drainage was installed on this type of Vertisol, the water level in the macropores reacted in tens of minutes upon the regulation mechanismus on the tile drainage while the water content in the matrix of peds was intact (Sutor, 1976, personal communication). A detailed review and the two-domain model is presented by Jarvis and Leeds-Harrison (1987).

REFERENCES

Abdine, A.Z., 1971. Factors affecting hydraulic conductivity in the soil profile of the Gezira. Appendix III. Technical Report attached to Terminal Statement: FAO Project SF/SUD/15.

Bear, J., 1972. Dynamics of Fluids in Porous Media. Elsevier, Amsterdam.

Beven, K. and Germann. P.F., 1981. Water flow in soil macropores: 2. A combined flow model. J. Soil Sci. 32: 15–29.

Bouma, J. and Wösten, J.H.M., 1979. Flow patterns during extended saturated flow in two undistributed swelling clay soils with different macrostructures. Soil. Sci. Soc. Am. J., 43: 16–22.

Bouma, J.A., Jongerius, A., Boersma O. and Jagger, A., 1977. The function of different types of macropores during saturated flow through four swelling soil horizons. Soil Sci. Soc. Amer. J., 41: 945–950.
Bronswijk, J.J.B., 1991. Relation between vertical soil movements and water content changes in cracking clays. Soil Sci. Soc. Amer. J. 55: 1220–1226.
Bui, E.N. and Mermut, A.R., 1988. Orientation of planar voids in Vertisols and soils with vertic properties. Soil Sci. Soc. Amer. J. 52171–178.
Corey, A.T., 1977. Mechanics of Heterogeneous Fluids in Porous Media. Water Res. Publ., Fort Collins, Co.
Durner, W., 1991. Vorhersage der hydraulischen Leitfähigkeit strukturierter Böden. Dissertation, Universit. Bayreuth.
Farbrother, H.G., 1987. Supplementary irrigation. In: IBSRAM: Management of Vertisols under Semi-Arid Conditions, Proc. No. 6, 1st Reg. Seminar, Nairobi, pp. 267–282.
Folorunso, O.F., Ohu, J.O. and Adeniji, F.A., 1988. The role of soil spatial variability investigation in the management of the Chad Basin Vertisols of N.E. Nigeria. Soil Technol. 1: 149–156.
Gairon, S. and Swartzendruber, D., 1973. Streaming potential effects in saturated water flow through a sand-kaolinite mixture. In: A. Hadas, D. Swartzendruber, P.E. Rijtema, M. Fuchs and B. Yaron (Editors), Physical Aspects of Soil Water and Salts in Ecosystems. Ecol. Studies 4, Springer, Berlin, pp. 141–151.
Gerber, T.D., Wilding, L.P. and Franklin, R.E., 1974. Ion diffusion across cutans: A methodology study. In: G.K. Rutherford (Editor), Soil Microscopy, 4th Int. Working Meeting on Soil Micromorphology. The Limestone Press, Kingston, ON, pp. 730–746.
Germann, P. and Beven, K., 1981. Water flow in soil macropores. I. An experimental approach. J. Soil Sci., 32: 1–13.
Germann, P. and Beven, K., 1985. Kinematic wave approximation to infiltration into soils with sorbing macropores. Water Res., 21: 990–996.
Gersevanov, N.M., 1937. The Foundation of Dynamics of Soils, 3rd ed. Strojizdat, Moscow-Leningrad.
Giraldez, J.V., Sposito, G. and Delgado, C., 1983. A general soil volume change equation: I. The two parameter model. Soil Sci. Soc. Amer. J., 47: 419–422.
Groenevelt, P.H. and Bolt, G.H., 1972. Water retention in soil. Soil Sci., 113: 238–245.
Haines, W.B., 1923. The volume changes associated with variations of water content in soil. J. Agric. Sci., 13: 293–310.
Hallaire, V., 1984. Evolution of crack networks during shrinkage of a clay soil under grass and winter wheat crops. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement in Heavy Clay Soils. ILRI Publ. 37, Wageningen, pp. 49–53.
Hatano, R., Nakamoto, H., Sakuma, T. and Okajima, H., 1988. Evapotranspiration in cracked clay field soil. Soil Sci. Plant Nutr., 34: 547–555.
Iwata, S., Tabuchi, T. and Warkentin, B.P., 1988. Soil Water Interactions. Mechanisms and Applications. Marcel Dekker, New York, Basel.
Jarvis, N.J. and Leeds-Harrison, P.B., 1987. Modelling water movement in drained clay soil. I. Description of the model sample output and sensibility analysis. II. Application of the model in Evesham series clay soil. J. Soil Sci., 38: 487–498, 499–509.
Kamphorst, A., 1988. Water and salt transport in the irrigated cracking clay soils of the Kachci Plains, Pakistan. I. Vertical transport. Soil Technol., 1: 271–281.
Kopecky, J., 1914. Die physikalische Eigenschaften des Bodens. Berlin.

Kutílek, M., 1964a. The influence of soil colloids upon the values of some hydrolimits. Rostlinná vyroba, 10: 609–623.
Kutílek, M., 1964b. The filtration of water in soils in the region of the laminar flow. Trans. VIII Congr. ISSS, Vol. II, Bucarest, pp. 45–51.
Kutílek, M., 1972. Non-Darcian flow of water in soils—laminar region. In: Fundamentals of Transport Phenomena in Porous Media. Elsevier, Amsterdam, pp. 327–340.
Kutílek, M., 1973. The influence of clay minerals and exchangeable cations on soil moisture potential. In: A. Hadas, D. Swartzendruber, P.E. Rijtema, M. Fuchs and B. Yaron (Editors), Physical Aspects of Soil Water and Salts in Ecosystems. Ecol. Studies 4, Springer, Berlin, pp. 153–160.
Kutílek, M., 1983. Soil physical properties of saline and alkali Vertisols. In: Isotope and Radiation Techniques in Soil Physics and Irrigation Studies, IAEA, Vienna, pp. 179–190.
Kutílek, M., 1984. Some theoretical and practical aspects of infiltration in clays with D = constant. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement in Heavy Clay Soils. ILRI Publ. 37, Wageningen, pp. 114–128.
Kutílek, M. and Novák, V., 1976. The influence of soil cracks upon infiltration and ponding time. In: M. Kutílek and J. Šútor (Editors), Water in Heavy Soils. Proc. I. ICID and ISSS, Bratislava, pp. 126–134.
Kutílek, M. and Semotán, J., 1975. Soil water properties of Gezira soils. In: New Development in the Field of Salt Affected Soils, Proc. Int. Symp. Cairo, pp. 299–308.
Kutílek, M., Svobodová, V., Doležal, F., Kuráž, V., Pařízková, M., and Semotán, J., 1976. Potential flow in tile drained clay soil—field experiment. In: M. Kutílek and J. Šútor (Editors), Water in Heavy Soils, Proc. I, ICID and ISSS, Bratislava, pp. 215–218.
McGarry, D. and Malafant, K.W., 1987. The analysis of volume change in unconfined units of soil. Soil Sci. Soc. Amer. J., 51: 290–297.
McIntyre, D.S., Watson, C.L. and Loveday, J., 1982. Swelling of a clay soil profile under ponding. Aust. J. Soil Res., 20: 71–79.
Novák, V., 1976. Cracks in swelling soil and the calculation of their characteristics. In: M. Kutílek and J. Šútor (Editors), Water in Heavy Soils, Proc. II, ICID and ISSS, Bratislava, pp. 21–41.
Novák, V., 1981. The structure of evapotranspiration (in Slovak) I and II. Vodohosp. As, 29: 476–492, 581–592.
Othmer, H., Diekkrüger, B. and Kutílek, M., 1991. Bimodal porosity and unsaturated hydraulic conductivity. Soil Sci., 152: 139–150.
Philip, J.R., 1957. The theory of infiltration: 4. Sorptivity and algebraic infiltration equations. Soil Sci., 84: 257–264.
Philip, J.R., 1969. Hydrostatics and hydrodynamics in swelling soils. Water Res., 5: 1070–1077.
Raats, P.A.C., 1984. Mechanics of cracking soils. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement in Heavy Clay Soils. ILRI Publ. 37, Wageningen, pp. 350–354.
Ringrose-Voase, A. and Bullock, P., 1984. The measurement of soil structural parameters by image analysis. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement in Heavy Clay Soils. ILRI Publ. 37, Wageningen, pp. 350–354.
Ritchie, J.T. and Adams, J.E., 1974. Field measurement of evaporation from soil shrinking

cracks. Soil Sci. Soc. Amer. Proc., 38: 131–134.
Schweikle, V., 1982. Gefügeeigenschaften von Tonböden. Verlag Eugen Ulmer, Stuttgart.
Shainberg, I., Rhoades, J.D. and Prather, J.R., 1981. Effect of low concentration on clay dispersion and hydraulic conductivity of a sodic soil. Soil Sci. Soc. Amer. J. 45: 273–277.
Smiles, D.E., 1976. Theory of liquid flow in saturated swelling materials: Some problem areas. In: M. Kutílek and J. Šútor (Editors), Water in Heavy Soils, Proc. I, ICID and ISSS, Bratislava, pp. 32–41.
Smiles, D.E. and Rosenthal, J., 1968. The movement of water in swelling materials. Austr. J. Soil Res., 6: 237–248.
Sposito, G. and Giraldez, J.V., 1976. On the theory of infiltration in swelling soils. In: M. Kutílek and J. Šútor (Editors), Water in Heavy Soils, Proc. I, ICID and ISSS, Bratislava, pp. 107–118.
Stroosnijder, L. and Bolt, G.H., 1984. The moisture characteristic of heavy clay soils. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement in Heavy Clay Soils. ILRI Publ. 37, Wageningen, pp. 324–329.
Swartzendruber, D., 1962. Modification of Darcy's law for the flow of water in soil. Soil Sci., 93: 22–29.
Swartzendruber, D., 1963. Non-Darcy behavior and the flow of water in unsaturated soils. Soil Sci. Soc. Amer. Proc., 27: 491–495.
Tariq, A. -ur-R. and Durnford, D.S., 1993. Analytical volume change model for swelling clay soils. Soil Sci. Soc. Amer. J., 57: 1183–1187.
Towner, G.D., 1987. Formulae for calculating water flow in macropores in soil. Int. Agrophysics (Akadémiai Kiadó/M. Nijhoff/Dr. W. Junk Publ.), 3: 5–15.
van Genuchten, M.T., 1980. A closed form equation for predicting the hydraulic conductivity of unsaturated soils. Soil Sci. Soc. Amer. J., 44: 892–898.
van Olphen, H., 1977. An Introduction to Clay Colloid Chemistry. 2nd ed. Interscience, New York.
Virgo, K.J., 1981. Observation of cracking in Somali Vertisols. Soil Sci., 131: 60–61.
Wagenet, R.J., Bouma, J. and Grossman, R.B., 1991. Minimum data sets for use of soil survey information in soil interpretive models. In: Spatial Variabilities of Soils and Landforms. SSSA Spec. Publ. No. 28, SSSA, Madison, WI, pp. 161–182.
White, R.E., 1985. The influence of macropores on the transport of dissolved and suspended matter through soil. In: B.A. Stewart (Editor), Advances in Soil Sci. 3, Springer Verlag, New York, pp. 95–120.
Wilding, L.P. and Hallmark, C.T., 1984. Development of structural and microfabric properties in shrinking and swelling clays. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement in Heavy Clay Soils. ILRI Publ. 37, Wageningen, pp. 1–18.
Yalon, D.H. and Kalmar, D., 1984. Extent and dynamics of cracking in heavy clay soil with xeric moisture regime. In: J. Bouma and P.A.C. Raats (Editors), Proc. of the ISSS Symp. on Water and Solute Movement. ILRI Publ. 37, Wageningen, pp. 45–48.
Yule, D.F. and Ritchie, J.T., 1980. Soil shrinkage relationships of Texas vertisols: I. Small cores. Soil Sci. Soc. Amer. J., 44: 1285–1291.
Zrubec, F., 1976. Processes of shrinking and swelling in heavy soils of Slovakia in relationship to some physical properties. In: M. Kutílek and J. Šútor (Editors), Water in Heavy Soils. Proc. III. ICID and ISSS, Bratislava, pp. 26–33.

Chapter 7

THE STRUCTURE AND GRAIN SIZE DISTRIBUTION OF VERTISOLS

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7.1. INTRODUCTION

Structure is the most striking, visual aspect of Vertisol morphology (Blokhuis, 1982). The type and degree of structure development in a Vertisol provides a digest of its genesis, constituent properties and agriculture management potential. The present challenge with the study of Vertisol structure is not to discuss the range of soil structure nor associated porosity or fabric, as these have been covered often. Rather it is to investigate the interaction of structure with constituent properties, external influences and processes in order to explain the links between "good structure", desirable physical properties, inherent agricultural potential and the maintenance of good structure (Fig. 7.1). Commercial practices may then evolve to activate or supplement constituent properties towards improving the inherently poor structure of some Vertisols, analogous to fertiliser application on low fertility soils. Human-induced structure degradation can also be targeted and repaired by activating or supplementing the in-built resiliency of Vertisol structure.

Grain size distribution varies greatly among Vertisols, and together with variations in clay mineralogy is a major source of heterogeneity in this soil order. Clay content can range from 30-80 percent and dominant clay types can be montmorillonite or kaolin with varying amounts of illite (Probert et al., 1987). The link between these variations and structure development will be considered.

This chapter will review the literature on the structure and grain size distribution of Vertisols. The principal aim is to assess the relative roles of constituent soil properties and external influences on both the degree of structure development and the type of structure developed. A conceptual model will be constructed, as a framework towards the sorting of Vertisols in terms of the influences that control structure development. This is a first step in deriving a numerical model of structure development in Vertisols towards finding commonality and interrelations in the soils' properties. As such the conceptual model will be a broad guide to further studies aimed at filling gaps in knowledge as well as an aid to sorting the highly varied information on what controls structure development in Vertisols. Major sections of the chapter are: previous reviews, morphological description, grain size distribution and clay mineralogy, organic matter, exchange capacity and exchangeable cations, climate, wetting and drying rates and soil colour. Outside