HILGARDIA

JOURNAL OF AGRICULTURAL SCIENCE PUBLISHED BY
HE CALIFORNIA AGRICULTURAL EXPERIMENT STATION

Volume 52 Number 2 • January 1984

Response of Soils to Sodic and Saline Conditions

I. Shainberg and J. Letey

NIVERSITY OF CALIFORNIA DIVISION OF AGRICULTURE AND NATURAL RESOURCES

The interaction between soil particles and the soil solution depends on the types and amounts of soil clays. Montmorillonite is the most reactive clay. Ca-clay forms tactoids (quasi crystals) and has limited swelling and dispersion. Na-clay forms single platelets and disperses freely. In a mixed Na/Ca system, "demixing" of the cations occurs as the Na ions concentrate on the external and the Ca ions on the internal surfaces of the tactoids. The demixing explains why a small percentage of exchangeable Na sharply increases the zeta potential and dispersion of montmorillonite clay.

The hydraulic conductivity of a soil depends on both Na and the total salt concentration of the percolating solution. High hydraulic conductivity may be maintained, even at high exchangeable sodium percentage (ESP) values, if the solution concentration is above a critical (threshold) level. When waters of very low salt content are used, decreases in hydraulic conductivity and clay dispersion occur even in soils with low ESP values (<10).

The main mechanism for hydraulic conductivity reduction in waters of medium to high salinity (>0.5 dS/m) is clay swelling. High content of expansible clays increases the susceptibility of soils to intermediate ($10<ESP<20$) sodicity. The effect of low sodicity ($ESP < 10$) on clay swelling is minimal and the hydraulic conductivity of these soils is maintained at high values.

When waters having low salt concentrations (salt concentration below the flocculation value of the clay) are used, clay dispersion, movement, and lodgement in the conducting pores comprise the main mechanism responsible for hydraulic conductivity reduction. These results occur even at low ESP. The most important soil property which determines its susceptibility to sodic conditions, when the soil is leached with rain or snow water, is the potential of the soil to release electrolytes from primary minerals and/or $CaCO_3$ dissolution. Soils which release sufficient electrolytes to maintain the concentration of the soil solution above the flocculation value of the clay will not disperse, and thus will not be sensitive to low sodicity.

THE AUTHORS:

I. Shainberg is the Head of the Department of Soil Physical Chemistry, Institute of Soils and Water, Volcani Center, ARO, Bet-Dagan, Israel and Visiting Professor, University of California, Riverside. J. Letey is Professor of Soil Physics, Department of Soil and Environmental Sciences, University of California, Riverside.

I. Shainberg and J. Letey

Response of Soils to Sodic and Saline Conditions[1]

INTRODUCTION

SALINE AND SODIC SOIL CONDITIONS reduce the value and productivity of soils in the arid and semiarid regions of the world. Accumulation of soluble salt in the soil solution imposes stress on growing crops that can lead to decreased yields and, in severe cases, complete crop failure. Accumulation of dispersive cations such as sodium and, possibly, potassium in the soil solution and the exchange phase affects soil physical properties such as structural stability, hydraulic conductivity, and infiltration rate, which consequently affect crop production.

Management of saline and sodic soils for crop production involves understanding three interrelated general areas: crop response to saline conditions, simultaneous transport of salt and water in soil, and response of soils to sodic and saline conditions.

Although the first area has been studied mainly by plant scientists and agronomists, the second and third areas have been studied mainly by soil scientists. Considerable information is available in the technical literature on all three areas, but knowledge in the third area is more limited and empirical than in the first two. Rhoades (1972) and Frenkel et al. (1978), evaluating water quality for irrigation, concluded, "Our greatest limitation in assessing the sodicity hazard [of irrigation water] is our inability to predict how the water will affect soil structure and permeability."

An indication of the superficiality of our understanding of the response of soils to sodic and saline conditions is the confusion we face in trying to answer the following five questions.

1) What is the critical concentration of exchangeable sodium which separates normal from sodic soils? The US Salinity Laboratory Staff (1954), after studying many soil samples from the western USA, suggested an exchangeable sodium percentage (ESP) of 15 as the critical level above which soil structure could be deleteriously affected. This value was considered to be far too high for Australian soils, and McIntyre (1979), studying the hydraulic conductivity of 71 soil samples from Australia, proposed an ESP of 5 more appropriate for separating normal from sodic soils. This difference in the threshold sodium percentage is very important in the applied world. Whereas ESP values above 15 are quite rare in soils of arid and semiarid regions, soils with ESP values around 5 are wide spread. If we adopt the Australian definition, most soils in semiarid and arid regions should be classified as sodic soils and care should be undertaken in their management.

---

[1] Accepted for publication April 13, 1983. Contribution from the Institute of Soils and Water, The Volcani Center, ARO, Bet-Dagan, Israel and the Department of Soil and Environmental Sciences, University of California, Riverside 92521. This review was supported by the University of California Kearney Foundation of Soil Science and a grant from the United States-Israel (Binational) Agricultural Research and Development Fund (BARD).

2) What is the effect of clay content and clay mineralogy on the response of soils to sodic conditions? McIntyre (1979) found that ESP influence on the hydraulic conductivity of the 71 soil samples was independent of their clay content and mineralogy. Conversely, McNeal and his co-workers (1966, 1968) found a good correlation between expansible clays (type and content) and hydraulic conductivity losses due to exchangeable sodium.

3) What determines the response of soils having similar clay content and clay mineralogy to exchangeable sodium? The effect of exchangeable sodium on the physical properties of soils is gradual, but some soils are sensitive to low percentages of exchangeable sodium when leached with distilled water (e.g., Fallbrook soil in California and Netanya soil in Israel), and others having similar texture and clays are not as sensitive (e.g., Gila soil in California and loess soil in Israel) (Frenkel et al., 1978; Shainberg et al., 1980; Alperovitch et al., 1981).

4) Which of the processes and properties in a given soil are more sensitive to exchangeable sodium, and why? Many of the soil properties are affected differently by exchangeable sodium. For example, clay swelling and clay dispersion are affected by exchangeable sodium to different degrees. Clay disperses at low ESP levels (Oster et al., 1980), whereas clay swelling is affected only at higher ESP values (McNeal et al., 1968 ; Shainberg et al., 1971). Similarly, hydraulic properties of the soil surface, which are important during the infiltration process, are much more sensitive than soil profile hydraulic conductivity to low ESP in the same soil (Oster and Schroer, 1979; Kazman et al., 1983).

5) What is the effect of exchangeable magnesium (Mg) on the physical properties of soils? The US Salinity Laboratory Staff (1954) grouped calcium (Ca) and Mg together as similar ions beneficial in developing and maintaining soil structure. However, laboratory studies have shown that Mg has deleterious effects in some circumstances (Abder-Rahman and Rowell, 1979; Alperovitch et al., 1981; Chi et al., 1977; Emerson and Chi, 1977; McNeal et al., 1968). Similarly, what is the effect of exchangeable potassium on the physical properties of soils? Is potassium a beneficial or dispersive cation?

The objective of this report is to synthesize information on the response of soils to sodic and saline condition by (1) describing the mechanisms by which exchangeable sodium affects soil properties; (2) developing an understanding of the soil processes which affect the response of soils to sodic conditions; and (3) defining the areas of research where knowledge is lacking and more research is to be done. The goal of this review is to provide answers to the questions listed above and to develop the ability to predict the effect of irrigation with waters of various qualities on soil's physical properties.

THE EFFECT OF SODICITY AND SALINITY ON THE COLLOIDAL PROPERTIES OF SOIL CLAYS

Mineralogical structure of clays and the diffuse double layer

The amount of interaction between soil solution and the soil particles depends on the types and amounts of the soil clays. The dominant clay mineral in the semiarid and arid regions is montmorillonite (a member of the smectite group, which also includes beidellite, nontronite, etc.), but illite and kaolinite are also common. The clay minerals' colloidal properties depend on their specific surface areas (surface area per gram of clay),

which depend on the size and shape of the clays. The specific surface areas of montmorillonite, illite, and kaolinite are 750, 150, and 30 square meters per gram (m^2/g), respectively. This explains why montmorillonite is the most reactive constituent among the soil mineral colloids. However, clay reactivity also depends on the cationic composition and the free electrolyte concentration in the equilibrium solution. This point will be discussed later.

Two structural units are involved in the atomic lattices of most clay minerals. One unit consists of two-dimensional arrays of silicon-oxygen tetrahedra and the other unit consists of sheets of aluminum (or iron or magnesium)-oxygen-hydroxyl octahedra. The clay minerals in the soil are classified according to the way in which these two sheets are stacked to form a unit layer. Grim (1968) classified clay minerals as follows. In *two-layer types,* the unit layer is composed of one sheet of silica tetrahedra and one sheet of aluminum octahedra. The two sheets are held together by shared oxygen atoms. The layers show little tendency to separate because of the hydrogen bonds between the layers. Thus, kaolinite clays, which belong to this group, have a relatively low specific surface area and do not swell. In *three-layer types*, the unit layers are composed of two sheets of silica tetrahedra with an octahedral sheet between. The three sheets are held together by shared oxygen atoms. Individual layers are stacked to form crystalline particles.

Two recognized groups of three-layer clays are an expanding-lattice or smectite group, and a non-expanding-lattice or illite group, distinguished from the smectites mainly by the absence of an expanding interlayer. When montmorillonite clay is wetted, water molecules penetrate between the unit layers, causing their separation, and the lattice expands. The extent of this expansion depends on the exchangeable cation and the electrolyte concentration in the solution.

Substitution of trivalent aluminum for quadrivalent silicon in the tetrahedral sheet, and of magnesium or ferrous ions for aluminum in the octahedral sheet (isomorphous substitution) results in a negative charge in the lattice. This charge is uniformly distributed within the lamellate clay particles. Charges also result from broken bonds at the edges of the tetrahedral and octahedral sheets. These surfaces are either positively or negatively charged, depending on the pH of the solution. At neutral pH, the charge at the edges is positive. The particle charge is compensated by the counter ions in the liquid immediately adjacent to the clay surface.

The diffuse double layer consists of the lattice charge and the compensating counter ions. The counter ions are subject to two opposing tendencies: the cations are attracted electrostatically to the negatively charged clay surface; and the cations tend to diffuse from the surface of the particle, where their concentration is high, into the bulk of the solution where their concentration is lower. The two opposing tendencies result in decreasing counter ion concentration from the clay surface to the bulk solution. Divalent ions are attracted to the surface with a force twice as large as that of monovalent ions. Thus, the diffuse double layer in the divalent ion system is more compressed toward the surface. With an increase in the electrolyte concentration in the bulk of the solution, the tendency of the counter ions to diffuse away from the surface is diminished and the diffuse double layer compresses toward the surface. Complete descriptions of the diffuse-double-layer theory are presented in textbooks (Bolt, 1979; Bresler et al., 1982; van Olphen, 1977).

Swelling of Na/Ca clays

Homoionic Na and Ca systems. When two clay platelets approach each other, their diffuse counter ion atmospheres overlap as shown in Figure 1. Work must be performed to overcome the electrical repulsion forces between the two positively charged ionic atmospheres. The electric double-layer repulsion force, also called "swelling pressure," can be calculated by means of the diffuse-double-layer theory (Bolt, 1979; Bresler et al., 1982; Shainberg et al., 1971; Warkentin et al., 1957). The greater the compression of the ionic atmosphere toward the clay surface is, the smaller is the overlap of atmospheres for a given distance between particles. Consequently, the repulsion forces between the particles decrease with increases in the salt concentration and in the valence of the adsorbed ions. Because adsorbed sodium (Na) ions form a diffuse layer, high swelling pressures develop between Na-montmorillonite platelets, and single platelets tend to persist in dilute solution. This is also why studies on Na-montmorillonite show that the swelling curve can be predicted from the diffuse-double-layer theory (Fig. 2)(Warkentin et al., 1957; Shainberg et al., 1971). Conversely, the low swelling pressure between calcium (Ca) clay platelets and the electrical attraction forces between the exchangeable calcium ions and the negative clay surface (forces which are not considered in the diffuse-double-layer theory) prevent the indefinite swelling of Ca-montmorillonite, even in distilled water, and the Ca-platelets aggregate into tactoids (Blackmore and Miller, 1961) or quasi crystals (Aylmore and Quirk, 1959; Quirk and Aylmore, 1971).

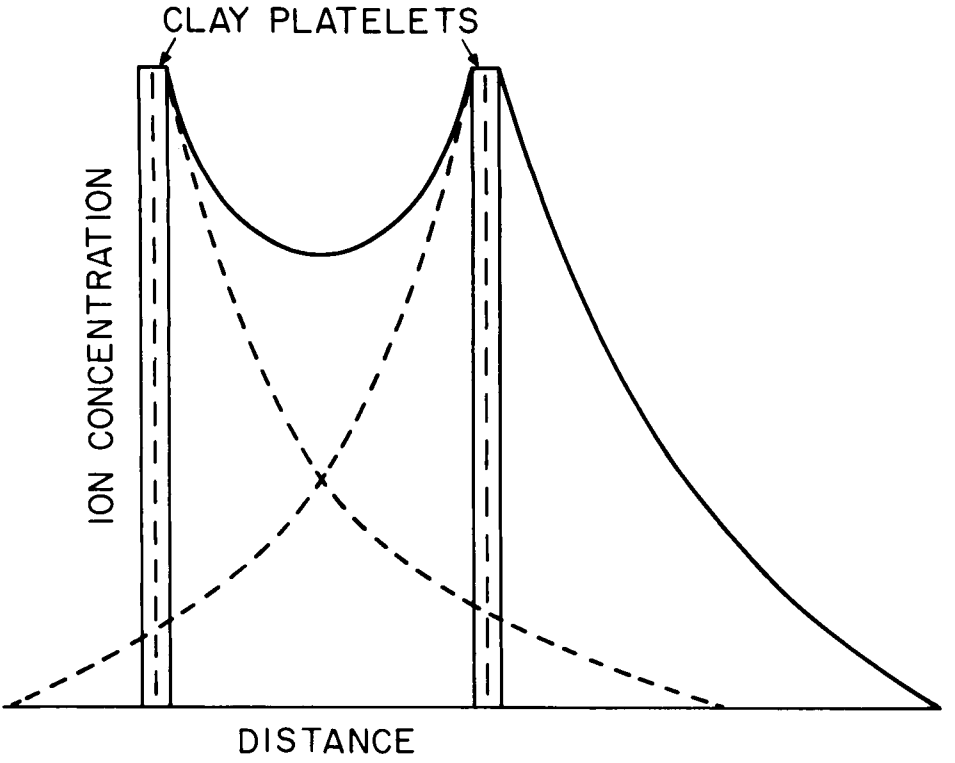

Fig. 1. Overlapping of two diffuse double layers (schematic). Broken line: ionic concentration at a single platelet without overlapping; solid line: ionic concentration for overlapping platelets (Shainberg and Oster, 1978).

Each tactoid consists of several (four to nine) clay platelets in parallel array, with interplatelet distances of 0.9 nm. The exchangeable Ca ions adsorbed on the internal surfaces of the tactoids do not form a diffuse double layer. If a diffuse ion layer is present, it acts only on the outside of these tactoids (Blackmore and Miller, 1961), and the clay behaves like a system having a much smaller surface area. The diffuse-double-layer theory has been used to describe the swelling pressure of Ca-montmorillonite (Shainberg et al., 1971; Russo and Bresler, 1977*a*,*b*). However, a reasonable agreement with experimental data was obtained only by adjusting the number of platelets in a tactoid (quasi crystal) (Fig. 2). The number of Ca-clay platelets per tactoid depends upon the salt concentration in the bulk solution, the applied consolidation pressure, the methods of preparing the Ca-montmorillonite sample, and the exchangeable cation composition. No independent method is available to predict this number.

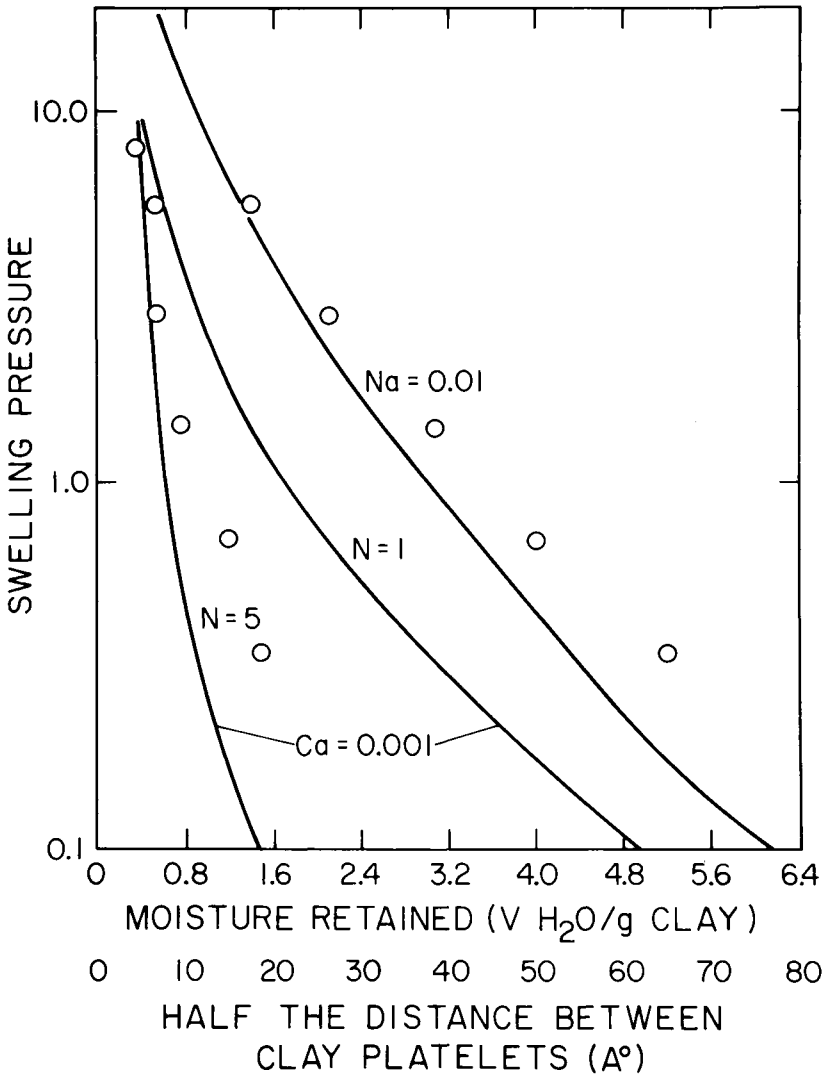

Fig. 2. Swelling pressure of Na- and Ca-montmorillonite as a function of moisture retained (theoretical curves and experimental points). N is the number of platelets in a tactoid (Shainberg et., 1971).

Mixed Na/Ca systems. The soil solution in a saline soil usually contains a mixture of Na and Ca ions. Consideration of a mixed Na/Ca system is therefore important. As indicated in the previous section, the platelets in sodium clays are separated when in equilibrium with dilute salt solution, whereas calcium-saturated clays exist in tactoids with only a fraction of the particles' surfaces active in the swelling-shrinking cycle. Thus, it is important to note how introducing exchangeable sodium to Ca-tactoids will affect both the swelling pressure and the size of the tactoids.

The volume of water retained by montmorillonite clay as a fuction of the exchangeable sodium percentage (ESP) for various consolidation pressures is presented in Figure 3A. Figure 3B shows diffuse-double-layer predictions (based on the Bresler method, 1970) of the moisture retained as a function of the ESP for the same pressure. In the computation, complete mixing of the adsorbed ions was assumed (following the classical diffuse-double-layer theory) and the existence of tactoids was not considered. It is evident that the agreement between the predicted and experimental results is not good. Experimentally, a slight addition of exchangeable sodium to mainly Ca-saturated clay had little effect on the amount of solution retained by the clay, but the classic diffuse-double-layer theory predicts an increase in the amount of moisture retained by the clay. Also, the predicted amount of moisture retained by Ca-clay was well above the experimental values. Russo and Bresler (1977*a,b*) improved the diffuse-double-layer prediction by considering the formation of tactoids; however, the number of platelets in the tactoids was used as an adjustment parameter to get the good agreement.

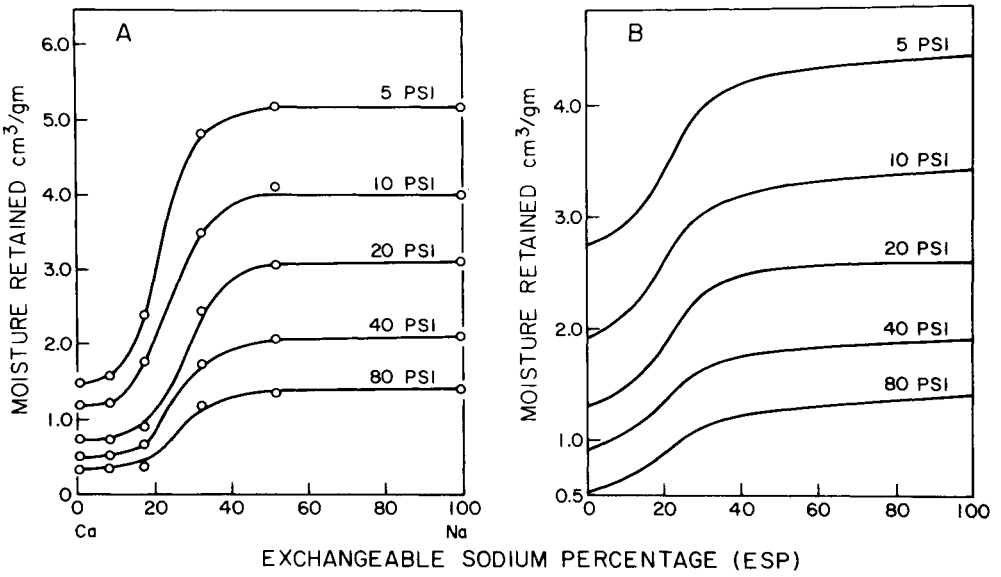

Fig. 3. A. Experimental water retention as a function of ESP and pressure applied. B. Theoretical calculations of water retained by montmorillonite as a function of ESP and swelling pressure (Shainberg et al., 1971).

The experimental results with montmorillonite clays compare favorably with the results of McNeal and Coleman (1966) on soil clays (Fig. 4). They also found that replacing about 15 percent of the adsorbed calcium with sodium had little effect on the swelling of the clay. With a further increase in the ESP of the soil, a very sharp increase in macroscopic swelling was found for the montmorillonitic (Gila and Waukena) soils. At

the high ESP range (>50), the agreement between the predicted (Fig. 3B) and the experimental results is reasonable. Clays with ESP values above 50 swell almost as much as do pure Na-clays.

The fact that clay swelling is affected only slightly by increases in the ESP below ESP 15 suggests that the swelling is not the direct or main cause for the deterioration of the physical properties of soils in this ESP range. The role of clay dispersion in affecting the physical properties of soils is discussed later.

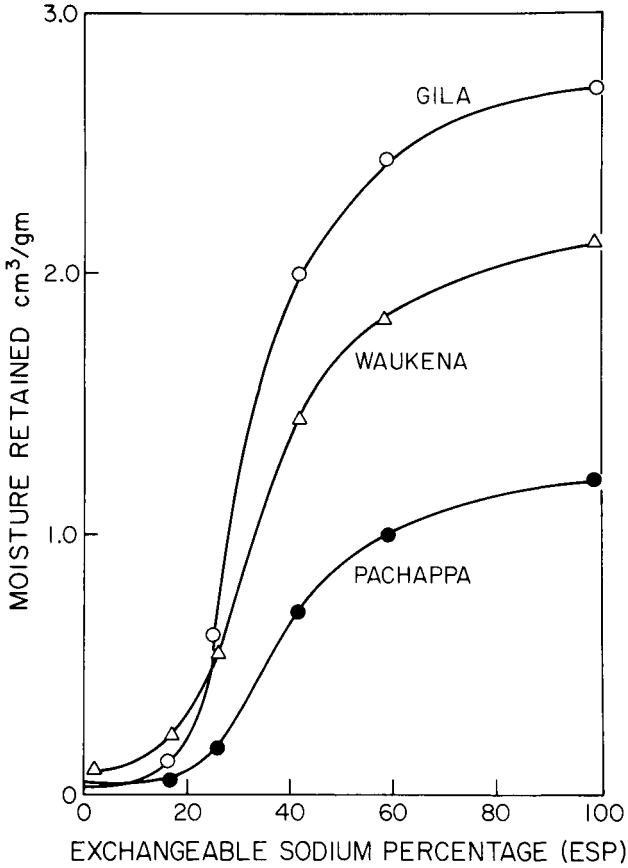

Fig. 4. Macroscopic swelling of extracted soil clays as a function of ESP. (Based on data of McNeal and Coleman [1966] for total salt concentration 12.5 meq/l.)

Adsorbed cation distribution in Na/Ca clays

The platelets in Na-saturated clays are completely separated when in equilibrium with dilute salt solutions, but in Ca-saturated clays they remain in tactoids. A mixture of monovalent and divalent ions is being considered here. In earlier studies of diffuse-double-layer phenomena in bi-ionic systems (Bresler, 1970), the two ions were pictured as being mixed at random throughout the exchange complex. This means that for any ratio of Na to Ca, there are as many sodium ions inside the tactoids as on their external surfaces.

Another hypothesis for the distribution of adsorbed cations in bi-ionic systems was advanced by Mering and Glasser (in Brown, 1961, p. 180), who suggested that "demixing" of the cations occurs such that some interlayer spaces contain mainly Na ions, and

others mainly Ca ions. Mering and Glasser also showed that complete mixing occurs in the first stages of substitution of calcium for sodium. The demixing started at about 30 percent Ca and was complete at 40 percent Ca. At ESP values up to 60, demixing predominated. From x-ray diffraction patterns of clay saturated with mixtures of mono- and divalent ions, McAtee (1961) also inferred that demixing occurs.

Using viscosity and light transmission measurements, Shainberg and Otoh (1968) and Shainberg and Kaiserman (1969) studied the size and shape of montmorillonite particles saturated with a mixture of Na and Ca ions in the adsorbed phase. They found (Fig. 5, curve B) that the introduction of a small percentage of Na into the exchange complex of Ca-tactoids was not sufficient to break the tactoids apart, but introduction of more Na into the exchange complex (over 20 percent) did result in tactoid breakdown. The platelets were completely separated when 50 to 60 percent of the adsorbed Ca had been replaced by Na.

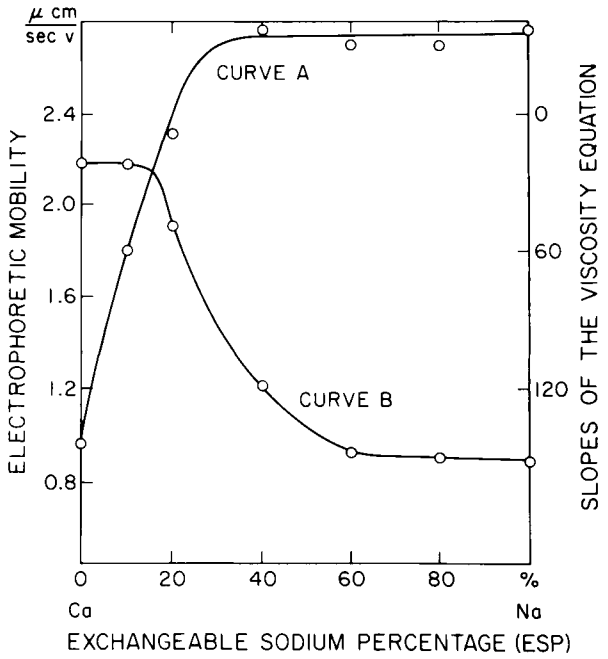

Fig. 5. Dependence of electrophoretic mobility (curve A) and the relative size (curve B) of montmorillonite particles on the exchangeable sodium percentage (ESP) (the relative size is expressed in units of the slope in Einstein equation for the viscosity of suspension). Bar-On et al., 1970.)

An indication of the location of the adsorbed ions in montmorillonite clay saturated with a mixture of mono- and divalent cations was obtained by Bar-On et al. (1970) through measurement of the electrophoretic mobility of clay particles in suspension. The electrophoretic mobility of clay is almost independent of the size of the colloid, but is sensitive to the valency of the cation that forms the diffuse double layer on the external surface. They found (Fig. 5, curve A) that a slight addition of exchangeable Na to Ca-saturated clay had a considerable effect on the electrophoretic mobility of the clay. When the ESP reached a value of about 35 percent, the electrophoretic mobility of this clay was identical to that of pure Na-montmorillonite. The fact that the size of the Ca-tactoids was not affected by a low percentage of exchangeable sodium (curve B) while the electrophoretic mobility increased rapidly supports the demixing model. When only

about 10 percent of the adsorbed Ca has been replaced by Na, most of the adsorbed Na concentrates on the external surfaces of the tactoids. As long as only a small amount of Na exists inside the tactoids, their size is not affected. Conversely, electrophoretic mobility increases very rapidly because the external surface is being saturated mainly with Na. As the ESP of the system increases, Na penetrates the tactoids and causes their disintegration. This stage is reflected in the descending part of curve B. The smaller tactoids formed at this stage already carry mostly adsorbed Na on their newly formed external surfaces, so electrophoretic mobility approaches its maximum value (curve A).

According to the second law of thermodynamics, a system has a spontaneous tendency to proceed to a state of higher entropy and more disorder. Thus, a random distribution of mono- and divalent cations in clay suspensions rather than a demixed distribution might be expected. However, demixing of the adsorbed ions as described above is favored in terms of energy. This can be explained qualitatively and demonstrated quantitatively by equations based on the diffuse-double-layer theory. Qualitatively, the more negative the electrical field is at the clay surface, the higher is the clay's affinity for the divalent ions. Thus, clays with high charge densities (and more negative electrical fields) have higher affinities for Ca and Mg (Babcock, 1963). Similarly, with increases in bulk solution concentration, the negative electrical potential at the clay surface decreases. Thus, with dilution, and consequently more negative electrical potential at the clay surface, affinity for Ca (over Na) increases. Similarly, when clay platelets are brought together and the double layers overlap (Fig. 1), the electrical field at the clay surface becomes more negative and divalent cations are more strongly attracted into the adsorbed phase than are monovalent ions (Keren, 1979). Where tactoids are present, and the distance between the clay platelets within the tactoids is 0.9 nm, strong overlapping of the double layers takes place and the negative surfaces within the tactoids attract divalent Ca ions preferentially over Na ions. Ca concentrates on the internal, and Na on the external surfaces.

This conclusion may be verified in a semiquantitative way by using Eriksson's exchange equation based on the diffuse-double-layer theory.

Eriksson's exchange equation may be written (Babcock, 1963)

$$\frac{\Gamma_1}{\Gamma} = \frac{r}{\Gamma\sqrt{B}} \sinh^{-1} \frac{\Gamma\sqrt{B}}{r + 4V_d\sqrt{m_2}} \tag{1}$$

In this expression, Γ_1 is surface charge neutralized by monovalent ions, Γ is the total surface charge density in me/cm², m_1 and m_2 are the molar concentrations in bulk solution of the mono- and divalent ions respectively, $r = \frac{m_1}{\sqrt{m_2}}$ is called the reduced ratio, B is a constant equal to 1.08×10^{15} cm/mmole for water at 25 °C, $\sinh^{-1}$ represents the inverse hyperbolic sine, and V_d is defined by

$$V_d = \cosh Y_d \tag{2}$$

where Y_d is the dimensionless potential in the plane midway between the platelets. In most applications of Equation 2, the value of V_d is taken as unity (equivalent to specifying that $Y_d = 0$ or that no interaction occurs between adjacent colloid platelets). Evidently, this assumption is not applicable to dominantly Ca-clay systems with up to 30 or 40 percent Na in the exchange complex. In this range of exchangeable Na percentage, tactoids are stable and the dimensionless potential in the midway plane within the tactoids is not zero. Shainberg and Kemper (1966) calculated Y_d values inside the tactoids

as a function of the total electrolyte concentration, and inserted these values in Eriksson's equation (Table 1). The results show that the Ca affinity of internal tactoid surfaces is three times that of external surfaces for an electrolyte concentration of 0.001M. The affinity ratio between internal and external surfaces decreased to 1.5 as the bulk solution concentration increased to 0.1M.

The usefulness of the ion-exchange equations based on the double-layer theory for nonsymmetrical electrolytes and for interacting diffuse double layers was tested by Bresler (1972). He concluded that the double-layer approach worked surprisingly well for soil systems, a conclusion verified experimentally by Bower (1959).

Dispersion and flocculation in Na/Ca clay systems

In a stable clay suspension, dispersed particles collide frequently because of their Brownian motion, but separate again because of diffuse-double-layer repulsion forces. When salt is added to the clay suspension, the particles stick together upon collision, forming flocs which settle. The suspension is then separated into the bottom sediment and a particle-free supernatant liquid. The minimum electrolyte concentration that causes flocculation is referred to as the "flocculation value." The flocculation values of clays depend on counter ion valency. The flocculation values of Na- and Ca-montmorillonite are 12 meq/liter NaCl and 0.25 meq/l $CaCl_2$, respectively (van Olphen, 1977). The flocculation values of Na- and Ca-illites are 40 to 50 meq/l NaCl and 0.25 meq/l $CaCl_2$, respectively (Arora and Coleman, 1979; El-Swaify, 1976; Greene et al., 1978).

Table 1. EFFECT OF PARTICLE COLLAPSE ON EXCHANGE EQUILIBRIA BETWEEN MONO- AND DIVALENT CATIONS. A COMPARISON WITH THE STANDARD DIFFUSE-LAYER THEORY*

Equilibrium solution			Midplane potential, theoretical estimate		Monovalent adsorbed fraction		
					Internal surfaces		External surfaces
Reduced ratio r	Ratio R	Concentration C_o, mol.	Standard Y_d	Mixed-ion Y_d	Standard $\Gamma\sqrt{\Gamma_s}$	Mixed-ion $\Gamma\sqrt{\Gamma_s}$	
1.0	1,000	1.00×10^{-3}	7.6	6.42	0.19	0.341	0.58
1.0	100	1.01×10^{-2}	4.8	4.12	0.26	0.341	0.55
1.0	0	1.1×10^{-1}	2.0	1.82	0.33	0.337	0.49

* See Shainberg and Kemper, 1966.

Oster et al. (1980) studied the flocculation values of montmorillonite and illite suspensions saturated with mixtures of Na and Ca ions in the exchange phase (Fig. 6). They found that a small addition of exchangeable Na to Ca-saturated montmorillonite had a considerable effect on the flocculation value. Similar effects of exchangeable sodium on the electrophoretic mobility of Ca-montmorillonite were reported by Bar-On et al. (1970) (Fig. 5, curve A). According to Verway and Overbeek's stability theory (van Olphen, 1977), these two properties should be interrelated and both depend on the diffusivity of the double layer. Both are explained by the demixing model, which postulates that Na, as the predominant cation on the external surfaces, results in a more stable suspension than in a system where Na and Ca are distributed evenly.

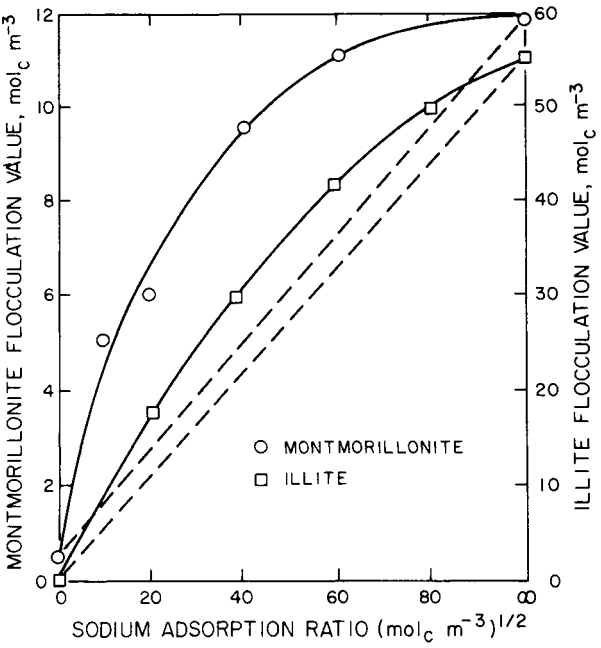

Fig. 6. Flocculation value for Wyoming montmorillonite (API 25) and Fithian illite (API 35) as a function of the sodium adsorption ratio (Oster at., 1980).

A Na-montmorillonite gel has about 250 grams of water per gram of clay, which corresponds to a film thickness of 330 nm per surface. This thickness exceeds the range of diffuse-double-layer forces and/or the van der Waals forces. The open structure typical of Na-montmorillonite gels suggests that edge-to-face attraction forces (van Olphen, 1977) are operating.

The effect of exchangeable Na on the flocculation value of Ca-illite suspensions follows more closely a simple linear relationship than on that of montmorillonite suspensions (Fig. 6). This observation suggests that the demixing of exchangeable ions in Na/Ca-illite is not as pronounced as in Na/Ca-montmorillonite. However, illite with low ESP is more easily dispersed than montmorillonite. The flocculation values for Na/Ca-montmorillonite are 3.0, 4.0, and 7.0 meq/l, and for Na/Ca-illite are 6, 10, and 18 meq/l for ESP values of 5, 10, and 20, respectively. This suggests that soils with illitic clays are more sensitive to dispersion and clay movement than are soils with montmorillonitic clays.

The high flocculation value of the illite as compared with the montmorillonite system is probably due to a smaller edge-to-face attraction force in Na-illite. Consideration of the shape of the Na-illite particle explains this: illite particles consist of platelets stacked together to form particles having 10nm thickness and 120 m^2/g specific surface area (for Fithian illite) compared with a 750 m^2/g specific surface area for montmorillonite. Electron-micrographs reveal that illite particles have irregular surfaces (Greene et al., 1978) and that their planar surfaces are terraced (Quirk, 1978). Upon the close approach of two or more particles, the unavoidable mismatch of terraces leads to poor contact between the edges and the surfaces, resulting in smaller edge-to-face attraction forces, and consequently a higher flocculation value for Na-illite.

The role of the edge-to-face attraction forces in the flocculation process can explain the effect of Na-polymetaphosphate (a peptizer) and high pH on the flocculation value

of clays. Na-polymetaphosphate $(NaPO_3)_{13}$ in concentrations of 0, 1, 10, and 100 mg/l results in flocculation values of Na-montmorillonite of 12, 20, 80, and 120 meq/l of NaCl, respectively (van Olphen, 1977; Oster et al., 1980). Van Olphen (1977) suggested that the polymetaphosphate anions are chemisorbed on the clay particle edges by reacting with the exposed aluminum. Thus, the charge of the edges becomes negative, the edge-to-face bond does not operate, and higher salt concentrations are needed for flocculation.

Increase in pH causes the charge at the edges to change from positive to negative. Consequently, the edge-to-face forces weaken and flocculation value increases. This may explain why increases in soil alkalinity result in increased soil dispersion (US Salinity Laboratory Staff, 1954).

Ion-exchange equilibria

The absolute concentration of various cations in irrigation water is not sufficient for the estimation of potential hazards. An important consideration is the extent of soil ESP increase resulting from sodium adsorption from the water. This increase depends on the ratio of sodium to divalent cations in solution. Cation-exchange equations describe the distribution of cations between the exchanger phase and its equilibrium solution.

Many equations have been proposed to describe the equilibrium distribution between the exchangeable and soluble cations. Because the soil clays contain a mixture of clay mineral types and at least four cationic species (Na, K, Mg, and Ca), a rigorous theoretical description of ionic distribution is difficult. An empirical approach is often preferred.

Several theoretical approaches have been used in deriving the exchange equations. The formal thermodynamic treatment, based on the mass action principle, gives no direct information about the molecular mechanisms and forces operating in the systems. The diffuse-double-layer theory provides a molecular description of the coulombic forces operative in ion-exchange processes. Discrepancies between experimental results and those predicted by the diffuse-double-layer theory facilitate the evaluation of the magnitude of other forces acting in the Na/Ca ionic distribution on clay surfaces. (A description of the diffuse-double-layer theory was given earlier in this section when discussing the effect of tactoid formation on Na/Ca distribution in clays.)

The thermodynamic exchange equilibrium constant for the exchange reaction

$$2NaX + Ca^{2+} = CaX_2 + 2\,Na^+ \tag{3}$$

is represented by

$$K = \frac{(Na^+)^2(CaX_2)}{(NaX)^2(Ca^{2+})} \tag{4}$$

where X is one equivalent of the anionic exchanger and parentheses refer to the activity of the chemical species. The cation activity, a, in solution is related to molar ion concentration, m, and molar activity coefficient, γ, by $a_i = m_i\gamma_i$ (Bohn et al., 1979). With increasing salt concentrations, the activity coefficient for divalent ions decreases more rapidly than that for monovalent cations. Thus, the ratio $\gamma_{Na}:\gamma_{Ca}^{1/2}$ (see below) remains relatively constant over a fairly wide concentration range, and diminishes some of the

objections to the use of concentration (rather than activity) in the SAR equation (Eq. 11, below).

The activity of a species in the adsorbed phase is more difficult to compute. If it is assumed that the exchangeable cation activities are equal to their equivalent fractions, E_i, then the Gaines-Thomas selectivity coefficient, K_{GT}, is obtained:

$$K_{GT} = \frac{E_{Ca} m_{Na}^2 \gamma_{Na}^2}{(E_{Na})^2 m_{Ca} \gamma_{Ca}} \tag{5}$$

If the activity of an exchangeable cation is equal to its mole fraction, N, Equation 4 gives the Vanselow selectivity coefficient, K_V.

$$K_V = \frac{N_{Ca} m_{Na}^2 \gamma_{Na}^2}{(N_{Na})^2 m_{Ca} \gamma_{Ca}} \tag{6}$$

Gaines-Thomas's and Vanselow's equations have been shown to adequately describe monovalent-divalent exchange in clays (van Bladel et al., 1972; Levy and Hillel, 1968; Sposito, 1977; Sposito and Mattigod, 1979). All groups found that for the exchange reaction of Na^+ with M^{2+} (M = Ca, Cd, Cu, Ni, and Zn) on Camp-Berteau montmorillonite, K_V was constant and independent of the exchange composition for $E_M < 0.7$. They found that as the fraction of the divalent ion increased, the affinity for the divalent cation also increased. This observation was explained using the tactoid and demixing model (Eq. 1). When E_M increased above 0.7, tactoids were formed and the affinity of the internal surfaces created by the stacking of platelets for the divalent ions increased (Shainberg et al., 1980).

An ion-exchange equation that has been widely used in salinity studies is the Gapon equation (US Salinity Laboratory Staff, 1954). This equation was derived for the exchange reaction:

$$Ca_{1/2}X + Na^+ = \tfrac{1}{2}\, Ca^{2+} + NaX \tag{7}$$

and the Gapon's equilibrium constant is

$$k_G = \frac{[Ca^{2+}]^{1/2} (NaX)}{[Na^+] (Ca_{1/2}X)} \tag{8}$$

or

$$\frac{E_{Na}}{E_{Ca_{1/2}}} = k_G \frac{[Na]}{[Ca]^{1/2}} \tag{9}$$

where brackets refer to the concentration of cations in the solution (rather than their activities), and the activities of the adsorbed cations are assumed to be equal to their equivalent fractions. Although ion activities are quite different from ion concentrations over the concentration ranges common to salt-affected soils, the corresponding ratio of ion concentrations is of similar magnitude to the corresponding ratio of ion activities

over a fairly wide range. This is because the activity coefficients for divalent ions decrease more rapidly with increasing salt concentrations than do the activity coefficients for monovalent ions, but the square root operation on the divalent ion term results in a fairly constant ratio of monovalent ion activity coefficient to the square root of divalent ion activity coefficient.

Equation 9 is also given as

$$ESR = k_G\ (SAR) \tag{10}$$

where ESR is the exchangeable sodium ratio, SAR is the sodium adsorption ratio, and calcium and magnesium in the exchange and solution phases, respectively, have been lumped together to give

$$ESR = \frac{E_{Na}}{E_{Ca_{1/2}} + E_{Mg_{1/2}}} = \frac{E_{Na}}{CEC - E_{Na}} \tag{11}$$

$$SAR = \frac{(Na)}{(Ca + Mg)^{1/2}}$$

In Equation 11, the cationic concentration in the solution phase (the SAR term) are expressed in mole/m³ (or mmole/l), and E_i, the exchangeable-ion equivalent fraction, is given in meq of i/100 g divided by the cation exchange capacity (CEC) of the clay in meq/100 g. Analysis of a large number of soil samples from the western USA (US Salinity Laboratory Staff, 1954) led to the empirical Gapon equation

$$ESR = 0.01475\ (SAR) - 0.0126 \tag{12}$$

and is presented in Figure 7. However, it should be realized that the slope and the intercept vary with soils depending on their clay mineralogy. If a more accurate estimation of ESR is required, the relationship for each soil should be determined experimentally. For example, the ESR-SAR lines for six soils from Israel are presented in Figure 8 (Levy and Hillel, 1968). The slopes of the lines range between 0.0072 and 0.0169.

Exchangeable sodium percentage (ESP) is related to the SAR of the solution by Equation 13 (US Salinity Laboratory Staff, 1954).

$$ESP = \frac{E_{Na} \cdot 100}{CEC} = \frac{100(-0.0126 + 0.01475\ SAR)}{1 + (-0.0126 + 0.01475\ SAR)} \tag{13}$$

This equation was employed to obtain the average relation between ESP and SAR presented in Figure 9. It is evident that at the ESP range most common in agricultural soils (ESP 0 to 30), the numerical values of the SAR and ESP are almost equal. It is experimentally easier to determine the SAR of the equilibrium solution than the ESP of the soil, so these two parameters are often interchangeable. For example: the US Salinity Laboratory Staff (1954) used the ESP of the soil itself to predict the presence of sodic soil conditions. However, the Terminology Committee of the Soil Science Society of America has recommended that the SAR of the saturation extract be used for sodic soil characterization. Although ESP and SAR are not equal numerically, the committee proposed an SAR value of 15 as the dividing line between sodic and nonsodic conditions (Bresler et al., 1982).

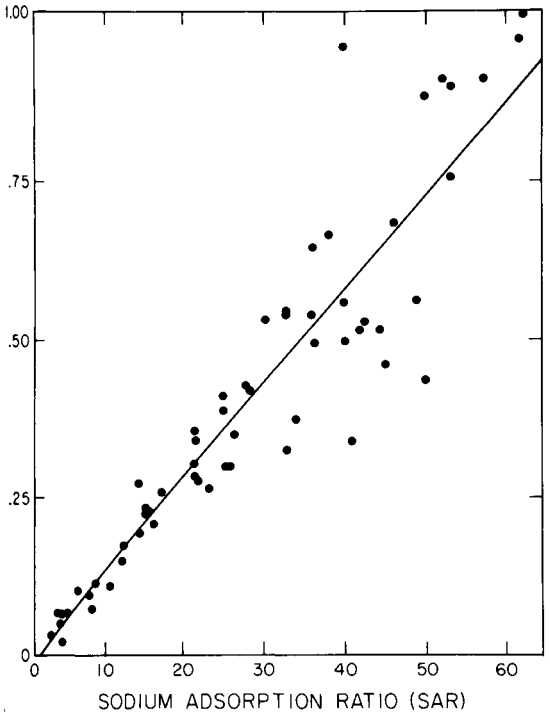

Fig. 7. Exchangeable sodium ratio (ESR) or soil samples from western USA as related to sodium adsorption ratio (SAR) of soil extracts (US Salinity laboratory Staff, 1954).

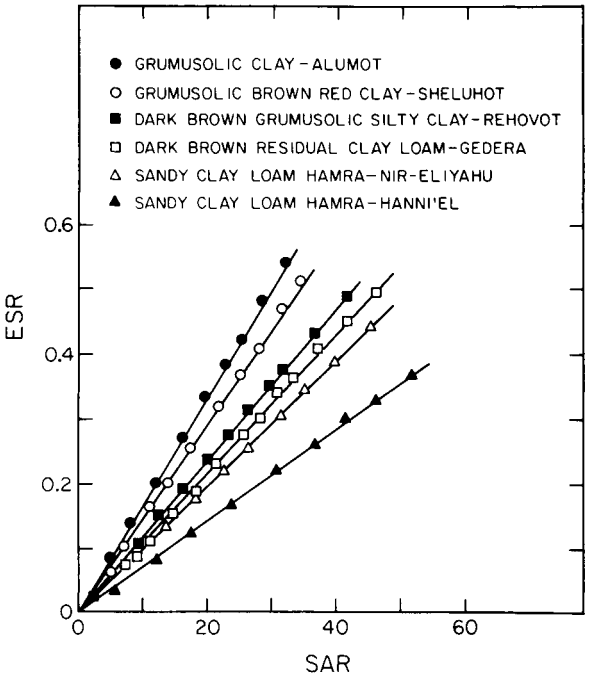

Fig. 8. ESR of six Israel soils as related to the SAR of the soil saturation extract (Levy and Hillel, 1968).

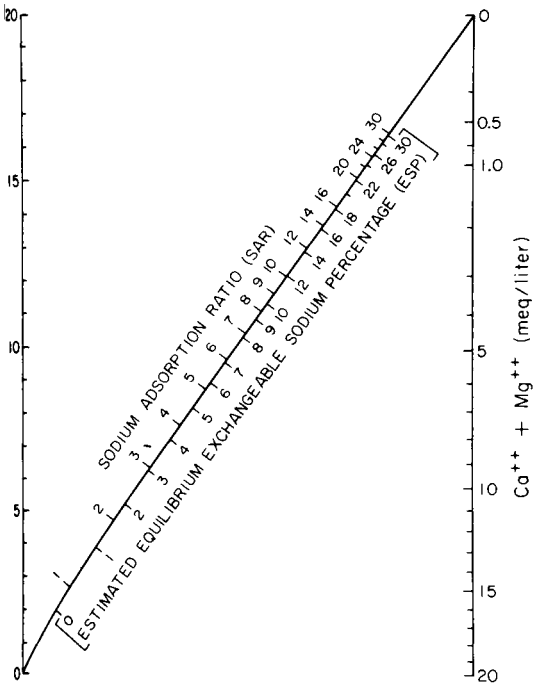

Fig. 9. Nomogram for determining the SAR value of irrigation water and for estimating the corresponding ESP value of a soil equilibrated with this water (US Salinity Laboratory Staff, 1954).

Summary

The swelling and dispersion of clays are affected by the composition of the clays' exchangeable cations. Adsorbed sodium ions form a diffuse double layer, create high swelling pressures, and form single clay platelets which tend to persist in dilute solution.

The low swelling pressure between Ca-clay platelets prevents their dispersion, and, because of the electrostatic attraction between divalent cations and negative surfaces, the platelets stack into tactoids or quasi crystals. Each tactoid consists of several clay platelets, with a 0.45-nm-thick film of water on each internal surface.

In a mixed Na/Ca system, introducing a small amount of sodium (ESP < 15) to a largely Ca-saturated clay has little effect on the swelling. A larger amount of sodium (ESP >15) brings about a breakdown of the tactoid and intensive swelling. Conversely, the electrophoretic mobility and the flocculation value of the clays are very sensitive to small amounts of exchangeable sodium. It should be remembered, however, that clay dispersion can take place only when electrolyte concentration is below flocculation value. These phenomena suggest that when sodium is introduced into calcium-montmorillonite, the sodium ions will concentrate on the external surfaces of the tactoids (demixing), thus increasing sharply the zeta potential of the clay. In a montmorillonitic soil with a given experimental ESP, the ESP at the external surfaces is higher, and the ESP at the internal surfaces lower than the experimental ESP as a result of demixing. This phenomenon explains the pronounced effect of sodium even in clays and soils of low sodicity.

Swelling and/or dispersion of soil colloids alters the geometry of soil pores and thus affects the soil's hydraulic conductivity. Thus, in saline and sodic soils soil cannot be considered an inert porous medium whose hydraulic conductivity can be predicted from its texture or pore size distribution.

THE EFFECT OF SODICITY AND SALINITY ON THE HYDRAULIC CONDUCTIVITY (HC) OF SOILS

Definition and experimental background (saturated flow)

Water flow in soils takes place in accordance with Darcy's equation

$$\frac{Q}{At} = K\frac{\Delta H}{\Delta L} \tag{14}$$

where Q is the volume of water passing through the soil in time t, A is the cross-sectional area of the soil column, ΔH is the hydraulic head difference, ΔL is the soil interval ($\Delta H/\Delta L$ is the hydraulic head gradient), and K is the hydraulic conductivity of the soil.

Hydraulic conductivity (HC) of soil is usually measured empirically and includes soil properties (tortuosity, pore size distribution, etc.) and percolating fluid properties (viscosity). With the objective of eliminating the fluid properties from the conductivity parameter, the intrinsic soil permeability, k, is defined

$$k = \frac{K\eta}{\zeta g} \tag{15}$$

where η is the fluid viscosity, ζ is the fluid density, and g is gravity. A relationship between soil permeability and the properties of the porous medium is given by the Kozeny-Carman equation

$$k = \frac{\Theta^3}{mt^2S^2(1-\Theta)^2} \tag{16}$$

where Θ is the porosity, m is the pore shape factor ($\simeq 2.5$), t is the tortuosity ($\simeq 2^{1/2}$), and S is the specific surface area per unit volume of particles.

Equation 16 is valid for factoring out the effects of fluid viscosity and density on flow through soil, but does not eliminate the effect of fluid properties if the fluid properties alter the soil matrix geometry. Swelling and dispersion of clay in response to the electrolyte composition and concentration of the permeating solution change the pore geometry and affect hydraulic conductivity.

The most relevant data on the effects of sodicity and salinity on soil permeability should be obtained from field studies. However, only rarely has water composition been considered an important variable during hydraulic conductivity measurements in the field. Thus, no field data on the effect of water quality on *in situ* hydraulic conductivity are available.

Most definitive data on the effects of salt concentration and ESP on soil permeability have been provided from laboratory studies. Such studies have usually employed disturbed soils, precluding any direct extrapolation to the field. However, relative values for soil permeability obtained in the laboratory during variations in soil chemical conditions should describe relative behavior in the field under similar conditions, even though absolute hydraulic conductivity values in the field cannot be estimated reliably from laboratory studies. This should be particularly true for the weakly structured soils common to irrigated agriculture and for the regularly tilled surface soils where reductions in soil permeability due to salinity usually are more pronounced (see section beginning page 33). Laboratory data would not provide valid estimates of relative behavior in the field wherever substantial amounts of water could pass through soil discontinuities such as desiccation cracks, gopher holes, or worm channels.

The most common measure of the effect of sodicity and salinity on soil permeability is the determination of relative hydraulic conductivity values in the laboratory. High salt concentration solutions are used to measure the inherent permeability of the soil matrix, and to assure exchange equilibrium between the solution and the adsorbed phase. Thereafter the soil column is leached with a series of percolating solutions of decreasing concentrations, but of appropriate composition to maintain a given exchangeable cation composition (SAR value). The measurements are carried out on fragmented samples which have been dried, ground, passed through a standard sieve, and packed in a uniform manner. Samples are also commonly treated with CO_2 or wetted under suction and treated with a biological inhibitor to remove most of the influence of entrapped air and microbial activity on time-dependent soil hydraulic conductivity changes. A filter paper, or a thin layer of washed sand, is placed on the soil surface to avoid possible disturbance when solutions are added to or taken from the soil column. (A more detailed description of the experiment is presented in publications cited in this section.)

The following distinction is made between hydraulic conductivity and infiltration rate in this report. Hydraulic conductivity measurements are made under conditions where the soil surface is not disturbed, whereas considerable surface disturbance is possible under infiltration rate measurement. Soil disturbance during infiltration rate measure-

ment leads to a surface layer (crust) with different characteristics than the underlying soil, and thus, different water transmission properties. As will be pointed out later, soil surface disturbance has a profound effect on the interaction between water quality characteristics and soil physical parameters. Thus the distinction between hydraulic conductivity (no soil surface disturbance) and infiltration rate (soil surface disturbance) is required for a better understanding of the response of soils to sodic and saline conditions.

Effect of electrolytes on the HC of soils

Permeability of soil to water depends on the ESP of the soil and on the salt concentration of the percolating solution (Quirk and Schofield, 1955; McNeal et al., 1966, 1968; Yaron and Thomas, 1968; Frenkel et al., 1978). The higher the proportion of exchangeable sodium is and the lower the electrolyte concentration of the percolate, the larger is the HC reduction.

To describe the relationship between hydraulic conductivity and solution composition, Quirk and Schofield (1955) developed the concept of "threshold concentration." This is the concentration of salts in the percolating solutions which causes a 10- to 15-percent decrease in the soil permeability at a given ESP value. Soil permeability can be maintained, even at high ESP values, provided that the salt concentration of the water is above a critical (threshold) level. For their particular soil, a noncalcareous silty loam, the threshold concentration values were approximately 0.6 meq/l for the Ca-soil, 2.3 and 9.5 meq/l for soils with ESP of 5.8 and 21 respectively, and 250 meq/l for Na-soil. According to Quirk and Schofield, even Ca-soil may show a reduction in HC, provided that the salt concentration is below 0.6 meq/l. This has been verified by Emerson and Chi (1977), who observed dispersion of soils saturated with divalent cations at salt concentrations below 0.5 to 2 meq/l. When water of salinity below 1 meq/l (rain or snow water) is applied to a soil, even Ca soil and soils of low ESP may disperse and lose some permeability. This will be discussed later in detail.

The basic approach of Quirk and Schofield (1955) has been extended to a large number of additional soils by workers such as McNeal and Coleman (1966) and Cass and Sumner (1982*a*,*b*). In most of these studies, the electrolyte concentrations of the percolating solutions were maintained above 3 meq/l (e.g., McNeal et al., 1966, 1968; Yaron and Thomas, 1968; Rhoades and Ingvalson, 1969). Typical effects of salt concentration and ESP on soil HC are demonstrated in Figure 10 (from McNeal and Coleman, 1966). These studies showed that soils responded differently to the same combination of salt concentration and ESP—thus, a unique threshold concentration exists for each soil (Fig. 11). Soil properties significantly affecting the response of soil permeability to sodic and electrolyte concentration seemed to be clay content, clay mineralogy, iron oxide or aluminum oxide content, organic matter content, and bulk density. However, other soil properties such as primary mineral and $CaCO_3$ content are very important in determining the response of sodic soils to leaching with distilled water (simulating rain and snow water). In the following discussion of the effect of soil properties on the response of soils to sodic conditions a separation is therefore made between percolating solutions of low concentration (below 3 meq/l) and those having concentrations above 3 meq/l. This separation is justified in terms of the mechanisms responsible for the HC decrease.

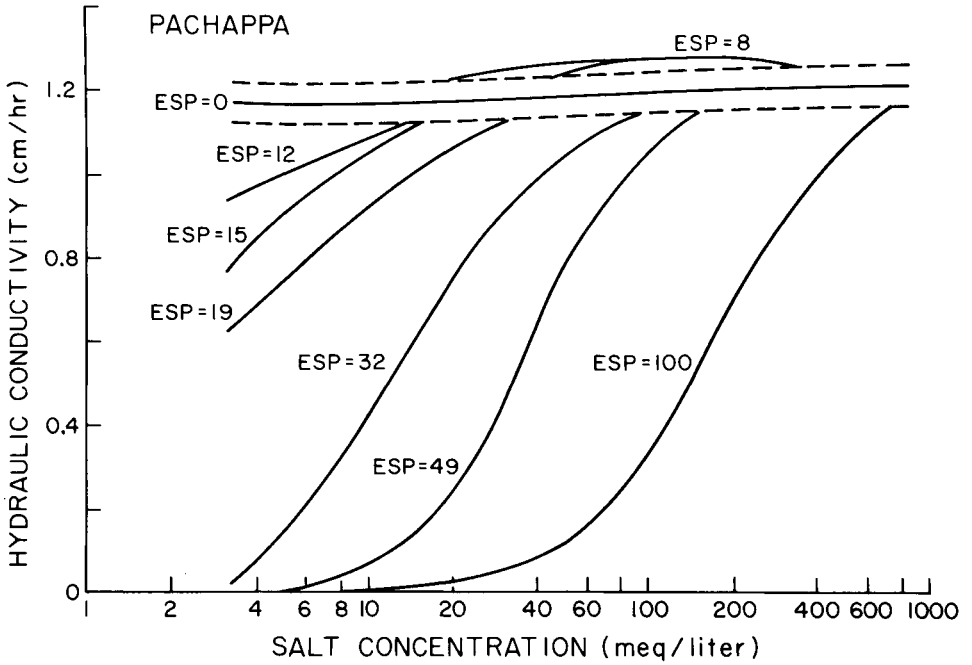

Fig. 10. Hydraulic conductivity of Pachappa sandy loam as related to salt concentration and ESP (McNeal and Coleman, 1966).

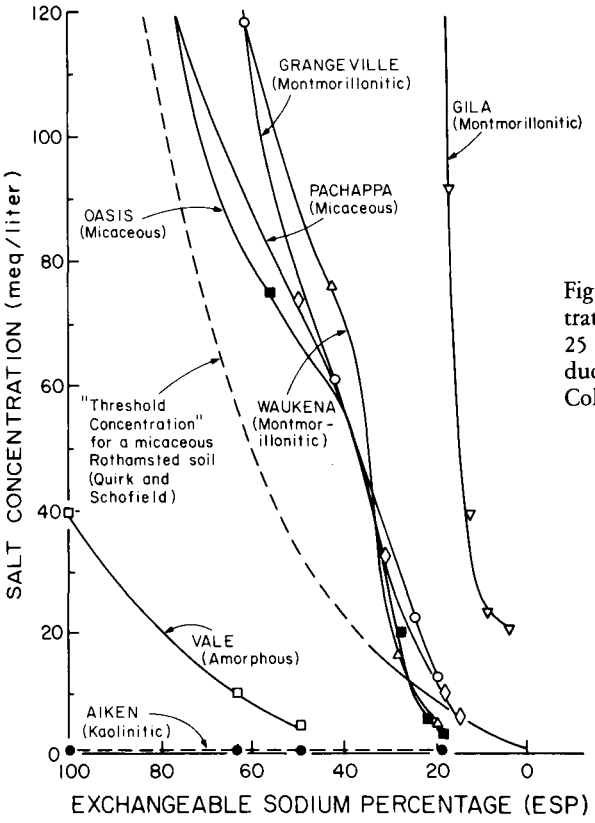

Fig. 11. Combinations of salt concentration and ESP required to produce a 25 percent reduction in hydraulic conductivity for selected soils (McNeal & Coleman, 1966).

Two main mechanisms have been proposed to explain the soil permeability decrease. Quirk and Schofield (1955) suggested that the swelling of clay particles, which increases with an increase in clay sodicity, could result in blocking or partial blocking of the conducting pores. Rowell et al. (1969) showed that initial reductions in HC could be attributed to swelling. McNeal et al. (1966) found a linear relationship between HC reduction and macroscopic swelling of extracted soil clays with correlation coefficients varying from 0.94 to 0.97. They were able to relate clay-swelling measurement to theoretically calculated interlayer swelling values.

Deflocculation and dispersion were proposed as the second main mechanism (Quirk and Schofield, 1955). This mechanism operates when the charged plates, which are moving apart in the process of swelling, have separated enough so that attractive forces are no longer strong enough to oppose repulsive forces and the platelets can move by an external force. The plugging of soil pores by dispersed clay particles is another mechanism by which the HC of sodic soils is reduced. The importance of dispersion in affecting soil permeability was recognized by Felhendler et al. (1974), Frenkel et al. (1978), Pupisky and Shainberg (1979), Rhoades and Ingvalson (1969), and Shainberg et al. (1981*a*,*b*).

The differences between swelling and dispersion processes are very important. Clay swelling is not greatly affected by low ESP values (below 10 to 15) but increases markedly as the ESP increases above 15 (see section beginning page 2). Clay dispersion is very sensitive to low levels of sodicity, and increases markedly at the low ESP range. Deflocculation and dispersion of clay are possible only at solution concentrations below the flocculation value, but swelling is a continuous process and decreases gradually with increases in solution concentration. The flocculation values of montmorillonite clay with ESP values of 0, 10, and 20 are 0.25, 4, and 7 meq/l. As these numbers were obtained in test tubes and with reference clays, their numerical values may vary in soils. Nevertheless, it may be concluded that the dispersion mechanism in most agricultural sodic soils can operate only at very dilute salt solutions, and under these conditions it is operative even at low ESP values. Although swelling is essentially a reversible process—reduction in permeability which can be reversed by adding electrolytes or divalent cations to the system—dispersion and particle movement are essentially irreversible and cause the formation of an impermeable clay layer in the soil profile.

Figure 12 shows the importance of particle dispersion and movement as compared with swelling. In this experiment, a loamy soil of 20 percent clay (montmorillonitic) and a clay soil of 50 percent clay (montmorillonitic) were equilibrated by leaching with SAR 10 and 20 meq/l solutions. Following equilibration, the soils were leached with distilled water. The loamy soil was completely sealed upon leaching with distilled water, but the clay soil maintained a relative HC of about 20 percent. This observation contradicts the generalization that soils having a high content of expansible clays are more sensitive to sodic conditions than are those of lower clay content—although these results may be explained by the two mechanisms described above. In the calcareous clay soil (Grumosol), salt concentration in the soil solution was maintained at 4 meq/l by dissolution of carbonates. At ESP 10, this concentration prevented clay dispersion, and the HC decrease was due only to clay swelling. Conversely, in the noncalcareous loamy soil (Netanya), soil solution concentration was lowered below 1 meq/l; thus clay dispersion occurred and the soil was sealed completely. These processes, too, will be discussed later in detail.

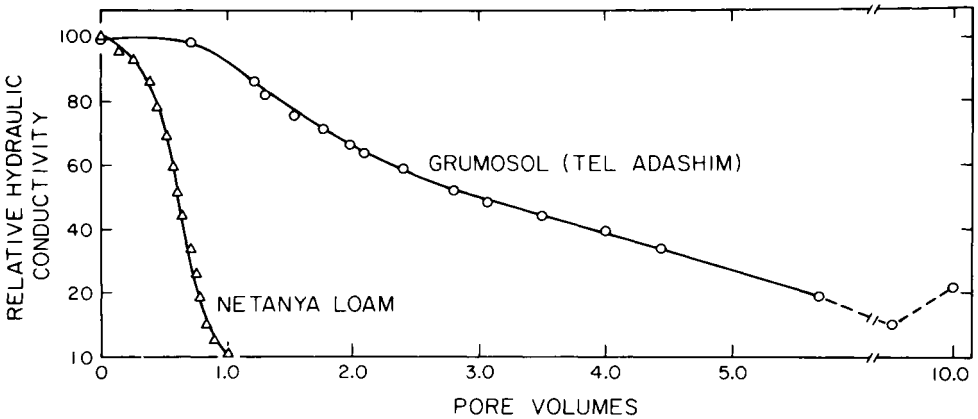

Fig. 12. Hydraulic conductivity of two montmorillonitic soils having an ESP of 10 and leached with distilled water (Shainberg and Oster, 1978).

Soil properties affecting the response of sodic soils to electrolyte solutions of medium to high salinities

Effect of soil texture. McNeal et al. (1968) showed that for soils having variable clay content but nearly uniform clay mineralogy fraction (a clay fraction consisting of 42 percent montmorillonite, 29 percent mica, 16 percent quartz plus feldspars, and 13 percent of other species including chlorite, vermiculite, and amorphous minerals), HC reductions caused by increases in exchangeable sodium or decreases in total electrolyte concentration were greater for soils having higher clay content. This effect was in addition to the decreases in absolute HC associated with increases in clay content. Frenkel et al. (1978) also reported that the susceptibility of soils to sodic conditions increased with an increase in clay content.

Effect of clay mineralogy. McNeal and Coleman (1966) and Yaron and Thomas (1968) concluded that the most labile soils were those high in 2:1 layer silicates, especially montmorillonite, and the least labile were those high in kaolinite and sesquioxides. McNeal et al. (1966) found a good correlation between relative soil permeability and the swelling of extracted soil clays, which in turn was affected by soil solution concentration and composition. Several workers have used predictions of montmorillonite swelling to predict HC response to solutions of various salt concentrations and ionic compositions. Lagerwerff et al. (1969) coupled predictions of clay swelling with the Kozeny-Carman equation (Eq. 16) to estimate HC decreases during expansion of the soil matrix. However, being unable to calculate *a priori* the electrical potential at the midplane between swelling soil particles in mixed-cation systems, they used the midplane electrical potential as an empirical index which was allowed to vary until a reasonable fit was provided between theoretical and experimental values. A semi-empirical approach was taken by McNeal et al. (1968), who described saturated permeability with the equation

$$1 - r = fx^n/(1 + fx^n) \tag{17}$$

where r is the ratio of HC of a given solution to HC of divalent cation-saturated soil at high salt concentration solution, f is an empirical constant characteristic of the soil, x is a colloid swelling factor (based upon a demixed-ion swelling model), and n is a factor which varies with the estimated soil ESP.

Russo and Bresler (1977*a,b*) tested the effects of mixed Na-Ca salt solutions on soil water diffusivity and unsaturated HC for a loamy soil from Gilat, Israel. They found that for any given moisture content, the diffusivity and HC decreased either as the soil solution salt concentration decreased or as the SAR increased. The negative effect of high SAR and low solution concentration decreased with decreasing soil water content. Russo and Bresler (1977*a,b*) developed a model to predict the dependence of HC on moisture content, SAR, and solution concentration. The model was based on the diffuse-double-layer theory for swelling and the Marshall equation (Marshall, 1958) for estimating HC as a function of moisture content. They obtained relatively good agreement between model calculations and experimental data, but they also had an adjustable parameter (the number of platelets in a tactoid) which was allowed to vary until a reasonable fit was obtained between theoretical and experimental values.

Frenkel et al. (1978), studying the response of kaolinitic soils to sodic conditions, found that the HC values of these soils were less affected by the ESP and the electrolyte concentration than were those of montmorillonitic soils. Similarly, Rhoades and Ingvalson (1969) concluded that a much higher ESP was needed to appreciably reduce HC for vermiculitic as compared to montmorillonitic soils.

It is evident that swelling was considered the main mechanism for HC decreases as the ESP of the soil increased and the salt concentration decreased. The models and experimental results showed that the effect was normally greatest for soils having high contents of swelling minerals (such as smectites) and for soils having high clay contents.

Effect of free sesquioxides. Studying the effect of sodicity and electrolyte concentration on a group of soils from the Hawaiian Islands, McNeal et al. (1968) found that despite the high textural clay contents in these highly weathered and hydrous oxide-stabilized soils, the effect of ESP and electrolyte concentration on the HC was essentially negligible until iron- and aluminum-containing hydrous oxides were chemically removed. El-Swaify (1973), studying tropical soils, arrived at similar conclusions. McNeal's (1968) conclusion was that the cementing action of iron oxides prevented swelling and dispersion. However, Deshpande et al. (1968) concluded that aluminum oxides, rather than iron oxides, had the greatest effect on soil stability during leaching with sodic water.

Effect of soil density. Frenkel et al. (1978) studied the effect of soil bulk density on the HC decreases resulting from sodic and electrolyte concentration, and concluded that soil sensitivity to excessive ESP and low electrolyte concentration increased with increase in bulk density.

Response of sodic soils to leaching with distilled water and dilute salt solution

"Despite the obvious effect of electrolytes on soil permeability, the main emphasis has been given to the exchangeable sodium percentage and [electrical] conductivity of the irrigation water as an index of the degree to which sodium will occupy exchange positions on the clay surface. Soils in arid and semiarid regions often have high ESP before being irrigated and for this reason the quantity of electrolyte in the irrigation water should be given greater consideration."

Although the above paragraph was written in 1955 (Quirk and Schofield), it is still pertinent today. Even in arid regions where irrigation is essential for maintaining agricultural production, intermittent rain may lower the electrolyte concentration below the threshold value typical to the soil. In Israel, where water having a SAR value of 26

and an electrical conductivity of 4.6 dS/m is used for commercial irrigation (Frenkel and Shainberg, 1975), there are no permeability problems during the irrigation season (summer)—electrolyte concentration in the irrigation water is enough to prevent the dispersive effect of the sodium. In the rainy season (winter) however, the soil disperses, soil permeability is reduced, crust is formed, and rain infiltration is very low. Similarly, in some areas in the Central Valley of California waters of very low salinity (0.05 to 0.2 dS/m) are used for irrigation. The electrolyte concentrations of these waters are below the threshold concentration typical to the soils and permeability problems often arise (Mohammed et al., 1979).

Felhendler et al. (1974) measured the HC of two montmorillonite soils (a sandy loam and a silt loam) as functions of the SAR and salt concentration of the percolating solution, and found (Fig. 13) that both soils were only slightly affected by the SAR of the percolating solution up to SAR 20, as long as the concentration of the percolating solution exceeded 10 meq/l. However, when the percolating salt solution was displaced by distilled water (DW) simulating rainfall, the response of the two soils differed drastically. The HC of the silty soil dropped to 42 and 18 percent of the initial value for soils having ESP values of 10 and 20, respectively. The HC of the sandy loam soil dropped to 5 and 0 percent of the initial value for the same conditions, respectively. They also noted that the clays in the sandy loam soil were mobile and appeared in the leachate, whereas no clay dispersion took place in the silty loam soil. They postulated that the HC response was associated with the potential for clay to disperse when the soil was leached with distilled water. Clay in the sandy loam soil dispersed strongly when the soil was leached with distilled water, but clay in the silty loam soil did not. However, they presented no hypothesis to explain why clay particles of similar mineralogy will disperse in one soil and not in the other.

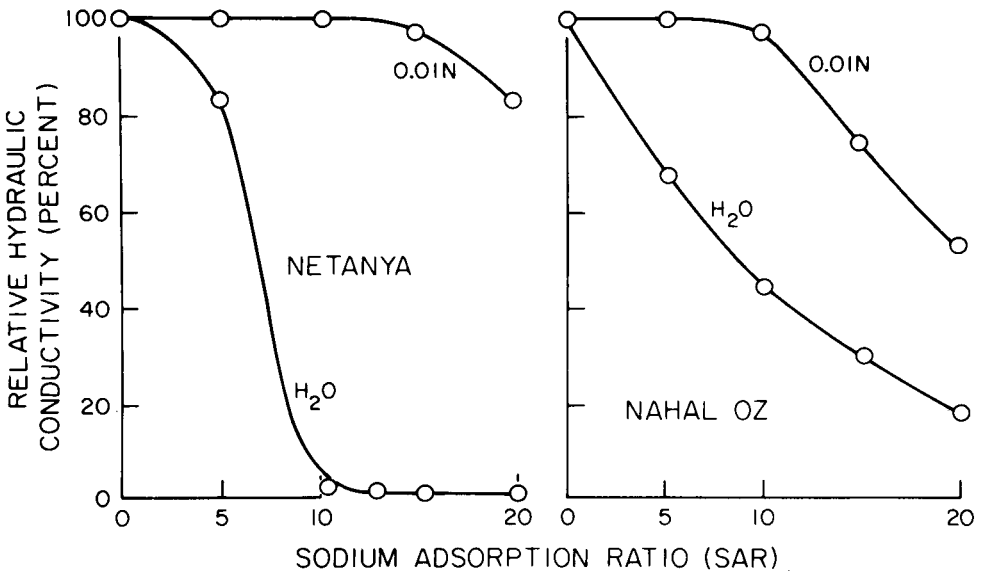

Fig. 13. Hydraulic conductivity of a sandy loam (Netanya) soil and a silty loam (Nahal-Oz) soil as a function of the SAR and the concentration of the leaching solutions (Felhendler et al., 1974).

Effect of low electrolyte concentration on clay dispersion and HC of a sodic soil. Figure 14 shows the effect of displacing 0.01 N solutions of SAR 10, 15, 20, and 30 with distilled water or salt solutions of 1, 2, or 3 meq/l on the relative HC of Fallbrook soil-sand mixture. It is evident that when leached with distilled water, even a low ESP was enough to appreciably reduce the HC of the Fallbrook soil. Electrolyte concentration of 2 meq/l in the percolating solution prevented the adverse effect of ESP 10 on the HC of this soil. The adverse effect of 15 percent Na in the exchange complex was prevented by a solution of 3 meq/l.

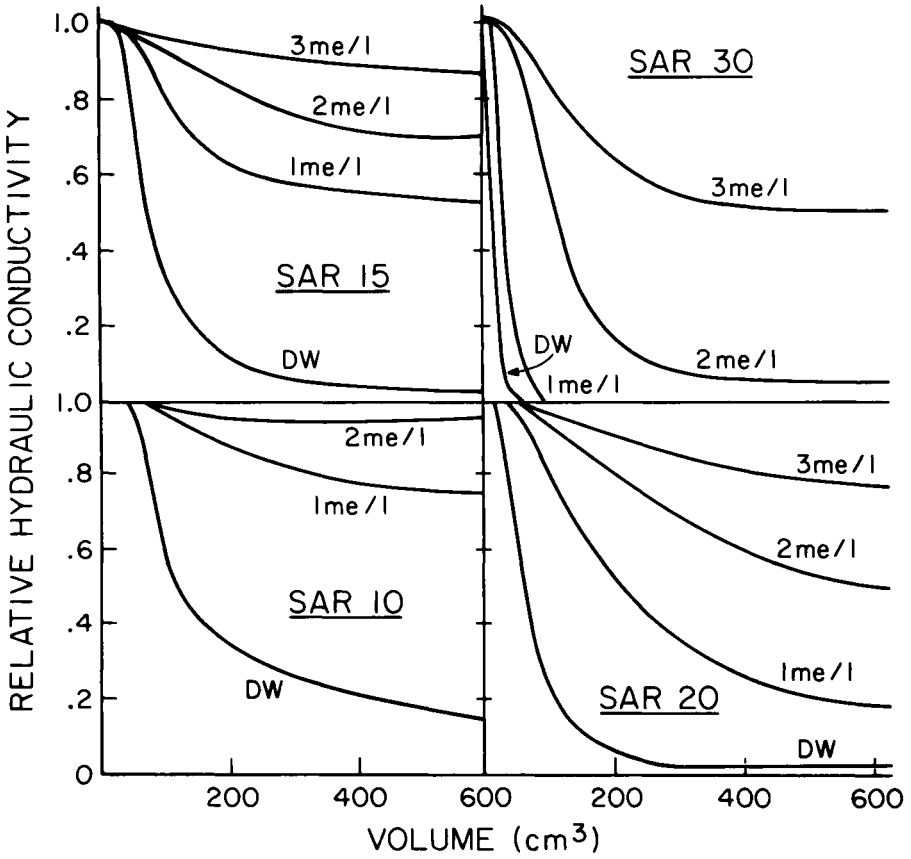

Fig. 14. Relative HC of Fallbrook soil-sand mixture, equilibrated with 0.01 N solutions of SAR 10, 15, 20, or 30, and leached with distilled water (DW) or salt solutions of 1, 2, or 3 meq/liter (Shainberg et al., 1981*a*).

Figure 15 shows the effect of the solution concentration on the concentration of clay in the Fallbrook soil effluent. It is clear that clay dispersiveness and movement are very sensitive to the electrolyte concentration in the percolating solution. For example, in soil equilibrated with SAR 15 solution, leaching with distilled water or 1 or 2 meq/l solutions resulted in peak clay concentrations of 1.0, 0.1, and 0.02 percent clay, respectively, in the effluent.

Figure 16 summarizes the response of the Fallbrook soil to a combination of sodicity and electrolyte concentration. When the percolating solution concentration was maintained at $\geq$3 meq/l, no reduction in HC took place until the SAR exceeded 12. For solution concentrations of 2, 1, and 0.5 meq/l, threshold SAR values were 9, 6, and 4,

respectively. When Fallbrook soil was leached with distilled water, the detrimental effect of exchangeable sodium started at an ESP value of about 1 percent. Once the threshold ESP was exceeded the HC was always very sensitive to further increases in exchangeable sodium. The more dilute the leaching solution was, the more sensitive was the soil to an incremental increase in the ESP beyond the threshold value.

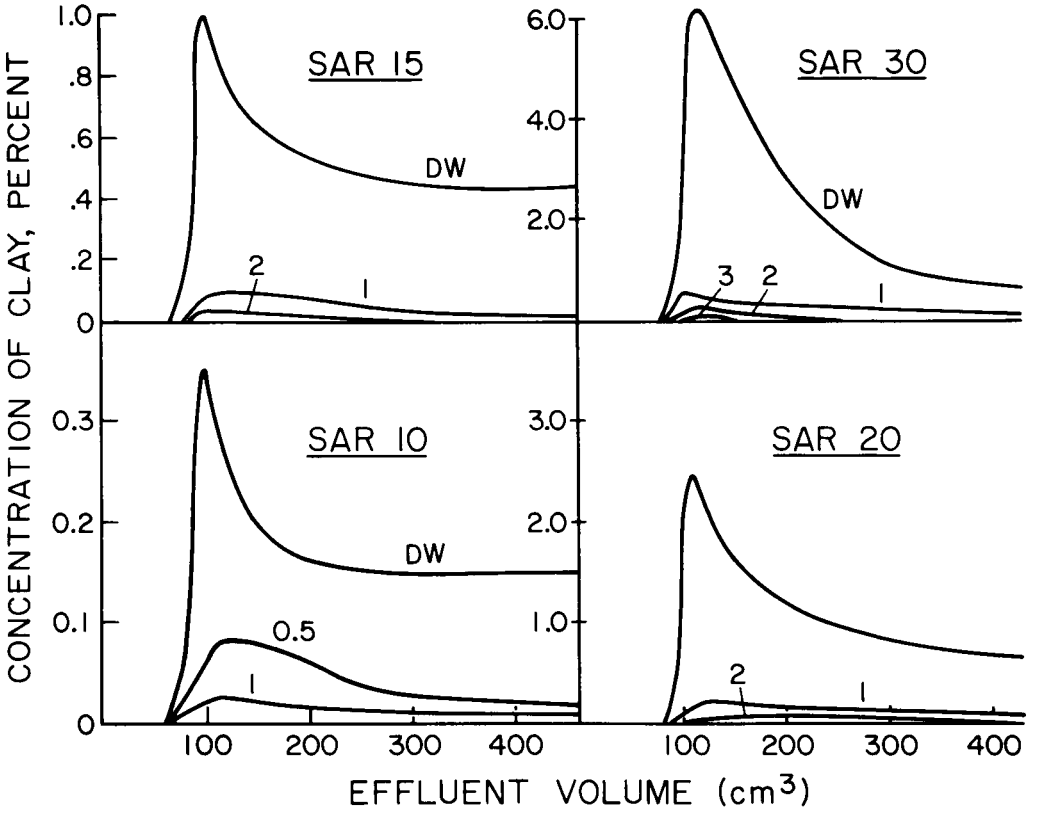

Fig. 15. Concentration of clay in effluent of the low salt concentration and distilled water (DW) leachates from the Fallbrook soil mixture equilibrated with 0.01 N solutions of SAR 10, 15, 20, and 30 (Shainberg et al., 1981*a*,*b*).

The effect of ESP on the HC of soils leached periodically with rainwater should be reemphasized; even a low ESP may enhance loss of HC of the soil. Five percent sodium on the exchange complex of a soil like Fallbrook reduced the HC of the soil to 20 percent of its initial value.

The high sensitivity of soils to low levels of exchangeable sodium when leached with "high quality water" can explain the results of McIntyre (1979), who concluded, following the measurement of the HC of 71 Australian soils, that an ESP value of 5 should be accepted for Australian soils as separating sodic from normal soils. The US Salinity Laboratory Staff (1954) suggested an ESP value of 15 to designate sodic soils. In McIntyre's studies, the electrolyte concentration in the tap water used to determine the soils' HC was 0.7 meq/l, whereas in the US Salinity Laboratory studies it was above 3 meq/l. Thus it is expected that when high quality water (e.g., rain or snow water) is used, an ESP of 5 can be detrimental to the physical properties of soils. When waters of higher salinities are used (>3 meq/l), only ESP values higher than 15 will cause soil damage.

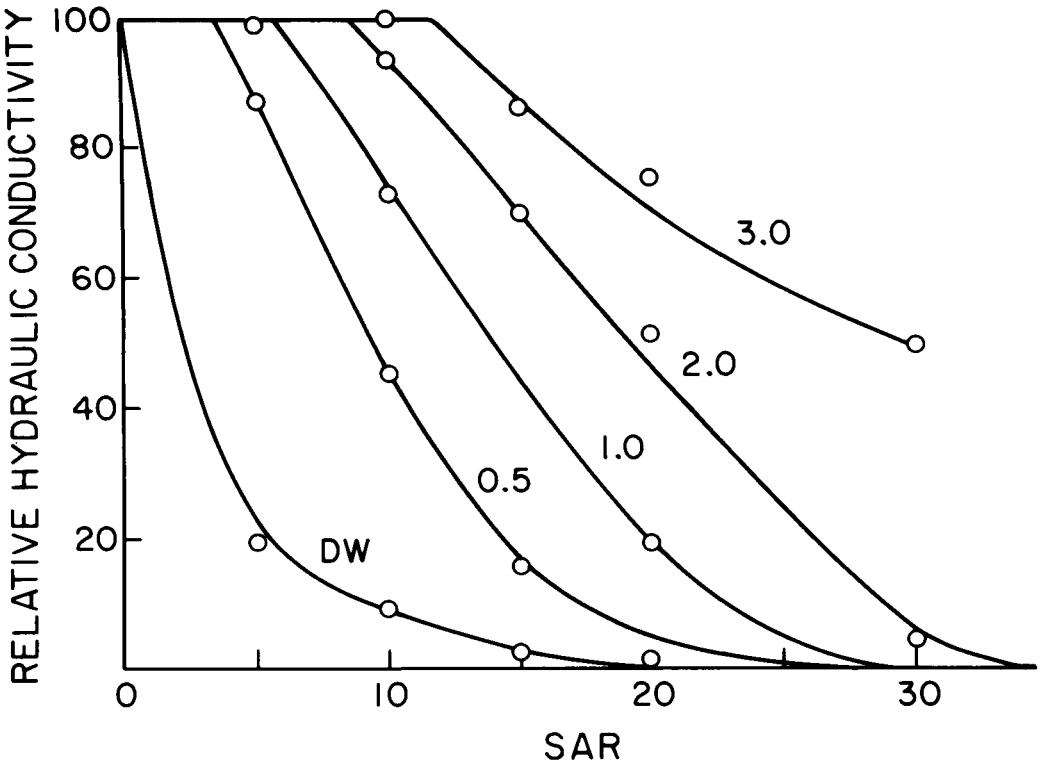

Fig. 16. Effects on Fallbrook soil-sand mixture's HC of the ESP of distilled water (DW) and various soil solution salt concentrations (Shainberg et al., 1982*a,b*).

Effect of soil minerals' weathering on their response to leaching with distilled water. Arid land soils release 3 to 5 meq/l of calcium and magnesium to solution as a result of the dissolution of plagioclase, fledspars, hornblende, and other minerals (Rhoades et al., 1968). The solution composition of a calcareous soil at a given ESP when placed in contact with distilled water can easily be calculated (Shainberg et al., 1981*b*). The $CaCO_3$ dissolves and replaces sodium from exchange sites until the solution is in simultaneous equilibrium with exchange sites and with the $CaCO_3$ solid phase. Assuming that the apparent $CaCO_3$ solubility product is $10^{-8.0}$ and in equilibrium with a CO_2 pressure of $10^{-3.5}$ Atm, the electrical conductivities of the solutions in equilibrium with soils having ESP 5, 10, and 20 are 0.4, 0.6, and 1.2 dS/m, respectively (Shainberg et al., 1981*b*). In a study of the dissolution of two salt-free soils from Israel (Netanya sandy loam and Nahal-Oz silty loam), Oster and Shainberg (1979) found (Fig. 17) that the concentrations of the soil solutions ranged between 0.5 and 3 meq/l respectively within 2 hours after the soils were mixed with distilled water. The more weathered Netanya sandy loam was from a region having an annual precipitation of 600 mm, whereas the Nahal-Oz calcareous silty loam was from the northern Negev where the annual precipitation was 400 mm. The leaching intensity in a region determines the degree of soil weathering.

The sensitivity of sodic soils to solutions of low electrolyte concentrations led Shainberg et al. (1981*a,b*) to hypothesize that a major factor causing differences among various sodic soils in their susceptibilities to HC decreases when leached with low-electrolyte water was their rate of salt release by mineral dissolution. Mineral dissolution

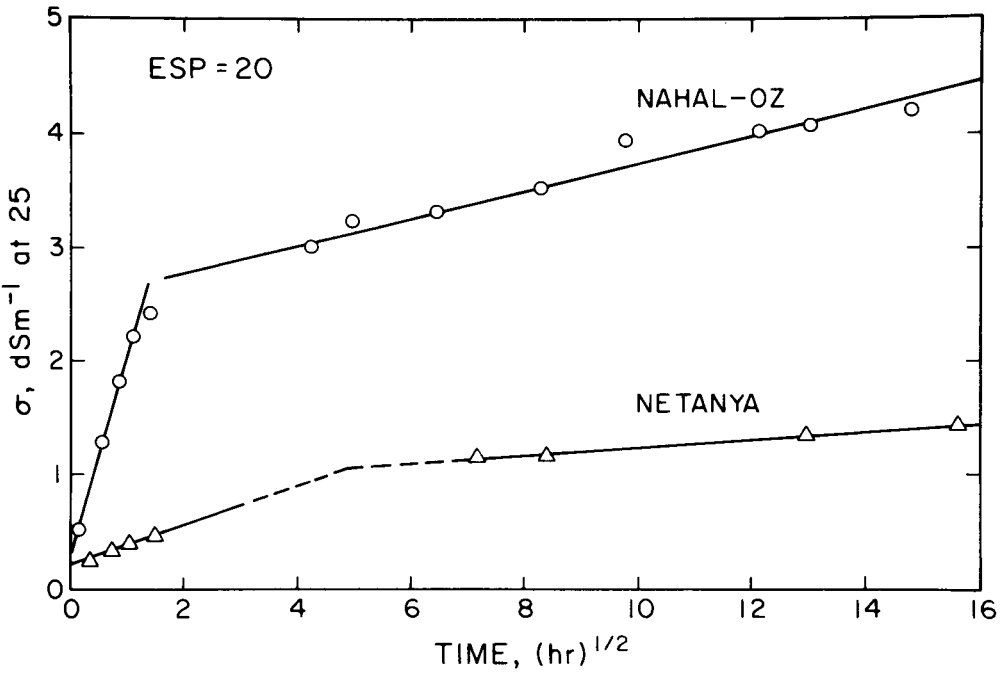

Fig. 17. Relationship between specific conductance, σ, and time, for aqueous suspensions of two soils (Oster and Shainberg, 1979).

determines the electrolyte concentrations of percolating solutions. They postulated that sodic soils containing minerals ($CaCO_3$ and a few primary minerals) that readily release soluble electrolytes will not readily disperse when leached with distilled water (simulating rain water) at moderate ESPs, because they will maintain high enough salt concentrations in the soil solution ($\simeq 3$ meq/l) to prevent clay dispersion. Additionally, the ESPs in these soils will be reduced, because most of the released salts are Ca and Mg. Conversely, salt concentration in the soil solution of soils lacking readily weatherable minerals will be below a threshold concentration (the flocculation value), and the soils will be more susceptible to clay dispersion. Lodgement of dispersed clay particles in conductivity pores may markedly decrease the HC of such soils.

The data presented in Figures 12 and 13 on two Israel soils supported the hypothesis. Soil that was more stable chemically (Netanya sandy loam) was also more susceptible to sodic conditions when leached with distilled water. The hypothesis is further supported by measurements of clay dispersivity, HC, and the mineral weathering properties of three California soils (Figs. 18, 19, and 20). The chemically stable Fallbrook soil was the most sensitive soil to exchangeable Na effects on clay dispersion and loss in HC in spite of the presence of sesquioxides and kaolinite in its clay fraction. Gila and Pachappa soils, which have higher mineral dissolution rates, were less affected by exchangeable Na.

Clay will be in the leachate only in sandy soils (Fig. 20). When the soil's clay content is high, the conducting pore size is usually small (resulting in low HC), and the dispersed clay moves only short distances before it plugs the pores. Thus, in loamy and clay soils the dispersion mechanism still operates, but no macroscopic movement of the clay particles is observed. When the clay soil is mixed with sand, the dispersion mechanism is evident.

The different capacities of soils for salt release and dispersion are important in affecting crust formation under rainfall conditions. Soils having moderate ESP levels may maintain reasonable physical properties through most of the profile and yet be susceptible to dispersion near the surface. This is because the soil solution electrolyte concentration may be insufficient at the surface to maintain physical structure.

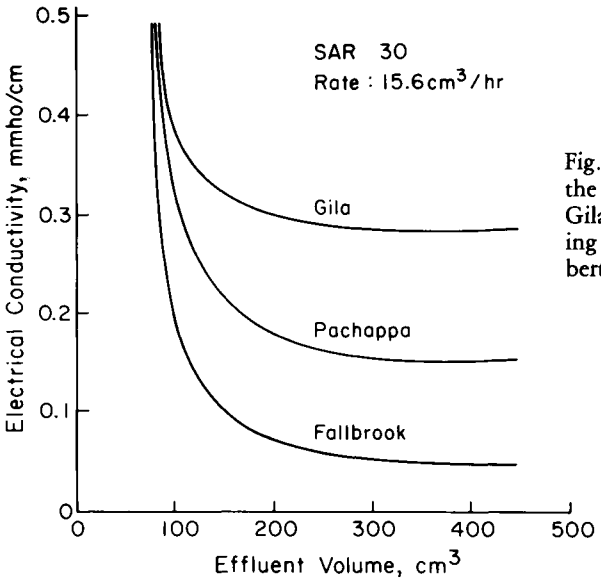

Fig. 18. Effect of effluent volume on the EC of effluents from columns of Gila, Pachappa, and Fallbrook soils during leaching with distilled water (Shainbert et al., 1981*b*).

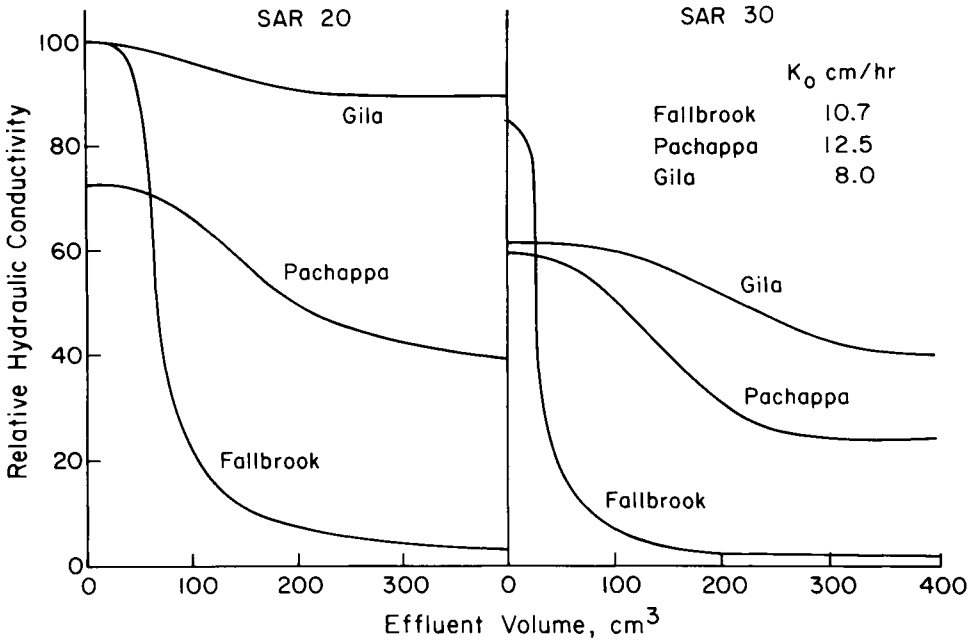

Fig. 19. Effect on relative HCs during leaching of ESP 20 and ESP 30 soil mixtures with distilled water (Shainberg et al., 1981*b*).

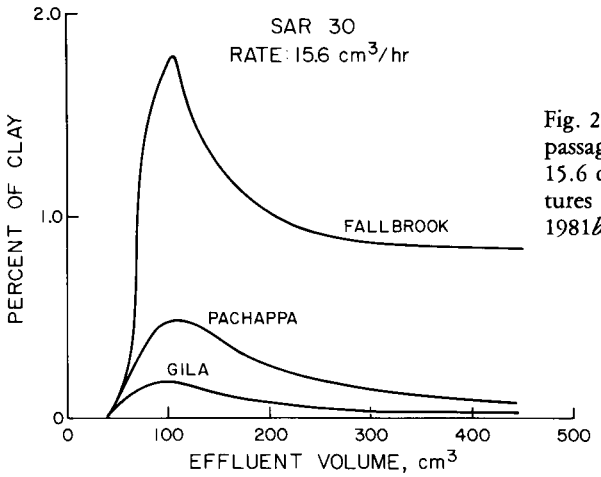

Fig. 20. Effluent clay concentration with passage of distilled water at the rate of 15.6 cm^2 $hour^{-1}$ for the three soil mixtures at ESP = 30 (Shainberg et al., 1981*b*).

Effect of exchangeable Mg and soil amendments on the response of soils to leaching with distilled water

Specific effect of exchangeable magnesium on the HC of sodic soils. The US Salinity Laboratory Staff (1954) grouped Ca and Mg together as similar ions beneficial in developing and maintaining soil structure. However, it has often been suspected that Mg can cause soil structure to deteriorate and develop a "Mg solonetz" (Ellis and Caldwell, 1935). Laboratory studies also imply that Mg is a deleterious ion in some circumstances. For instance, McNeal et al. (1968) showed that mixed Na/Mg soils developed lower hydraulic conductivities (HC) than did Na/Ca soils under similar conditions. In recent work (Chi et al., 1977; Emerson and Chi, 1977; Abder-Rahman and Rowell, 1979) a distinction was made between the direct effect of exchangeable Mg in causing decreases in HC, which has been termed a specific effect, and the ability of Mg in irrigation waters to cause higher exchangeable Na levels to build up in soil. In the work referred to above, Mg was shown to have a specific effect in soils dominated by illite, but not in montmorillonitic soils. The montmorillonite results were obtained for both pure montmorillonite clay and for heavy-textured soil having high organic matter content, where clay swelling was the dominant process causing structural deterioration. In more recent work (Rowell and Shainberg, 1979), exchangeable Mg appeared to have no specific effect on soil HC for the A- and C-horizons of a sodic sandy loam soil dominated by montmorillonite and kaolinite clay minerals leached with solutions having electrolyte concentrations of ≥ 10 meq/l. However, there was evidence of a specific effect of Mg when the soils were subsequently leached with distilled water (i.e., under conditions favorable for clay dispersion).

The effect of low electrolyte concentration and mineral weathering on clay dispersion and the HC of sodic soils was discussed in the previous section. Alperovitch et al. (1981) used a similar approach to explain and predict in which soils and under what conditions exchangeable magnesium would have dispersive properties. Figure 21 shows some of their data. Kesufim, Netanya, and Hermon are all sandy loam soils having montmorillonite as the dominant clay. Kesufim is a calcareous soil which releases electrolytes into the soil solution, Hermon is a chemically very stable soil, and Netanya occupies an intermediate position. The data in Figure 21 suggested to Alperovitch et al. (1981) that exchangeable Mg does not have a specific effect on the hydraulic conductivity (HC) and

clay dispersion of calcareous soils. Conversely, in well-weathered soils, which do not contain $CaCO_3$, a specific effect of exchangeable Mg is evident on both HC losses and clay dispersion when the Na/Mg soils are leached with distilled water (simulating rainwater). Mg-saturated soil clays are chemically more stable and do not release electrolytes into the solutions (Kreit et al., 1982); they thus disperse more easily when leached with distilled water. The presence of exchangeable Mg enhances the dissolution of $CaCO_3$ in calcareous soils, and the electrolytes prevent the dispersion of clay and HC losses in Na/Mg calcareous soils.

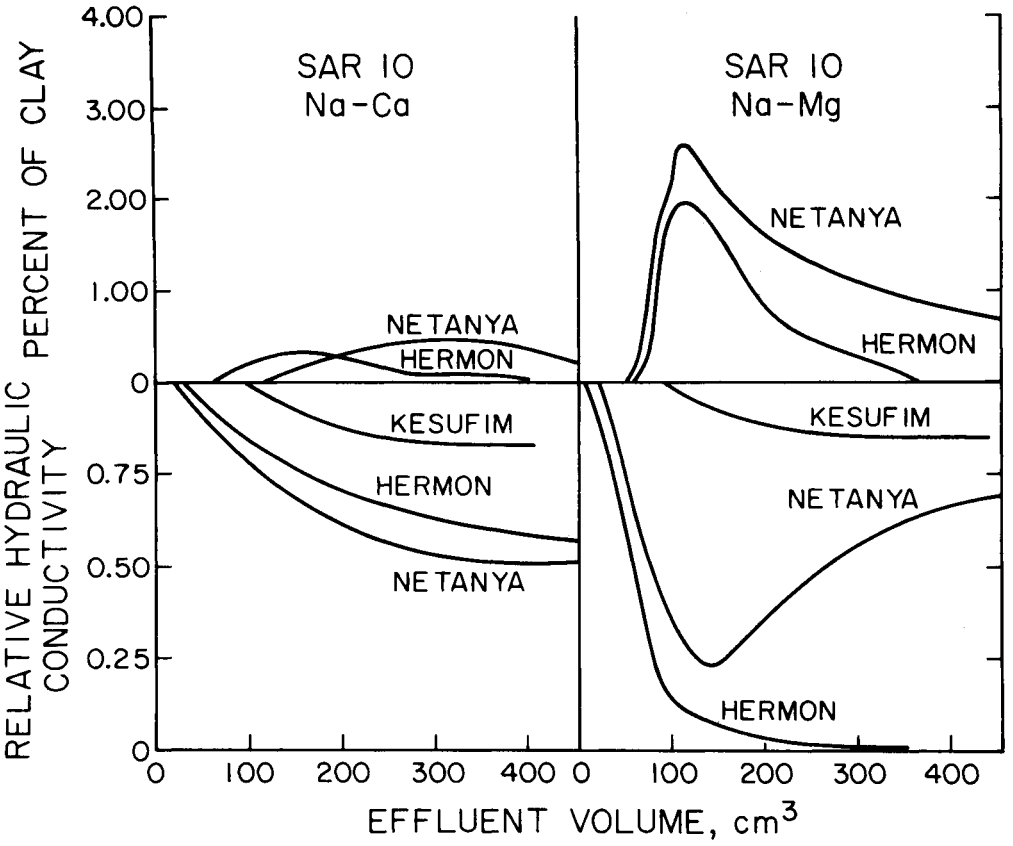

Fig. 21. Relative HC and concentration of clay in the effluent of six Israel soils equilibrated with 0.01 N solutions of SAR 10 and leached with distilled water. Where no data for clay concentration are given, no clay appeared in the effluent (Alperovitch et al., 1981).

Response of sodic soils to gypsum and $CaCl_2$ application. Gypsum is the most commonly used amendment for sodic soil reclamation, primarily because of its low cost (Oster, 1982). The amount of gypsum required to reclaim a sodic soil depends on the amount of exchangeable Na in the soil profile (US Salinity Laboratory Staff, 1954). Gypsum added to a sodic soil can initiate permeability increases due to both electrolyte concentration and cation exchange effects (Loveday, 1976). The relative significance of the two effects is of interest for several reasons. If the electrolyte effect is sufficiently great to prevent dispersion and swelling of soil clays, surface application of gypsum may be worthwhile. In this case the amount required depends on the amount of high-quality water applied and the rate of gypsum dissolution (Oster, 1982)—it is somewhat in-

dependent of the amount of exchangeable Na in the soil profile. Conversely, in soils where the electrolyte concentration effect is insignificant and the main effect results from cation exchange, the amount of gypsum required depends on the quantity of exchangeable Na in a selected depth of soil. The cation exchange process has formed the basis of several "gypsum requirement" tests (US Salinity Laboratory Staff, 1954; Oster and Frenkel, 1980; Keren and O'Connor, 1982; Oster, 1982). The relative significance of the electrolyte effect was estimated recently (Shainberg et al. 1982) by comparing the effects of equivalent amounts of gypsum and $CaCl_2$ applied at the soil surface on the hydraulic conductivity of three Israel soils leached with distilled water. The exchange reclamation was similar with both amendments, but there was a difference in the electrolyte effect. The slow dissolution and the relatively long-term electrolyte effect of gypsum were very important in maintaining high hydraulic conductivity in a chemically stable soil which did not release salt into the soil solution. For a corresponding $CaCl_2$ treatment, complete sealing of the soil took place. Conversely, the efficiency of the two amendments was similar for a calcareous soil. Even without gypsum, this soil released enough electrolytes into the soil solution to prevent clay dispersion and hydraulic conductivity losses. The ability of gypsum to maintain a moderate concentration of electrolytes in the percolating water is very important in preventing crust formation and soil sealing under rainfall conditions.

Effect of lime on the response of soils to sodic conditions. The presence of fine $CaCO_3$ particles in soils is known to improve the physical conditions of sodic soils (US Salinity Laboratory Staff, 1954). Because of low $CaCO_3$ solubility in soils having pH greater than 7.5, lime is not effective in the exchange mechanism by which exchangeable sodium is replaced by calcium, and thus sodic soils containing $CaCO_3$ are common in the semiarid and arid regions of the world. To explain the effect of $CaCO_3$, the US Salinity Laboratory Staff (1954) and Rimmer and Greenland (1976) suggested that lime in soils acts as a cementing agent which stabilizes soil aggregates and prevents clay dispersion. Another mechanism which explains the beneficial effect of Ca CO_3 is its potential for dissolving and maintaining the concentration of the soil solution at concentration levels above the flocculation value of the soil clays, thus preventing their dispersion. According to the dissolution mechanism, exchange reclamation is still negligible because of the low concentration of Ca ions in the soil solution, and clay dispersion is prevented mainly because of the electrolyte concentration effect (Shainberg et al., 1981*a*).

The mechanism by which $CaCO_3$ stabilizes soil structure is important in studies on infiltration rate and crust formation in calcareous and noncalcareous soils under rainfall conditions (e.g., Agassi et al., 1981). If $CaCO_3$ acts as a cementing material, calcareous soils will not be as sensitive to crust formation as will noncalcareous soils. However, if the dissolution mechanism is dominant, the concentration of electrolytes at the soil's surface exposed to rain will be insufficient, and both soils will be sensitive to crust formation.

The rate of $CaCO_3$ dissolution in preventing HC losses of sodic soils was determined by Shainberg and Gal (1982) by mixing lime-free soils with low percentages of powdered lime (0.5 and 2.0 percent). Although the HC of the lime-free soils dropped sharply when 0.01 N solutions of SAR 20 were displaced with distilled water, mixing the soils with powdered lime prevented both HC losses and clay dispersion (Fig. 22). The increase in electrolyte concentration in the soil solution due to $CaCO_3$ dissolution was suggested as the mechanism responsible for the beneficial effect of lime. The dissolution mechanism explains why both calcareous and noncalcareous soils are so sensitive to crust formation when soils with low ESP levels are exposed to rainfall (Agassi et al., 1981; Kazman et al., 1983) (see section beginning page 33).

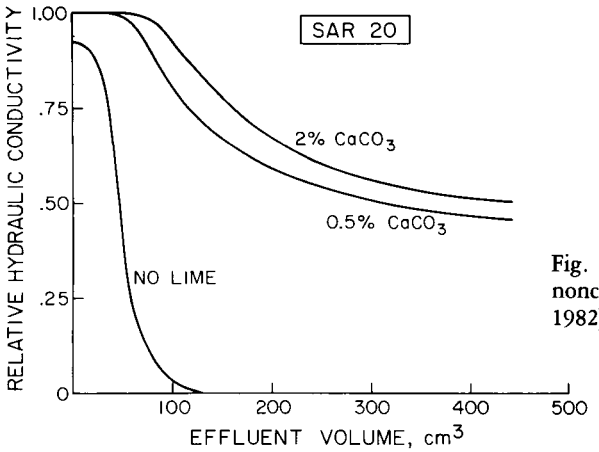

Fig. 22. Effect of lime on the HC of noncalcareous soil (Shainberg and Gal, 1982).

Summary

Permeability of a soil to water depends both on its ESP and the salt concentration of the percolating solution. Soil permeability can be maintained, even at high ESP values, if the solution concentration is above a critical (threshold) level. Conversely, at low ESP values where little soil swelling is expected, clay dispersion and a marked reduction in soil permeability occur in very dilute salt solutions. When the salt concentration is 3 meq/l, decreases in HC and clay dispersion occur in soils of ESP values >12. In distilled water, clay dispersion and HC reduction occur at ESP values as low as 2.

For soils of intermediate sodicity (ESP 10 to 20), the flocculation value of the soil clays is below 5 meq/l. Thus, when the salt concentration in the percolating solution exceeds approximately 5 meq/l, the main mechanism for HC reduction is clay swelling. At this concentration range, susceptibility of soils to sodic conditions is enhanced by higher clay content and the presence of expansible minerals (smectites).

When the salt concentration in the percolating water is below the flocculation value, clay dispersion, movement, and lodgement in the conducting pores comprise the main mechanism responsible for HC reduction. The most important soil property determining susceptibility to sodic conditions when the soil is leached with very dilute salt solutions (or rainwater) is the soil's potential for releasing electrolytes through dissolution of primary minerals and/or $CaCO_3$. Soils which release salt at a rate sufficient to maintain the concentration of soil solution above the flocculation value of the clay will not disperse, and will not be sensitive to low ESP values.

Differences among soils in their abilities to release salt and to undergo dispersion are not so important in affecting water infiltration and crust formation under rainfall conditions (or very good quality water application). Soils having moderate ESP levels may maintain reasonable physical properties through most of the profile and yet be susceptible to dispersion near the surface. This is because the soil solution electrolyte concentration may be insufficient there to maintain physical structure. This subject is discussed in detail in the next section.

EFFECT OF SODICITY AND SALINITY ON THE HYDRAULIC CONDUCTIVITY OF SOIL SURFACE (CRUST FORMATION) AS DETERMINED BY THE INFILTRATION RATE (IR)

Definition and experimental background

Infiltration—or the downward entry of water into soil—is one of the more important processes in the soil phase of the hydrological cycle. The rate of this process relative to the rate of water supply determines how much water will enter the root zone, and how much, if any, will run off. Where the infiltration rate (IR) is restricted, plants may be denied sufficient moisture while the amount of erosion increases.

The IR is defined as the volume flux of water flowing into the profile per unit surface area of soil, and this flux has the dimension of velocity. In general, the soil infiltration capacity is initially high, particularly when the soil is initially dry, but it tends to decrease monotonically until it asymptotically approaches a constant rate—the final infiltration rate or steady-state IR. If water delivery rate to the surface is smaller than the soil infiltration capacity, water infiltrates as fast as it is delivered and the supply rate determines the IR. When delivery rate exceeds soil infiltration capacity it is the latter which determines the actual IR, and the process becomes surface-controlled.

In soils having stable surface structures, decreases in infiltration capacity result from the inevitable decrease in the matric suction gradient which occurs as infiltration proceeds (Baver et al., 1972; Hillel, 1980). In the initial wetting process, the surface is saturated but the soil below is dry and the matric suction gradient is very steep. As the wetted zone deepens, this gradient is gradually reduced. Decreases in infiltration capacity of a soil from an initially high rate can also result from gradual deterioration of the soil's structure and the formation of a surface crust.

Numerous formulations have been proposed in attempts to express IR as a function of time or of the total quantity of water infiltrated into the soil (Baver et al., 1972; Hillel, 1980). One of the first and the most realistic equations is that of Horton (Baver et al., 1972; Hillel, 1980)

$$i = i_c + (i_o - i_c)e^{-kt} \tag{18}$$

where i is the infiltration rate; i_c is the asymptotic steady infiltration rate reached when the time t and the cumulative volume of water infiltrated become large; i_o is the initial infiltration capacity of the soil; and k is a constant which determines how quickly i will decrease from i_o to i_c. As will be shown below, the Horton equation describes best the IR of rain into crusted soils. Soil crusting is the general case in arid and semiarid regions where irrigation is practiced and salinity and sodicity problems arise.

Several studies have been conducted on the effect of rainfall on the structure and hydraulic properties of soil crusts (McIntyre, 1958; Evans and Buol, 1968; Chen et al., 1980; Keren and Shainberg, 1981; Morin et al., 1981). McIntyre (1958) found the crust consists of two distinct parts: an upper skin seal attributed to compaction by raindrop impact, and a "washed-in" region of decreased porosity attributed to the accumulation of particles. McIntyre (1958) measured thicknesses of 0.1 and 2 mm for the skin seal and the washed-in zone, respectively. The washed-in layer was formed only in easily dispersed soils. Permeability of the deeper layer was about 800 times that of the washed-in

layer, and about 2,000 times that of the skin seal. Chen et al. (1980) examined scanning electron micrographs of crusts of loessial soil and also found a thin skin seal about 0.1 mm in thickness that had formed at the uppermost layer of the soil. They did not, however, find an accumulation of fine particles in the 0.1-to-2.8-mm zone (the washed-in layer) as observed by McIntyre (1958). Sealing efficiency of the crust is caused by suction forces which hold the clay particles together in a continuous dense skin (Morin et al., 1981). Suction forces at the soil/crust interface are created as a result of the large differences in HC between the crust and the underlying soil. The suction mechanism accounts for the stability of the crust HC and the similarity in HC of the crusts of soils varying greatly in their texture and mineralogy (Morin et al., 1981).

It is thus evident that crust formation is associated with clay dispersion and movement in the soil. Soil surfaces are especially susceptible to the chemistry (electrolyte concentration and cationic composition) of the applied water because of the mechanical action of the falling drops and the relative freedom of particle movement at the soil surface. Indeed, Oster and Schroer (1979), studying the IR of undisturbed loam soil columns, found that the effects of the applied water chemistry were far greater than expected. When the ESP of the surface soil layer was 8, the IR decreased from 15 to 1 millimeter per hour (mm/h) as the concentration of the irrigation water decreased from 28 to 8 meq/l. Comparable reductions in saturated hydraulic conductivity of saturated soils having an ESP of 10 occur only when the concentration of the percolating solution decreases below 2 to 3 meq/l (Shainberg et al., 1981*a*).

A rain simulator such as that developed by Morin et al. (1967) is a tool for studying the effects of soil sodicity and irrigation water salinity on the infiltration rate and crust formation of soils. Soil samples are packed in 30 × 50 cm trays which are then placed in the simulator at a predetermined slope and subjected to simulated rain. Intensity of the "rain" and impact energy of the drops may be controlled—the chemistry of the water may also be changed according to experimental plan. When rain water is to be simulated, water of electrical conductivity 0.1 dS/m is applied. Irrigation water of any desired quality can also be placed in the rain simulator. Volume of runoff water and that of percolating water are measured, and the IR is calculated.

Effect of soil sodicity on the IR of soils

The effect of soil sodicity (ESP) on the IR and crust formation of four soils varying in texture, clay mineralogy, and $CaCO_3$ content was studied by Kazman et al. (1982) using distilled water (EC .05 to .1 dS/m) in a rain simulator. The IR of the four soils was very sensitive to low levels of ESP. The IRs of two soils, the same soils on which the HC studies were made (Fig. 13), are presented in Figures 23 and 24. The final IR of the Netanya sandy loam of ESP 1.0 was maintained at 7.5 mm/h, whereas the final IRs of the soils having ESP values of 2.2, 4.6, and 11.6 dropped to 2.3, 0.7, and 0.6 mm/h, respectively. An ESP value of 2.2 was enough to cause a drop of 70 percent in the final IR. Similar effects of ESP were observed on the Nahal-Oz soil. The amount of rain required to approach the final infiltration rate was also affected by ESP (Figs. 23 and 24).

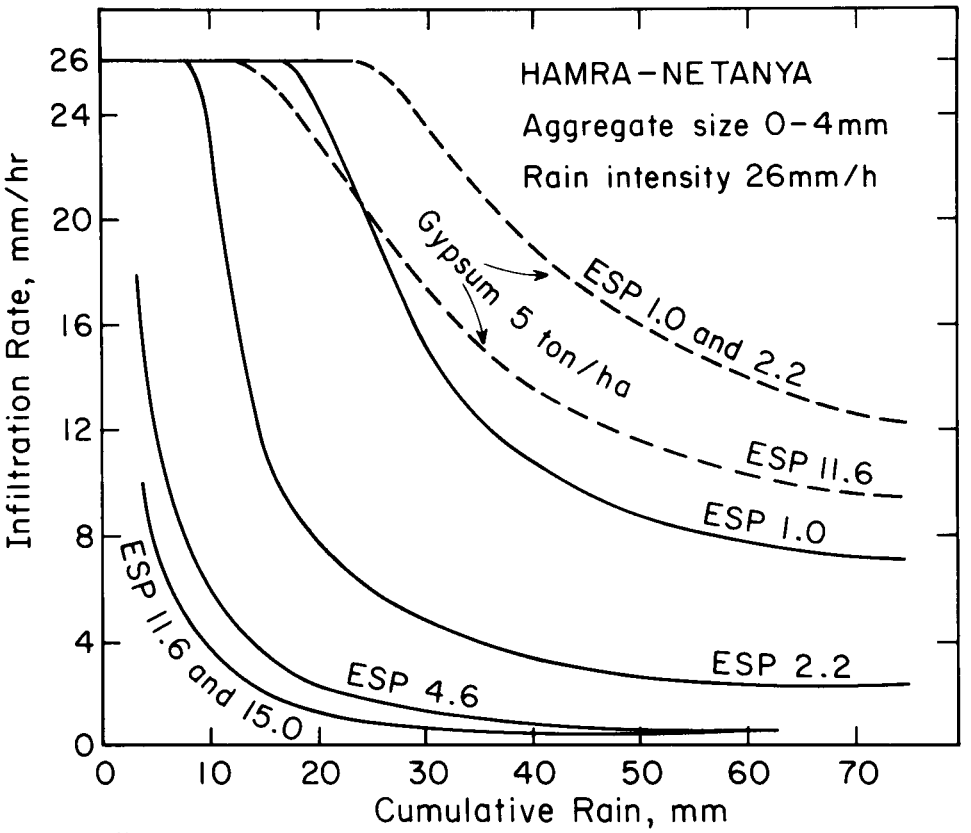

Fig. 23. Effect of soil ESP (and phosphogypsum) on the infiltration rate of the Hamra (Netanya) soil (Kazman et al., 1982).

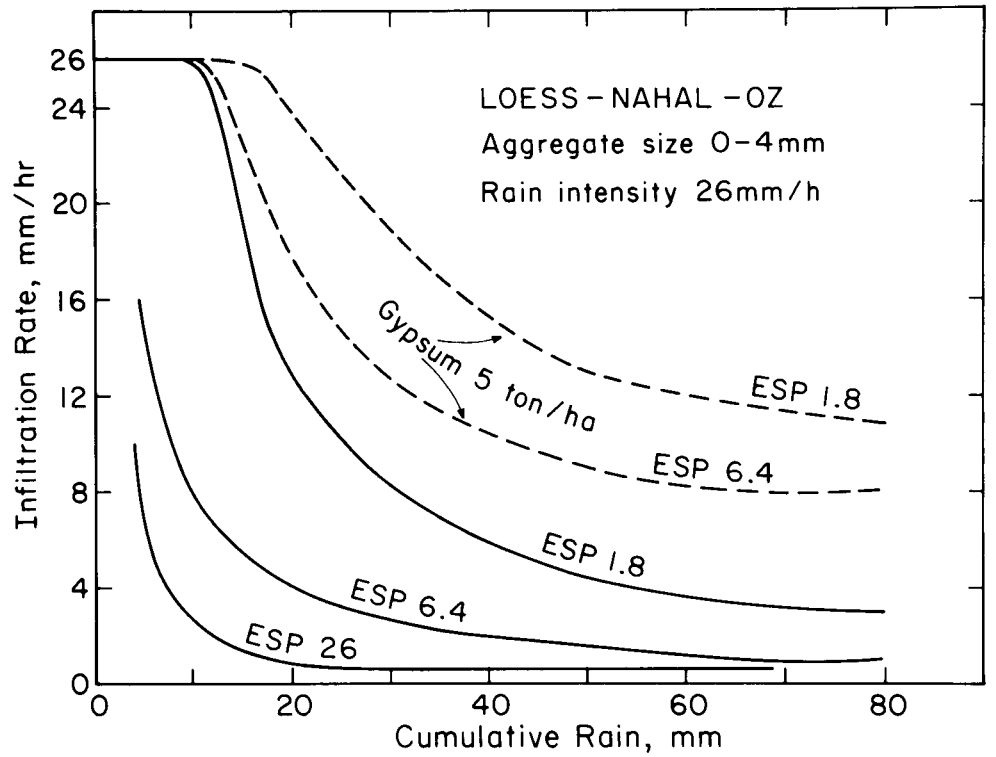

Fig. 24. Effect of soil ESP (and phosphogypsum) on the infiltration rate of the loess soil (Kazman et al., 1982).

Comparing the IR data (Figs. 23 and 24) with the HC data (Fig. 13), it is evident that the IR is more sensitive to a low ESP than is the HC. Furthermore, the IRs of the calcareous and noncalcareous soils are equally sensitive to low levels of exchangeable sodium, whereas the HC of the calcareous soil is not sensitive to this condition. These two phenomena are explained by the same mechanism. The soil solution electrolyte concentration determines the response of soils to low ESP values. When a sodic calcareous soil is leached with distilled water, $CaCO_3$ dissolution maintains electrolyte concentration in the soil solution at values close to 3 meq/l, which is sufficient to prevent clay dispersion with subsequent decreased HC. However, the soil solution concentration at the soil surface is determined solely by the concentration of the applied water (nearly distilled water for rainwater), thus aggregate dispersion takes place at the surface at low levels of ESP even in calcareous soils. The higher susceptibility of the surface of soils to low levels of ESP is supplemented by two other factors: mechanical impact of the raindrops, which enhances dispersion; and absence of the soils matrix (sand particles), which slows clay dispersion and clay movement (Oster and Schroer, 1979).

The raindrop impact energy was the same in all the experiments, so differences in IR curves for the various soil samples were due to the potential for soil dispersion caused by sodicity. Apparently crust formation in soils exposed to rain was due to two types of mechanisms: a physical dispersion caused by impact action of the raindrops, and a chemical dispersion which depended on the ESP of the soil and the electrolyte concentration of applied water. The physical dispersion mechanism dominated in soils having very low sodium in the exchange complex. When the physical dispersion mechanism dominated, the steady IRs of Netanya and Nahal-Oz soils dropped to values around 7 to 8 mm/h.

When the impact energy of the raindrops was prevented (by mulching) in soils with ESP below 5 to 6, the rain application rate was smaller than the infiltration capacity of the soil and the rain intensity determined the IR curve (an IR line parallel to the x-axis at an IR of 26 mm/h). Thus the drop in IR in soils with ESPs below 1 was mainly due to mechanical action of the raindrops. The degree of the chemical dispersion depended on the ESP of the soil—increasing ESP resulted in increasing chemical dispersion.

The mechanical and chemical mechanisms are complementary. The kinetics of the chemical dispersion depends markedly on the intensity of mechanical mixing. Thus, the mechanical impact of the raindrops has two effects: breakdown of soil aggregates followed by compaction of the soil surface and production of thin skin seals (McIntyre, 1958); and stirring of soil particles which enhances their rate of chemical dispersion. Without the stirring process, the rate of chemical dispersion might be slow. In HC measurements, the stirring process is kept at a minimal value and a higher percentage of exchangeable sodium is needed to cause dispersion.

Chemical dispersion may be prevented by spreading phosphogypsum (or another readily available electrolyte source) at the surface of sodic soils (Keren and Shainberg, 1981) (Figs. 23 and 24). Phosphogypsum is a by-product of the phosphate fertilizer industry and is available in large quantities—when it was applied to both soils at a rate of 5 tons per hectare (ton/ha) the IR drop and the steady IR were both affected. The steady IRs of Netanya soils having ESP values of 1.0 and 2.2 were maintained at 12.2 mm/h when treated with phosphogypsum, as compared with 7.5 and 2.2 mm/h in the control. It is evident that phosphogypsum was effective even in soils of ESP 1.0, thus indicating that some chemical dispersion took place even at very low ESP values, and possibly even in Ca-soils.

Keren and Shainberg (1981) compared mined gypsum to phosphogypsum and found

that the efficiency of phosphogypsum in maintaining high IR was greater than that of mined gypsum. They also found that the rate of dissolution of phosphogypsum was much higher than that of mined gypsum. The dissolution rate was important because of the short contact time between rainwater and gypsum at the soil surface. Electrolyte concentration in the surface soil solution treated with mined gypsum was apparently not sufficient to prevent dispersion and crust formation because of slow dissolution. The beneficial effects of phosphogypsum in maintaining IR appear to be caused by rapid dissolution resulting in a relatively high electrolyte concentration rather than a cation-exchange phenomenon.

Although most of the data on the effect of soil sodicity on the IRs of soils were obtained under rain conditions, it is believed that similar results will be obtained when low-salinity irrigation water is applied—the similarity between rain and sprinkler irrigation is obvious. The spectrum of the sprinkler drops' size and their terminal velocity depends on the type of sprinkler used and is different from that of the raindrops. Though it is expected that the impact energy of the drops will have an effect on the IR and on crust formation in sodic soils, the magnitude of this effect should not alter basic conclusions previously stated.

Similarly, with surface irrigation particle disturbance is introduced at the soil surface by the flowing water and by release of air bubbles formed during water infiltration. The stirring energy in surface irrigation is probably not as large as in a rainstorm or in sprinkler irrigation, but is enough to enhance clay dispersion. The other factors that were described in connection with rain infiltraton (salt concentration at the soil surface determined by the irrigation water quality and not by the soil potential for releasing salts, and the absence of soil matrix) are also pertinent to surface irrigation. Thus it is expected that in surface irrigation the IR will also be more sensitive to low levels of ESP than will the HC of the soil profile. Oster and Schroer (1979) presented data on the infiltration rate of a soil exposed to surface irrigation only, and they also showed the high susceptibility of soils having low SAR values (2 to 5) when irrigated with waters of very low electrolyte concentrations.

Effect of electrolyte concentration on the IR of sodic soils

Agassi et al. (1981) studied the effect of electrolyte concentration and soil sodicity on the infiltration rate of two loamy soils in Israel (Netanya and Nahal-Oz soils), using a rainfall simulator. Figures 25 and 26 show their results. They found that the IR was more sensitive than the HC to the electrolyte concentration of applied water. For example, the HC of Netanya soil, when in equilibrium with SAR 10 solutions, was not affected by the electrolyte concentration in the percolating solutions as long as the electrolyte concentration was above 5 meq/l ($\simeq$0.5 dS/m) (Fig. 13). Similar observations were made with Nahal-Oz soil from Israel (Fig. 13) and Fallbrook soil from California (Fig. 16). In all these soils, concentrations above 5 meq/l were more than enough to prevent the adverse effect of sodium in the ESP range of 0 to 15.

However, the IRs of these soils in the above ESP range were affected by the electrolyte concentration up to 50 meq/l. For example, the final IR of sandy loam Netanya-Hamra with ESP 13.6 increased from 1.2 to 2.2, 5.3, and 7.5 mm/h as the electrical conductivity of the applied water increased from distilled water to 0.5, 2.3, and 5.6 dS/m, respectively (Fig. 25). Similar results were obtained by Oster and Schroer (1979) on Heimdal soil from North Dakota. They also found that cation concentration greatly

affected infiltration rates even at low SAR levels of 2 to 5. Their data showed that for SAR values between 2 and 4.6 the final IR increased from 2 to 28 mm/h as the cation concentration of the applied water increased from 5 to 28 meq/l.

Figures 25 and 26 show that salt concentration had an effect on the final IR and on the rate at which the IR of the soil dropped from the initial to the final IR value. The rate of the IR decrease accelerates with reduced electrolyte concentration. Similarly at the same electrolyte concentration, the higher the sodicity of the soil is, the greater is the rate at which the IR drops to its final value, and the lower is this final value. The rate of IR decrease is important in agricultural practice—it suggests that when low-salinity water is used, the amount of runoff is high, and there is little advantage in cultivating the crusted soil because the rate of crust formation is high. It was concluded (Agassi et al., 1981) that crust formation is a kinetic process which depends on the electrolyte concentration in the applied water, the ESP of the soil, and the disturbance caused by applied water at the soil surface. The impact of the raindrops has two effects: breaking down the aggregate, and enhancing the rate of the chemical dispersion by stirring the soil surface.

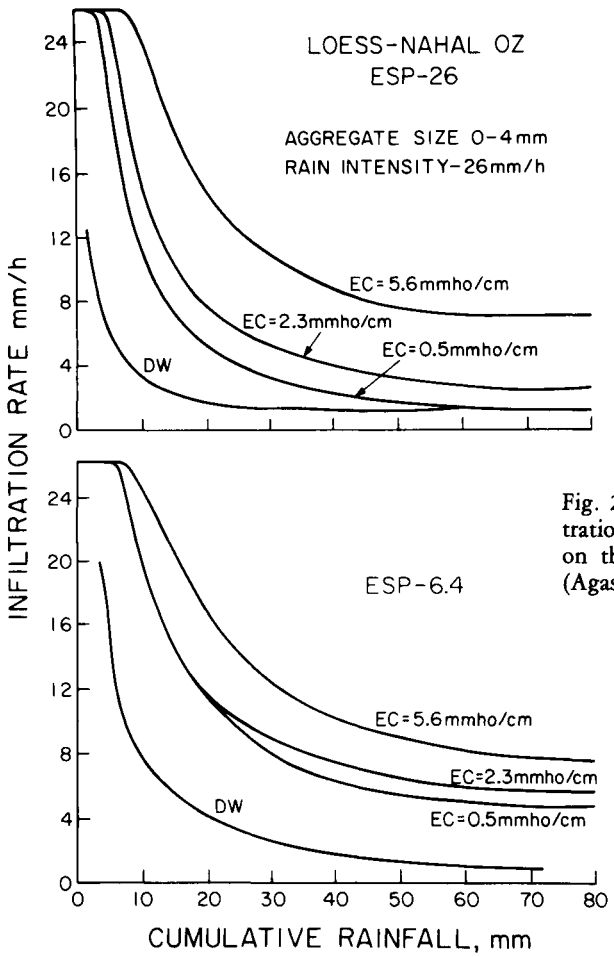

Fig. 25. Effect of electrolyte concentration in rain-simulation experiments on the infiltration rate of Hamra soil (Agassi et al., 1981).

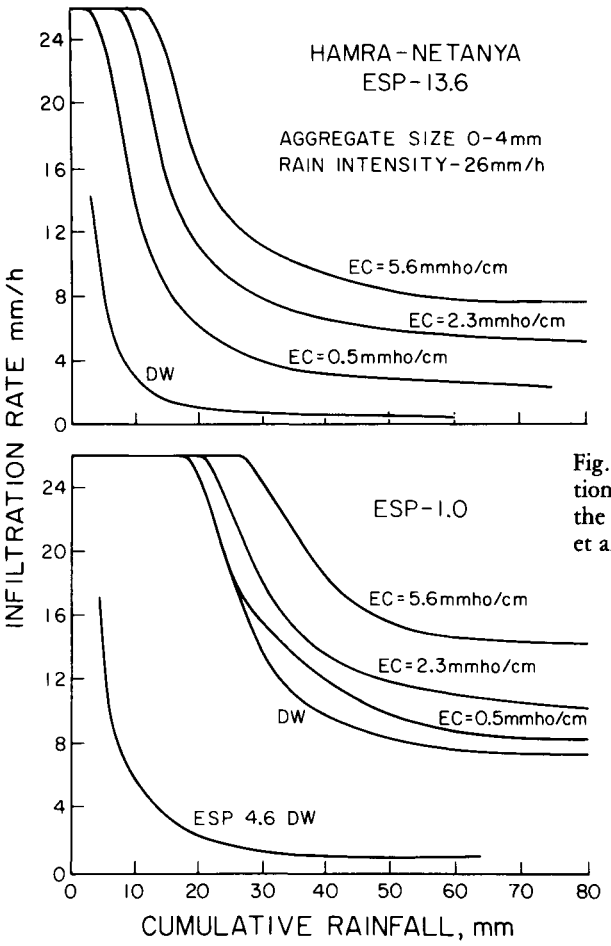

Fig. 26. Effect of electrolyte concentration in rain-simulation experiments on the infiltration rate of loess soil (Agassi et al., 1981).

The rate of the chemical dispersion is worth re-emphasizing. When solutions of high concentration are applied, the impact energy of the drops is the main force causing breakdown of the soil aggregates and a compacted layer of low permeability is produced at the soil surface. The rate at which this compacted layer is formed is relatively slow, and the final IR values are maintained at 8 to 15 mm/h. When low-salinity water is applied to the soil, even with low levels of exchangeable sodium, chemical dispersion of the soil clays also occurs. The dispersed clay particles can be washed into the soil with the infiltrating water, and the pores immediately beneath the surface become clogged. The final IR values of the two soils (Figs. 25 and 26) when rained upon with distilled water dropped to a value around 1 mm/h, independent of the soil ESP (provided the ESP value was above 4 to 5). Increasing the ESP of the soil had an effect on the rate at which the final IR value was reached, but not on the final IR value. The IR characteristics of the calcareous Nahal-Oz and noncalcareous Netanya soils responded in a similar way to low sodicity and low electrolyte concentration. This is in spite of two factors which should stabilize the structure of the Netanya soil: the Netanya soil contains a high percentage of sesquioxides which act as cementing agents, stabilizing the soil structure; and the loessial soil contains a high level of silt (37.8 percent compared with only 5.0 percent in the Netanya soil), and it is generally accepted (Cary and Evans, 1974) that a high content of

silt weakens the soil structure. Apparently, the chemical instability of the Nahal-Oz soil is more than enough to counter these factors, and the IR characteristics of the two soils are similar.

Effect of intermittent application of rain and saline water on the IR of soils

Intermittent application of rain and saline irrigation water predominate in many semiarid regions of the world. It was shown in the previous section that when saline water was applied to soils of low sodicity, a final IR of 7 to 8 mm/h was maintained, whereas when distilled water was applied crust was formed and the IR dropped to 1 to 2 mm/h. Hardy et al. (1983) studied the stability and the reversibility of the crust formed on the Netanya and Nahal-Oz soils using a rainfall simulator. They studied the effect of subsequent saline water (SW) irrigation on the stability of crusts formed under distilled water (DW) rain. Figure 27 presents their results.

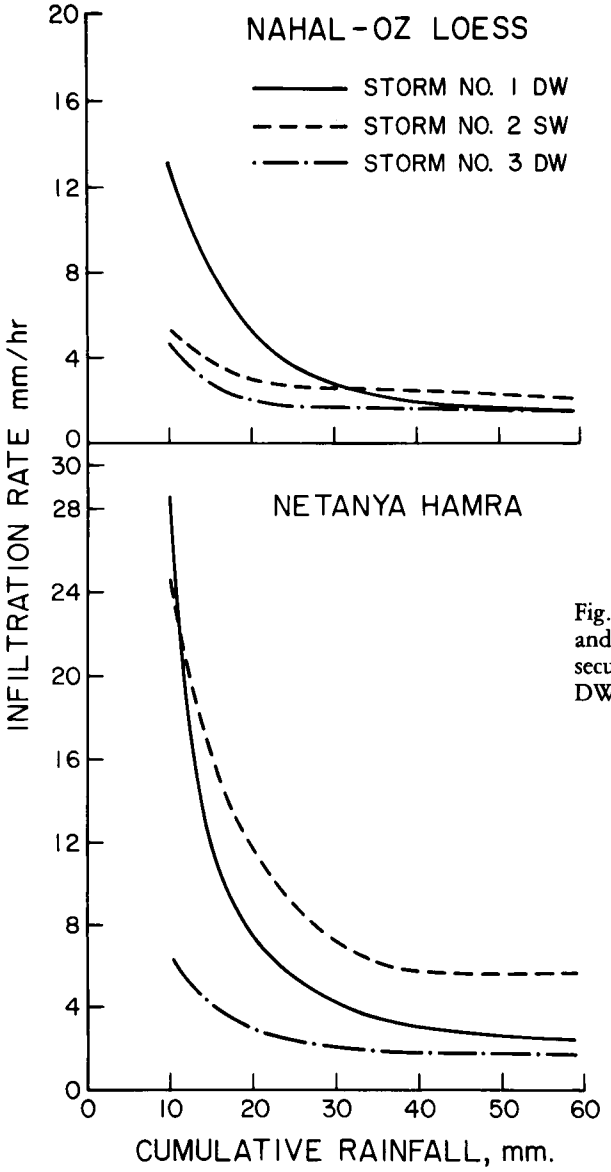

Fig. 27. Infiltration rates of Nahal-Oz and Netanya soils subjected to three consecutive storms of sequence DW-SW-DW (Hardy et al., 1983).

When two consecutive storms of DW were applied to Netanya and Nahal-Oz soils, curves similar to those designated as storms number 1 and 3 in Figure 27 were obtained. The final IR in the second (or subsequent) storm was obtained more rapidly (compared with the first storm) because the surface was already crusted (Morin et al., 1981). Application of SW in the second storm to the already-crusted loessial soil (Fig. 27) resulted in a small increase in final IR, from 1.8 to 2.5 mm/h. This final value was reached after the application of 40 mm of rain. A third storm, of DW, resulted in an infiltration curve essentially identical to that obtained for a second consecutive storm of DW. The crust formed in the first DW storm was stable, only a small increase in IR in the SW storm was observed, and when a third DW storm was applied the IR curve was that typical of a second consecutive DW storm—the intermittent SW storm had no effect on the properties of the crust.

Application of SW in the second storm to the crusted sandy loam Netanya soil caused a much larger increase in final IR, to 5.6 mm/h. Application of a third storm, of DW, resulted in a much lower IR at any given rain depth. The two soils responded very differently to a storm sequence of alternating water quality. The crust in Netanya soil behaved in a reversible way and the final IR of consecutive storms was almost independent of the penultimate storm, but the loessial crust behaved in an irreversible way and the effect of the crust formed in the first DW storm was maintained in the second (SW) storm. This pronounced difference between the soils was explained by observing the pictures of crust formation during rainstorms (Figs. 28 and 29). Crust formation during the initial stages of the first storm is illustrated in the left-hand columns of Figures 28 and 29. The soil surfaces of both Nahal-Oz and Netanya changed from raw tilth to smooth crusted surface during the initial 12 mm of rainfall. Initially, raindrop impact caused a breakdown of the soil structure and the formation of small craters or pits. As rain proceeded, soil erosion took place and the surface became progressively less pitted until a smooth crust formed. This phenomenon is well-documented (McIntyre, 1958). Pitting was more pronounced in the sandy loam than in the loess, and the loess began to take on a smooth appearance after 4 mm of rain compared with the 7 mm of rain needed to reach this stage in the sandy loam soil.

The right-hand columns of Figures 28 and 29 show the soil surface during the initial stages of the second storm. The crust that formed on the loessial soil during the first storm did not alter in appearance during the second storm (Fig. 28). The surface of the sandy loam behaved very differently during the second storm (Fig. 29); rainfall impact upon its crust caused pitting and a rearrangement of soil particles. Maximum pitting was obtained at 5 mm rainfall, and the soil erosion that took place with further rain caused the reformation of a smooth soil crust.

In loessial soil the crust formed during the first storm maintained a continuous crust structure during subsequent storms, while in sandy loam the crust was reworked by raindrop impact. In the sandy loam the crust formed during any particular storm thus has little relationship to the crusts formed in previous storms, whereas in loessial soil the crust, which is not destroyed by subsequent storms, determines the IR curves of consecutive storms. The response of the crusts of the two soils to the beating action of the raindrops explains the IR curves in Figure 27. As no major rearrangement of the loessial crust occurs during the second (SW) storm, the washed-in zone remains complete and *in situ*. Flocculation of the clay particles by the SW may occur but movement may not. Possibilities for increased hydraulic conductivity of this limiting zone are therefore small, and only a small increase in IR is found. Rainfall impact on sandy loam soil in the second storm causes destruction of the crust formed during the preceding DW storm, while the

SW causes flocculation of the clay particles. Therefore, the new crust will result mainly from the physical action of the raindrops. Clay movement and formation of the washed-in zone will be limited.

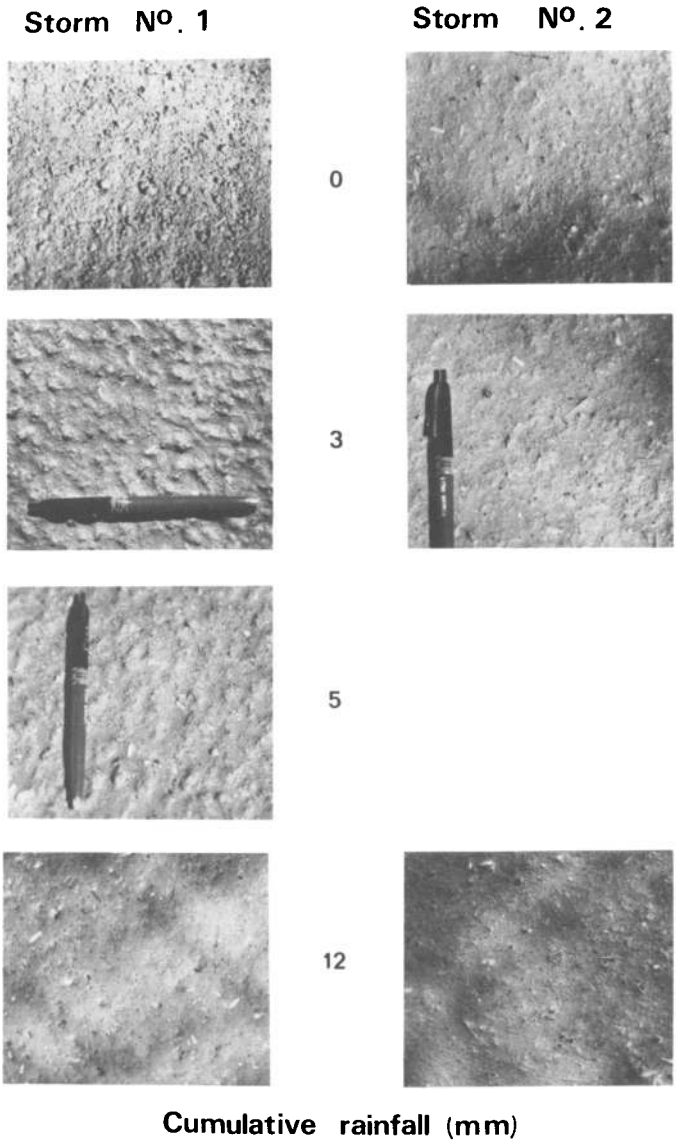

Fig. 28. Surface appearance of Nahal-Oz soil during the initial stages of the first and second storms—the numbers 0, 3, 5, and 12 refer to cumulative rainfall in mm (Hardy et al., 1983).

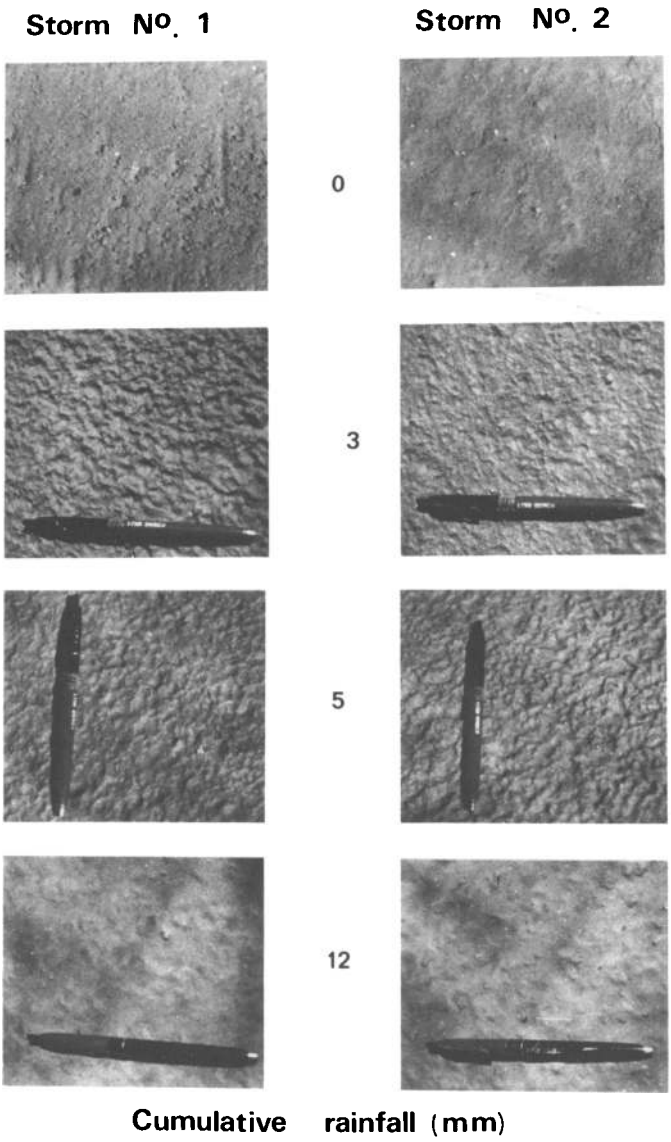

Fig. 29. Surface appearance of Netanya soil during the initial stages of the first and second storms—the numbers 0, 3, 5, and 12 refer to cumulative rainfall in mm (Hardy et al., 1983).

The underlying difference between the soils is one of texture. The impact of the raindrops mechanically disrupts the crusted surface of the sandy loam since its coarse texture results in only a relatively small number of contact points between soil particles, and a large average pore size. Conversely, soils having high silt content have sufficient particle-to-particle contact points to form strong bonds when the soil dries, thus forming a strong crust (Cary and Evans, 1974). The hydraulic conductivity of the crust is lower

than that of the underlying soil, so water flow in the soil under the crust is unsaturated (Baver et al., 1972), and this results in a state of suction from the soil surface downward. This suction catalyzes the processes of clay leaching into the soil surface and crust formation (Morin et al., 1981). Suction pressure increases with decreasing particle size, so suction is greater in Nahal-Oz than in Netanya. The combined effects of reduced pore size, higher attainable negative water pressure, and increased number of contact points all contribute to enhance bond formation in the crusted loessial soil. Rainfall impact is therefore unable to cause the destruction of the crust in the loessial soil as it does in the sandy loam.

Summary

Infiltration of water into soil is one of the most important processes in the hydrological cycle and is very important in agriculture. This is also the process most sensitive to the ESP of the soil and the electrolyte concentration in the applied water. The mechanical impact of raindrops (or the surface stirring and mixing introduced by applied irrigation water) and the relative freedom of particle movement at the soil surface may account for the greater sensitivity of the infiltration rate as compared to hydraulic conductivity. Experimental observations suggest that crust formation is due to two complementary mechanisms: a physical breakdown of soil aggregates followed by thin-layer compaction caused by impact action of the drops, and a chemical dispersion dependent on the soil ESP and the electrolyte concentration in the applied water. The kinetics of crust formation is also dependent on soil ESP and salt concentration. Stirring of the soil surface by drop impact and irrigation water flow enhances the rate of clay dispersion and crust formation.

The strength of the crust, as determined by the cohesion forces between the particles forming the crust, also determines its stability under the reworking and pitting effects of water drops. The crust formed on a silty loam soil during a rain storm is a stable and permenanent crust impermeable to saline water. The crust on a sandy loam soil, however, is not stable and is permeable to saline water.

The importance of the infiltration process and its sensitivity to sodicity and electrolyte concentration suggest that in evaluating water for irrigation the effect of water quality on the IR should be the most important consideration.

IRRIGATION WATER QUALITY CRITERIA AS RELATED TO SOIL PROPERTIES

The quality of irrigation water has been discussed in many papers and reviews (e.g., Ayers and Tanji, 1981; Ayers and Westcot, 1976; Bresler et al., 1982; Rhoades, 1972; US Salinity Laboratory Staff, 1954). The suitability of such water has been evaluated on the bases of criteria which determine its potential for causing soil conditions injurious to crop growth. Important water quality criteria are discussed below.

Salinity. Salinity generally refers to total salt concentration and is most commonly measured and reported in terms of the electrical conductivity (EC). A salinity problem occurs if the concentration of salt in the irrigation water is high enough to affect crop

yield. High salt concentration in the soil solution does not damage soil properties and is not discussed in this review. Very low salt concentration, however, can cause dispersion and greatly decrease infiltration rate and hydraulic conductivity, and in this review is considered a water criterion that might affect crop growth.

Toxicity. This problem occurs when certain constituents in the water are taken up by the crop and accumulate in amounts that result in a reduced yield. These constituents are traditionally grouped as follows: (1) toxic chloride and/or boron concentration; (2) trace elements (a guideline on maximum permissible concentrations is given in Ayers and Westcot, 1976); (3) nitrate (the hazard of high nitrate levels is related mainly to potential human health problem); and (4) pesticides. Toxicity will not be discussed further here because it does not affect the soil's physical properties.

Sodicity. In evaluating the suitability of irrigation water, an important consideration is the extent to which ESP will increase in the soil due to adsorption of sodium from irrigation water via cation exchange. The best known index to irrigation water sodium hazard is the sodium adsorption ratio (SAR). The soil ESP increase is almost linear with the increase in the SAR of the waters (US Salinity Laboratory Staff, 1954).

The rate at which the ESP in soil increases as a result of irrigating with waters of a given SAR depends on the solution concentration. More sodium should be applied to soil per given depth of irrigation water when waters of higher salt concentration are used—see the diagram for classification of irrigation water prepared by the US Salinity Laboratory Staff (US Salinity Laboratory Staff, 1954, Fig. 25).

The quantification of the relation between SAR of the irrigation water (SAR_{iw}), soil ESP, and SAR of the soil solution (SAR_{sw}) is difficult for the following reasons. (1) Concentration of applied soil water is increased by evapotranspiration processes. Soil water concentration varies with time between irrigations and with depth in the soil profile. When salt concentration is doubled, affinity of the soil for the monovalent ion increases, and the SAR of the concentrated solution is multiplied by the factor $\sqrt{2}$ ($\simeq 1.4$) as compared with the original solution. (2) Composition of the soil water is affected by salt precipitation and mineral weathering processes. Effects of these processes have limited the applicability of the SAR_{iw} value, *per se,* as a suitable index of irrigation water sodicity to relatively highly saline and low carbonate water. However, recent improvements such as the introduction of the Langelier index, pH_c^*, which serves as an index to the tendency of $CaCO_3$ to dissolve or precipitate during water passage through soils (Bower and Wilcox, 1965; Bower et al., 1965), and semiquantitative description of silicate mineral weathering process (Rhoades, 1968) have increased our ability to quantify the relation between the SAR_{iw} and soil ESP.

In spite of the obvious effect of electrolyte composition and concentration on soil permeability, the main emphasis has been given (Bresler et al., 1982; Rhoades, 1972; US Salinity Laboratory Staff, 1954) to the soil ESP and the SAR of the irrigation water as indexes of the degree to which sodium will occupy exchange positions on the clay surface. Because the effect of exchangeable sodium on swelling and dispersion is countered by a high electrolyte concentration, soil sodicity hazard cannot be assessed independently of information on the accompanying level of electrolyte concentration. Thus, improving estimates of soil ESP after irrigation by using more sophisticated cation-exchange theories is of academic interest but is not important in predicting the sodicity hazard of irrigating water. A much more important parameter is the change in the electrolyte concentration of the applied water and the change in the soil solution concentration during the growing season (or in between the growing seasons). Even in arid and semiarid regions where irrigation is practiced, rainstorms occur occasionally. Thus,

sodicity damage to the soil will not take place in the irrigation season when the electrolyte concentration in the water may be high enough to counter the dispersive effect of exchangeable sodium, but will occur in the rainy season when electrolyte concentration diminishes to values below the threshold concentration determined by the properties of the soil and its ESP. Also, soils in arid and semiarid regions often have moderate ESP values before irrigation, and for this reason the quantity of electrolytes in the irrigation water should be given greater consideration.

Predicting Effect of Irrigation Water on Soils

We have made a distinction between two soil processes connected with water movement: hydraulic conductivity, and infiltration rate. The HC is a property of the soil profile (which is protected from the stirring impact of applied water), but the IR depends on the hydraulic changes that take place at the soil surface. Water infiltration into the soil is the soil property most sensitive to soil ESP and the electrolyte concentration in the applied water. Water infiltration is more sensitive than the hydraulic conductivity of the soil profile for the following two reasons:

(1) The impact of rain (or irrigation) drops causes the mechanical breakdown of soil aggregates, and the stirring of the soil surface enhances the kinetics of clay dispersion.

(2) Concentration of electrolytes in the surface soil solution is determined solely by the composition of the applied water, because dissolution of lime and primary minerals is too slow to affect the surface solution concentration. Thus, when low-salinity water is applied (rainstorms or snow water), salt concentration in the soil surface solution remains very low and clay dispersion is possible. On the other hand, soil solution concentration deeper in the soil profile, even when irrigated (or rained upon) with low-salinity water, is affected by $CaCO_3$ and primary mineral dissolution and might be maintained at a concentration above 3 meq/l, which is enough to prevent clay dispersion at low ESP values.

In evaluating the effect of irrigation water on soil properties, the effect of the waters on each process should be discussed separately, and theoretical and laboratory results should be compared with field observations (Ayers and Westcot, 1976; Ayers and Tanji, 1981; Rhoades, 1977).

The effect of water quality on HC

Many studies have been done on the effect of exchangeable sodium on the HC of soils when soil sodicity is moderate to high (ESP >10) and water salinity of the water is moderate to high (EC >0.5 dS/m). Much less study has been devoted to the effect of low-salinity water on the HC of low-sodicity soils. The soil processes causing the reduction in soil HC and the soil properties affecting this reduction are sufficiently different at the two concentration ranges that the two phenomena should be discussed separately.

Waters of moderate to high salinity and ESP. Because a soil's HC depends on its ESP and the salt concentration of the percolating solution, any change in the HC should be considered in terms of combination of the two parameters. Quirk and Schofield (1955) combined the two parameters in their "threshold concentration" concept, which was defined as the salt concentration which caused a 10- to 15-percent decrease in soil per-

meability at a given ESP value. McNeal et al. (1966) suggested that a 25-percent decrease was more appropriate. The choice of the percentage of decrease is arbitrary. However, upon examining the HC changes observed by Quirk and Schofield (1955) or McNeal et al. (1966) it is seen that for many soils and percolating solution concentrations only a gradual drop in HC was observed, and the whole change in HC was sometimes not more than 25 percent. Soils having a HC equal to 0.75 of the "base" value still have reasonably high permeability. Further, there is a high variability in the HC of soils—even the base hydraulic conductivity can only be defined within a ± 10-percent range. Finally, in laboratory studies the soils are usually more sensitive to sodicity and electrolyte concentration than are soils in the field. Thus, it seems most appropriate to define a 50-percent decrease in soil permeability as critical, and the combination of salt concentration and ESP required to produce a 50-percent change in permeability as the threshold concentration.

The HC of soils having an ESP below 10 is not affected by waters of salinity above 0.5 dS/m (Felhendler et al., 1974; Frenkel et al., 1978; McNeal et al., 1966, 1968; Pupisky and Shainberg, 1979; Shainberg et al., 1981*a,b*). An electrolyte concentration of 0.5 dS/m is enough to prevent the deleterious effect of low levels of exchangeable sodium.

Significant changes (>50 percent) in soil HC occur only when the sodicity of the soil exceeds an ESP of 15 to 20. This is illustrated in Figure 30, where the threshold concentration as a function of the ESP is presented for the Waukena and Pachappa soils studied by McNeal et al. (1966), and for the micaceous Rothamsted soil studied by Quirk and Schofield (1955). The HC of montmorillonitic soil (Waukena) was more sensitive to sodic conditions than the HC of the illitic soils (Pachappa and Rothamsted). The opposite might be expected for IR where dispersion predominates and illite is more sensitive than montmorillonite. The difference between the Pachappa and Waukena soils was not evident in Figure 11, where a 25-percent decrease in the HC was assumed to be critical for the change in HC. This is another justification for our choice of 50-percent change in HC as the critical value.

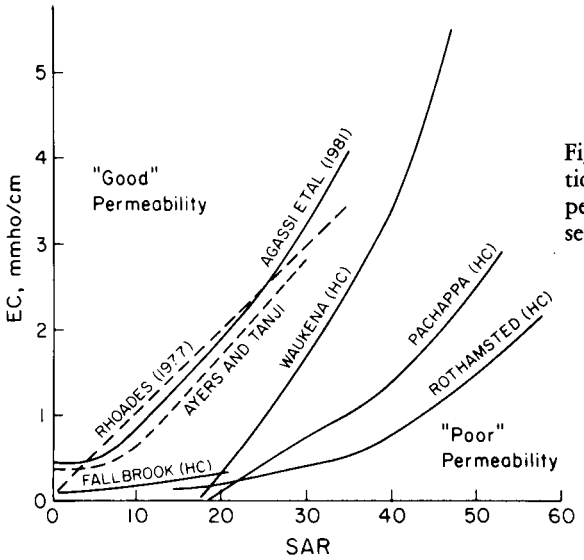

Fig. 30. Combination of salt concentration and ESP required to produce a 50-percent reduction in HC and IR for selected soils.

For Waukena and Pachappa soils there was a similar dependence of both clay swelling and increase in threshold concentration on the soil ESP. This similarity caused McNeal et al. (1966, 1968) to speculate that at moderate to high salinity clay swelling is the main mechanism affecting HC. Thus, the response of soils at the moderate to high concentration range depends on clay percentage (soils of higher clay percentage are more sensitive to ESP) and the clay type (montmorillonite soils swell more and are more sensitive to sodic conditions than are illitic soils).

Extrapolated threshold concentration curves to low salinity intersect with the x-axis at ESP values above 10 (Fig. 30), which suggests that the HCs of soils having ESPs below 10 are not affected by salt concentration in the percolating solution. According to this extrapolation, leaching with distilled water would result in soil HC being maintained at reasonable values.

This was not experimentally the case, and this fact illustrates the danger of extrapolating results when a change in mechanism occurs. The data at higher ESPs and water salinities seem to be related to clay swelling as the primary mechanism, whereas dispersion is the primary mechanism at low ESPs and water salinities.

Waters of low salinity. When salt concentration in percolating water is below the flocculation value of the clay, clay dispersion, movement, and lodgement in the conductivity pores comprise the main mechanism causing HC reduction. Macroscopic movement and lodgement of clay particles in a soil column leached with distilled water were reported by Felhendler et al. (1974), Frenkel et al. (1978), and Pupisky and Shainberg (1979). Macroscopic movement of clays in low-salinity water was reported by Shainberg et al. (1981*a*).

The flocculation values of montmorillonite clay at an ESP range up to 15 are below 5 meq/l (Oster et al., 1980) (Fig. 6). Hence, clay dispersion and movement take place only when salt concentration is below this value. Thus in agreement with Ayers and Westcot (1976), we also define waters of low salinity for montmorillonitic soils as waters with EC < 0.5 dS/m (~ 5 meq/l). The threshold concentration for waters of low salinity depends on soil ESP and clay type—illitic clays might be more dispersive and require a higher threshold concentration because flocculation values for illite are higher than for montmorillonite in a pure system (Fig. 6).

When waters of low salinity (e.g., rain or snow water) are used, the predominant mechanism causing HC reduction in soils of low sodicity is clay dispersion. Unlike swelling, dispersion does not require high soil clay content or montmorillonite clay, and may be more effective in soils of low clay content. Thus, Pupisky and Shainberg (1979) obtained an 80-percent reduction in the HC of sandy soil having only 3.1 percent clay when equilibrated with SAR 15 solution and leached with distilled water. Felhendler et al. (1974) obtained complete sealing of a soil of 10 percent clay when equilibrated with SAR 10 solution and leached with distilled water. Similar observations were made by Frenkel et al. (1978), Rowell and Shainberg (1979), Alperovitch et al. (1981), and Shainberg et al. (1981*a*). Thus when low salinity water is used, even sandy loam soils may seal.

The difference between swelling and dispersion mechanisms is important in another aspect. Swelling is essentially a reversible process: reduction in soil HC can be reversed by adding electrolytes or divalent cations to the system. On the other hand, dispersion and clay movement are essentially irreversible: they cause formation of crusts or impermeable clay pans which can only be disrupted by mechanical means.

An important soil property which determines the HC of soils leached with waters of very low salinity is the soils' potential for releasing electrolytes. Calcareous soils and soils

containing primary minerals such as feldspars and plagioclase are examples of soils which release electrolytes into solution readily in the presence of exchangeable sodium. Soil solution concentration is maintained above the flocculation value in these soils, so clay does not disperse and the HC change is limited. Conversely, the soil solution concentration in soils which do not release electrolytes is readily determined by the concentration of the applied solution. When these soils are leached with low-salinity water, the salt concentration remains low and causes clay dispersion and marked HC reduction.

The threshold concentration affecting the HC of Fallbrook soil at a low salinity level may be derived from Figure 16, and these numbers are presented in Figure 30. A high concentration of salts (>4 meq/l) prevents dispersion in the low ESP range.

Effect of water quality on the infiltration rate

The effect of electrolyte concentration of the applied water on the IR of soils was studied by Agassi et al. (1981) and Hardy et al. (1983), and was presented in Figures 25, 26, and 27. Considering, for example, the Nahal-Oz loess, one can see that the IR of a soil having an ESP of 6.4 responded to increased concentrations of electrolyte above EC 0.5 dS/m. Using the HC studies and the flocculation values of reference clay as a base (Fig. 6), we concluded that a salt concentration of 5 meq/l was enough to prevent dispersion of clays having an ESP of 6.4. Aggregates at the soil surface are more sensitive to breakdown and dispersion than are aggregates within the soil profile or those protected by filter paper or sand from the disturbance caused by applied water.

The combinations of soil ESP and applied-water salt concentration which result in 50 percent reduction in the IR can be calculated from the data in Figures 25 and 26, and are presented in Figure 30, identified as "Agassi et al."

Effect of water quality on soils—field observation

Summarizing the experience in the field, Ayers and Westcot (1976), Rhoades (1977), and Ayers and Tanji (1981) suggested EC values for irrigation water which are needed to prevent the deleterious effect of sodium. Table 2 gives the most recent values. It is evident from this table that waters with salinity below 0.2 dS/m cause severe permeability problems even at the ESP range of 0 to 3, and that threshold concentration increases as soil SAR and ESP increase. The combinations of SAR and EC of the irrigation water which result in moderate permeability problems (as presented in Table 2) are plotted in Figure 30, identified as "Ayers and Tanji."

Table 2. GUIDELINES FOR INTERPRETATION OF WATER QUALITY FOR IRRIGATION*

	EC dS/m		
SAR	No problem	Slight to moderate	Severe problem
0 to 3	> 0.9	0.9 to 0.2	< 0.2
3 to 6	> 1.3	1.3 to 0.25	< 0.25
6 to 12	> 2.0	2.0 to 0.35	< 0.35
12 to 20	> 3.1	3.1 to 0.9	< 0.9
20+	> 5.6	5.6 to 1.8	< 1.8

* Ayers and Tanji, 1981

Figure 30 also shows critical values of SAR and EC of the irrigation waters as proposed by Rhoades (1977). These numbers are based on judgment and field experience, and not on laboratory data. In the words of Rhoades, "...since soils vary widely with regards to the effects of exchangeable sodium and electrolyte concentration on permeability it is difficult to specify critical levels.... The critical values that I used to judge the likelihood of permeability problems are given in fig 8. [Fig. 30 in this paper] and apply to the more-sensitive soils."

The dotted lines in Figure 30, which are based on field observation, compare favorably with laboratory results on IR (the continuous line). The soils in the field are more sensitive to ESP than laboratory HC measurements suggest. Ayers and Tanji (1981) and Rhoades (1977) noted that their critical numbers were based on their experience with the IR of water into the soil. As was emphasized in this review, IR is more sensitive than HC to ESP.

Recent field studies on the effect of exchangeable sodium (at ESP values of 4.6 and 19.3) on surface runoff and soil erosion under natural rain conditions confirmed the IR laboratory studies (Frenkel and Hadas, 1981; Keren et al., 1983). Measurements of water runoff and soil loss were conducted in standard Wischmeier plots (25 m long and 6 m wide) on bare soil having a slope of about 2 to 4 percent (Keren et al., 1983). Soils having an ESP of 4.6 predominated in the studied area (northwestern Negev in Israel), whereas fields having ESP 19.3 had been commercially irrigated for 4 years with saline/sodic waters (EC$\simeq$5.5 dS/m^{-1} and SAR $\simeq$18). Table 3 shows surface runoff and the beneficial effect of phosphogypsum treatments. The pronounced effect of exchangeable sodium and phosphogypsum is evident. Even in a dry winter of only 182 mm of rain in seven storms, the total amount of runoff from soil having an ESP of 4.6 was 22.8 mm, and the amount of runoff from the sodic soil was 64.2 mm ($\simeq$35.3 percent of the total rain).[2] Spreading phosphogypsum at the rate of 5 ton/ha on soil of ESP 4.6 reduced the yearly amount of runoff from 22.8 to 3.8 mm. Similar gypsum treatment to soil with ESP 19.3 reduced runoff from 64.2 mm in the control to 30.1 mm in the mixed treatment (where gypsum was disced into the upper 10 cm layer) and 18.7 mm in the spread-over treatment. Considering the rain intensities, these results were in good agreement with the prediction based on laboratory studies (Kazman et al., 1983) (section starting on page 33). The high amount of runoff from this calcareous loess soil of ESP 4.6 and the beneficial effect of phosphogypsum confirmed the hypothesis that even low percentages of sodium are enough to cause aggregate dispersion and crust formation. The mechanical impact of the raindrops enhances the rate of crust formation and the amount of runoff. However, as rain intensity was the same for control and gypsum-treated soils, it is evident that the chemical dispersion (enhanced by sodicity and electrolyte concentration) plays a dominant role in affecting the infiltration rate.

[2] In a set winter (the present 1982/83 winter) the recorded rain was 520 mm and the percent of runoff in a 25 m long plot in a soil with ESP 4.6 was $\simeq$ 30 percent (157 mm) and in plots of 5 m long, the percent of runoff was 45 percent (234 mm).

Table 3. EFFECT OF PHOSPHOGYPSUM TREATMENTS ON SURFACE RUNOFF IN RAINSTORMS FROM LOESS SOIL AT TWO LEVELS OF ESP*

			Surface runoff (percent of rainfall)				
			ESP 4.6		ESP 19.3		
Storm number	Amount of rainfall (mm)	Time between storms (days)	Control	Phospho-gypsum spread over†	Control	Phospho-gypsum mixed in†	Phospho-gypsum spread over†
1	16	—	0	0	23.1	2.6	0
2	20	40	6.5	0	13.3	3.0	0.7
3	19	20	1.3	0.5	27.3	10.5	3.8
4	59	6	21.5	3.8	41.0	21.2	17.2
5	44	9	13.4	2.1	45.0	22.5	12.4
6	12	1	12.5	2.5	40.0	25.0	12.5
7	12	15	10.1	2.5	31.7	13.8	6.7
Total (mm)	182		22.8	3.8	64.2	30.1	18.7
Percent of annual rain			12.5	2.1	35.3	16.5	10.3

*From Keren et al., 1983.
†The rates of phosphogypsum application were 5 and 10 t/ha for the soils having ESP 4.6 and 19.3, respectively.

Research Needs

Water infiltration rate is the soil property most sensitive to the ESP of soil and the electrolyte concentration in applied water. Apparently IR measurements more closely approximate field observations than do HC measurements. Although many studies have been made on the effect of salinity and sodicity on soil HC, only a few studies (most of them within the last 3 years) have been done on the effect of water salinity and soil sodicity on the IR of water into soils. Future research should concentrate on the interaction between soil properties (texture, mineralogy, chemical composition, etc.) and applied water properties (solution composition, method of application), and their effect on the IR of soils. Furthermore, most of the studies on the effect of sodicity and salinity on IR were made in the laboratory. The new concepts have to be further tested in the field.

Laboratory studies: soil properties affecting the IR response to sodic conditions

Soil texture. Studies of the HC of soils indicate that soils having more clay are more sensitive to sodic condition. When clay swelling is the main mechanism causing changes in HC, this is a reasonable conclusion. Soil at the soil surface is not confined, and swelling results in increased soil porosity. When clay dispersion is the main mechanism causing the drop in IR, only a small amount of dispersed clay is needed to clog the pores. Thus even sandy loam soils should be sensitive to sodic conditions and low salinity—this explains why low infiltration is also found in coarse-textured soils. Although this explanation is plausible, only limited experimental data are available to provide proof.

Clay mineralogy. When swelling is the main mechanism for permeability changes, soils with smectite clay should be more sensitive to sodic conditions than are soils having illite clay. Because illitic clay is more dispersive than montmorillonitic clay, it is expected that the IR of illitic soils will be more sensitive to sodic conditions than that of montmorillonitic soils. This hypothesis has not been sufficiently investigated and must be considered tentative.

Effect of cementing materials. Sesquioxides and organic matter are known to stabilize soil aggregates. Stronger dispersive conditions (such as higher sodicity, lower salinity, and higher energy or impact of the irrigation water) should be needed to disperse a stable aggregate. There are not enough data on this subject to draw conclusions.

Specific effect of exchangeable magnesium and exchangeable potassium. Specific effect of magnesium on the HC of soils has been observed only in very dilute solution (rainwater). Swelling properties of Mg-montmorillonite are similar to those of Ca-montmorillonite, thus no specific effect of magnesium on the HC is expected or observed. However, it is possible that dispersion and low infiltration are more dominant in Mg-soils than in Ca-soils. Similarly, the effect of low percentages of potassium (up to 10 percent) on clay dispersion and IR has not been studied. Many soils throughout the world are rich in one or the other of these cations, and further investigations are warranted.

Effect of amendments on the IR of soils. The effect of gypsum (various particle sizes and various rates of application) and other amendments on the IR of soil should be quantitatively studied.

Laboratory studies: effect of applied water parameters on the IR (in addition to the SAR and salinity of the waters)

Effect of drop impact energy on the IR and crust formation. The size of the water drops and their terminal velocities determine their impact energy. The higher the impact energy, the more sensitive might be the soil to soil sodicity and solution salinity. The spectrum of drop impact energy is very wide for rainstorms and for the various methods of irrigation, and should be investigated from a water- and soil-quality point of view.

Effect of the rate of water flow in furrow and flood irrigation on IR and crust formation. Higher rate of flow introduces higher stirring of the soil surface, which in turn enhances aggregate and clay dispersion. Fast wetting also results in more stirring as well as release of air bubbles through the soil surface. Understanding the effect of water flow on the IR and crust formation may enable us to understand basic principles involved in determining effectiveness of the surge method of irrigation (Bishop et al., 1981).

Field studies

Most of the concepts developed in this review are based on laboratory experiments. It is difficult to study the effect of sodicity and salinity in controlled field experiments where soils of various sodicities and waters of various salinities and compositions are all at the same location. Thus, in field experiments only a limited aspect of the approach described in this review can be tested. Because of their complexity, field studies should respond to the recognized infiltration problems in the regions. The following two cases are given as examples.

Improving rainwater infiltration in dry-land agriculture. Annual rainfall in the semiarid and arid regions is low, uncertain, and patchy. Paradoxical as it may seem, in spite of the deficiency in rainfall, enormous amounts of water can be lost due to runoff. For example, the amount of runoff measured in the northern part of the Negev in Israel in small plots (6 m^2) ranged between 30 and 50 percent of the annual precipitation (Hillel, 1971). The arable soils in this region contain 2 to 8 percent exchangeable sodium in the upper layer. Agassi et al. (1982) postulated that the combination of these ESP values and the low salinity of the rainwater caused crust formation and thus high runoff. This hypothesis must still be verified in other arid and semiarid regions of the world—the effect of other soil properties which could enhance crust formation should also be studied.

If the above hypothesis is valid, phosphogypsum or other amendments allowing rapid dissolution of electrolytes are helpful in preventing crust formation, maintaining high infiltration rates, and reducing runoff. Studies on the rate, timing, and method of application of cheap chemical amendments to determine their effectiveness in combination with soil properties should be done.

Improving water infiltration in the Central Valley of California. Many growers in the Central Valley who originally irrigated with well waters having a moderate concentration of salt (EC$\simeq$0.7 dS/m, SAR 2 to 5) found that when they started using imported Friant-Kern canal water, which had a very low salt concentration (EC$\simeq$0.049 dS/m, SAR $<$0.5), a marked reduction in water infiltration occurred (Mohammed et al., 1979). Applying gypsum to the waters at rates of 30 to 675 ppm significantly increased the hydraulic conductivity (Mohammed et al., 1979). This field problem is similar to that described in the previous example, and it may again be hypothesized that the low ESP in the soil, in combination with the low-salinity irrigation water, caused the drop in infiltration rate. Here, research similar to that described in the previous example is needed.

LITERATURE CITED

ABDER-RAHMAN, W., and D. L. ROWELL
1979. The influence of Mg in saline and sodic soils: a specific effect or a problem of cation exchange. J. Soil Sci. 30:535–46.
AGASSI, M., J. MORIN, and I. SHAINBERG
1982. Infiltration and runoff control in the semi-arid region of Israel. Geoderma (in press).
AGASSI, M., I. SHAINBERG, and J. MORIN
1981. Effect of electrolyte concentration and soil sodicity on infiltration rate and crust formation. Soil Sci. Soc. Am. J. 45:848–51.
ALPEROVITCH, N., I. SHAINBERG, and R. KEREN
1981. Specific effect of magnesium on the hydraulic conductivity of sodic soils. J. Soil Sci. 32:543–54.
ARORA, H. S., and N.T. COLEMAN
1979. The influence of electrolyte concentration on flocculation of clay suspensions. Soil Sci. 127:134–39.
AYERS, R. S., and K. K. TANJI
1981. Agronomic aspects of crop irrigation with waste water. Proc. Conf. Water Forum '81. Am. Soc. Civil Eng. (Pub.).
AYERS, R. S., and D. W. WESTCOT
1976. Water quality for agriculture. FAO Irrigation and Drainage Paper No. 29. Food and Agriculture Organization of the United Nations, Rome.
AYLMORE, L. A. G., and J. P. QUIRK
1959. Swelling of clay-water systems. Nature 183:1752–53.
BABCOCK, K. L.
1963. Theory of the chemical properties of soil colloidal systems at equilibrium. Hilgardia 34:417–542.
BAR-ON, P., I. SHAINBERG, and I. MICHAELI
1970. The electrophoretic mobility of Na/Ca montmorillonite particles. J. Colloid and Interface Sci. 33:471–72.
BAVER, L. D., W. H. GARDNER, and W. R. GARDNER
1972. Soil Physics. 4th ed. New York: John Wiley & Sons.
BISHOP, A. A., W. R. WALKER, N. L. ALLEN, and G. J. POOLE
1981. Furrow advance rate under surge flow systems. J. Irrig. and Drain. Div. ASCE. 107(IR3):257–64.
BLACKMORE, A. V., and R. D. MILLER
1961. Tactoid size and osmotic swelling in Ca montmorillonite. Soil Sci. Soc. Am. Proc. 25:169–73.
BOHN, H. L., B. L. McNEAL, and G. A. O'CONNOR
1979. Soil Chemistry. New York: Wiley-Interscience. 329 pp.
BOLT, G. H. (ed.)
1979. Soil Chemistry. B. Physico-Chemical Models. Amsterdam: Elsevier Scientific Pub. Comp. 479 pp.
BOWER, C. A.
1959. Cation exchange equilibria in soils affected by sodium salts. Soil Sci. 87:32–35.
BOWER, C. A., and L. V. WILCOX
1965. Precipitation and solution of calcium carbonate in irrigation operations. Soil Sci. Soc. Am. Proc. 29:93–94.
BOWER, G. A., L. V. WILCOX, G. W. AIKEN, and M. G. KEYES
1965. An index of the tendency of $CaCO_3$ to precipitate from irrigation waters. Soil Sci. Soc. Am. Proc. 29:91–92.
BRESLER, E.
1970. Numerical solution of the equation for interacting diffuse layers in mixed ionic system with non-symmetrical electrolytes. J. Colloid and Interface Sci. 33:278–83.
1972. Interacting diffuse layers in mixed mono-divalent ionic systems. Soil Sci. Soc. Am. Proc. 36: 891–96.
BRESLER, E., B. L. McNEAL, and D. L. CARTER
1982. Saline and Sodic Soils. New York: Springer-Verlag.
BROWN, G.
1961. The x-ray identification and crystal structures of clay minerals. Mineralogical Society. London.
CARY, J. W., and D. D. EVANS (eds.)
1974. Soil crusts. Tech. Bull. No. 214, Univ. of Arizona, Tucson.
CASS, A., and M. E. SUMNER
1982*a*. Soil pore structural stability and irrigation water quality. I. Empirical sodium stability model. Soil Sci. Soc. Am. J. 46:503–06.
1982*b*. Soil pore structural stability and irrigation water quality. II. Sodium stability data. Soil Sci. Soc. Am. J. 46:507–12.
CHEN, Y., J. TARCHITZKY, J. MORIN, and A. BANIN
1980. Scanning electron microscope observations on soil crusts and their formation. Soil Sci. 130:49–55.
CHI, C. L., W. W. EMERSON, and D. G. LEWIS

1977. Exchangeable Ca, Mg, and Na and the dispersion of illite in water. I. Characterization of illites and exchange reactions. Aust. J. of Soil Res. 15:243–53.

DESHPANDE, T. L., D. J. GREENLAND, and J. P. QUIRK
1968. Changes in soil properties associated with the removal of iron and aluminum oxides. J. Soil Sci. 19:108–22.

ELLIS, J. H., and O. G. CALDWELL
1935. Magnesium clay solonetz. Trans. of the 3rd Intern. Cong. of Soil Sci. 1:348–50.

EL-SWAIFY, S. A.
1973. Structural changes in tropical soils due to anions in irrigation water. Soil Sci. 115:64–72.
1976. Changes in the physical properties of soil clays due to precipitated Al and Fe hydroxides. II. Colloidal interaction in the absence of drying. Soil Sci. Soc. Am. Proc. 40:516–20.

EL-SWAIFY, S. A., and W. W. EMERSON
1975. Changes in the physical properties of soil clays due to precipitated Al and Fe hydroxides. I. Swelling and aggregate stability after drying. Soil Sci. Soc. Am. Proc. 39:1056–63.

EMERSON, W. W.
1977. Physical properties and structure. *In* Soil Factors in Crop Production in a Semi-Arid Environment (J. S. Russell and E. L. Greacen, Eds.), pp. 78–104. St. Lucia, Queensland: Univ. of Queensland Press.

EMERSON, W. W., and A. C. BAKKER
1973. The comparative effect of exchangeable Ca, Mg, and Na on some physical properties of red brown earth subsoils. II. The spontaneous dispersion of aggregates in water. Aust. J. of Soil Res. 11:151–57.

EMERSON, W. W., and C. L. CHI
1977. Exchangeable calcium, magnesium and sodium and the dispersion of illites. II. Dispersion of illites in water. Aust. J. of Soil Res. 15:255–63.

EVANS, D. D., and S. W. BUOL
1968. Micromorphological study of soil crust. Soil Sci. Soc. Am. Proc. 32:19–22.

FELHENDLER, R., I. SHAINBERG, and H. FRENKEL
1974. Dispersion and hydraulic conductivity of soils in mixed solution. Trans. of the 10th Intern. Cong. of Soil Sci. (Moscow) I:103–12. Moscow: Nauka Pub. House.

FRENKEL, H., J. O. GOERTZEN, and J. D. RHOADES
1978. Effects of clay type and content, exchangeable sodium percentage, and electrolyte concentration on clay dispersion and soil hydraulic conductivity. Soil Sci. Soc. Am. J. 42:32–39.

FRENKEL, H., and A. HADAS
1981. Effect of tillage and gypsum incorporation on rain runoff and crust strength in field soils irrigated with saline-sodic water. Soil Sci. Soc. Am. J. 45:156–58.

FRENKEL, H., and I. SHAINBERG
1975. Chemical and hydraulic changes in soils irrigated with brackish water. *In* Irrigation with Brackish Water. Intern. Symp. Beersheva, Israel.

GREENE, R. S. B., A. M. POSNER, and J. P. QUIRK
1978. A study of the coagulation of montmorillonite and illite suspensions by $CaCl_2$ using the electron microscope. *In* Modification of Soil Structure (W. W. Emerson, R. D. Bond, and A. R. Dexter, Eds.), pp. 35–40. New York: John Wiley and Sons.

GRIM, R. E.
1968. Clay Mineralogy. 2nd ed. New York: McGraw-Hill.

HARDY, N., I. SHAINBERG, M. GAL, and R. KEREN
1983. The effect of water quality and storm sequence upon infiltration rate and crust formation. J. Soil Sci. (in press).

HILLEL, D.
1971. Runoff inducement in arid lands. Final technical report submitted to USDA Proj. No. AID-SWC-36.
1980. Applications of Soil Physics. New York: Academic Press.

KAZMAN, Z., I. SHAINBERG, and M. GAL
1983. Effect of low levels of exchangeable Na and applied phosphogypsum on the infiltration rate of various soils. Soil Sci. 135:184–92.

KEREN, R.
1979. Na-Ca exchange in the presence of hydroxy Al.

KEREN, R., and G. A. O'CONNOR
1982. Gypsum dissolution and sodic soil reclamation as affected by water flow velocity. Soil Sci. Soc. Am. J. 46:726–32.

KEREN, R., and I. SHAINBERG
1981. Effect of dissolution rate on the efficiency of industrial and mined gypsum in improving infiltration of a sodic soil. Soil Sci. Soc. Am. J. 45:103–07.

KEREN, R., I. SHAINBERG, H. FRENKEL, and Y. KALO
1983. A field study of the effect of exchangeable sodium on surface runoff from loess soil. Soil Sci. Soc. Am. J. (in press).

KREIT, J. F., I. SHAINBERG, and A. J. HERBILLON
1982. Hydrolysis and decomposition of hectorite in dilute salt solution. Clays and Clay Minerals 30:223–31.

LAGERWERFF, J. V., F. S. NAKAYAMA, and M. H. FRERE
1969. Hydraulic conductivity related to porosity and swelling of soil. Soil Sci. Soc. Am. Proc. 33:3–11.
LEVY, R., and D. HILLEL
1968. Thermodynamics equilibrium constants of Na/Ca exchange in some Israeli soils. Soil Sci. 106:393–98.
LOVEDAY, J.
1976. Relative significance of electrolyte and cation exchange effects when gypsum is applied to a sodic clay soil. Aust. J. of Soil Res. 14:361–71.
MARSHALL, T. J.
1958. A relation between permeability and size distribution of pores. J. Soil Sci. 9:1–8.
McATEE, J. L.
1961. Heterogeneity in montmorillonites. Clays and Clay Minerals. Proc. Natl. Conf. Clays Clay Minerals, 5th Conf. 1956, 279–88.
McINTYRE, D. S.
1958. Permeability measurements of soil crusts formed by raindrop impact. Soil Sci. 85:185–89.
1979. Exchangeable sodium, subplasticity and hydraulic conductivity of some Australian soils. Aust. J. Soil Res. 17:115–20.
McNEAL, B. L.
1968. Prediction of the effect of mixed-salt solutions on soil hydraulic conductivity. Soil Sci. Soc. Am. Proc. 32:190–93.
McNEAL, B. L., and N. T. COLEMAN
1966. Effect of solution composition on soil hydraulic conductivity. Soil Sci. Soc. Am. Proc. 30:308–12.
McNEAL, B. L., D. A. LAYFIELD, W. A. NORVELL, and J. D. RHOADES
1968. Factors influencing hydraulic conductivity of soils in the presence of mixed-salt solutions. Soil Sci. Soc. Am. Proc. 32:187–90.
McNEAL, B. L., W. A. NORVELL, and N. T. COLEMAN
1966. Effect of solution composition on soil hydraulic conductivity and on the swelling of extracted soil clays. Soil Sci. Soc. Am. Proc. 30:308–15.
MOHAMMED, EL TAYIB YOUSIF, J. LETEY, and R. BRANSON
1979. Sulphur compounds in water treatment. Sulphur in Agriculture 3:7–11.
MORIN, J., Y. BENYAMINI, and A. MICHAELI
1981. The dynamics of soil crusting by rainfall impact and the water movement in the soil profile. J. of Hydrology 52:321–35.
MORIN, J., S. GOLDBERG, and I. SEGINER
1967. A rainfall simulator with a rotating disk. Trans. Am. Soc. Agric. Eng. 10:74–79.
OSTER, J. D.
1982. Gypsum usage in irrigated agriculture: A review. Fertilizer Research 3:73–89.
OSTER, J. D., and H. FRENKEL
1980. The chemistry of the reclamation of sodic soils with gypsum and lime. Soil Sci. Soc. Am. J. 44:41–45.
OSTER, J. D., and F. W. SCHROER
1979. Infiltration as influenced by irrigation water quality. Soil Sci. Soc. Am. J. 43:444–47.
OSTER, J. D., and I. SHAINBERG
1979. Exchangeable cation hydrolysis and soil weathering as affected by exchangeable sodium. Soil Sci. Soc. Am. J. 43:70–75.
OSTER, J. D., I. SHAINBERG, and J. D. WOOD
1980. Flocculation value and gel structure of Na/Ca montmorillonite and illite suspension. Soil Sci. Soc. Am. J. 44:955–59.
PUPISKY, H., and I. SHAINBERG
1979. Salt effects on the hydraulic conductivity of a sandy soil. Soil Sci. Soc. Am. J. 43:429–33.
QUIRK, J. P.
1978. Some physico-chemical aspects of soil structural stability: a review. *In* Modification of Soil Structure (W. W. Emerson, R. D. Bond, and A. R. Dexter eds.) New York: John Wiley and Sons. pp. 3–16.
QUIRK, J. P., and L. A. G. AYLMORE
1971. Domains and quasi-crystalline regions in clay systems. Soil Sci. Soc. Am. Proc. 35:652–54.
QUIRK, J. P., and R. K. SCHOFIELD
1955. The effect of electrolyte concentration on soil permeability. J. Soil Sci. 6:163–78.
RHOADES, J. D.
1968. Mineral-weathering correction for estimating the sodium hazard of irrigation waters. Soil Sci. Soc. Am. Proc. 32:648–52.
1972. Quality of water for irrigation. Soil Sci. 113:277–84.
1977. Potential for using saline agricultural drainage waters for irrigation. Proc. of Water Management for Irrigation and Drainage. ASCE, Nevada.
RHOADES, J. D., and R. D. INGVALSON
1969. Macroscopic swelling and hydraulic conductivity properties of four vermiculite soils. Soil Sci. Soc. Am. Proc. 33:364–69.

RHOADES, J. D., D. B. KRUEGER, and M. J. REED
1968. The effect of soil-mineral weathering on the sodium hazard of irrigation waters. Soil Sci. Soc. Am. Proc. 32:643–47.
RIMMER, D. L., and D. J. GREENLAND
1976. Effect of $CaCO_3$ on the swelling of a soil clay. J. Soil Sci. 27:129–39.
ROWELL, D. L., D. PAYNE, and N. AHMAD
1969. The effect of the concentration and movement of solutions on the swelling, dispersion and movement of clay in saline and alkali soils. J. Soil Sci. 20:176–88.
ROWELL, D. L., and I. SHAINBERG
1979. The influence of magnesium and of easily weathered minerals on hydraulic conductivity changes in a sodic soil. J. Soil Sci. 30:719–26.
RUSSO, D., and E. BRESLER
1977*a*. Effect of mixed Na/Ca solutions on the hydraulic properties of unsaturated soils. Soil Sci. Soc. Am. J. 41:713–17.
1977*b*. Analysis of the saturated-unsaturated hydraulic conductivity in a mixed Na/Ca soil system. Soil Sci. Soc. Am. J. 41:706–10.
SHAINBERG, I., E. BRESLER, and Y. KLAUSNER
1971. Studies on Na/Ca montmorillonite systems. I. The swelling pressure. Soil Sci. 111:214–19.
SHAINBERG, I., and M. GAL
1982. The effect of lime on the response of soils to sodic conditions. J. Soil Sci. (in press).
SHAINBERG, I., and A. KAISERMAN
1969. Kinetics of the formation and breakdown of Ca-montmorillonite tactoids. Soil Sci. Soc. Am. Proc. 33:547–51.
SHAINBERG, I., and W. D. KEMPER
1966. Electrostatic forces between clay and cations calculated and inferred from electrical conductivity. Clays and Clay Minerals. Proc. of 14th Conf. 117–32.
SHAINBERG, I., R. KEREN, and H. FRENKEL
1982. Response of sodic soils to gypsum and $CaCl_2$ application. Soil Sci. Soc. Am. J. 46:113–17.
SHAINBERG, I., and J. D. OSTER
1978. Quality of irrigation water. IIIC Publication No. 2, Volcani Center, P.O. Box 6, Bet-Dagan, Israel.
SHAINBERG, I., J. D. OSTER, and J. D. WOOD
1980. Na/Ca exchange in montmorillonite and illite suspensions: Effect of salt concentration and exchanger composition. Soil Sci. Soc. Am. J. 44:960–64.
SHAINBERG, I., and H OTOH
1968. Size and shape of montmorillonite particles saturated with Na/Ca ions. Israel J. Chem. 6:251–59.
SHAINBERG, I., J. D. RHOADES, and R. J. PRATHER
1981*a*. Effect of low electrolyte concentration on clay dispersion and hydraulic conductivity of a sodic soil. Soil Sci. Soc. Am. J. 45:273–77.
SHAINBERG, I., J. D. RHOADES, D. L. SUAREZ, and R. J. PRATHER
1981*b*. Effect of mineral weathering on clay dispersion and hydraulic conductivity of sodic soils. Soil Sci. Soc. Am. J. 45:287–91.
SPOSITO, G.
1977. The Gapon and the Vanselow selectivity coefficients. Soil Sci. Soc. Am. J. 41:1205–06.
SPOSITO, G., and S. F. MATTIGOD
1979. Ideal behavior in Na^+-trace metal cation exhcange on Camp Berteau montmorillonite. Clays and Clay Minerals. 27:125–28.
US SALINITY LABORATORY STAFF
1954. Diagnosis and Improvement of Saline and Alkali Soils. USDA Agric. Handb. No. 60.
VAN BLADEL, R., G. GAVRIA, and H. LAUDELOUT
1972. A comparison of the thermodynamic, double-layer theory and empirical studies of the Na-Ca exchange equilibria in clay water systems. Proc. Intern. Clay Conf., pp. 385–98.
VAN OLPHEN, H.
1977. An Introduction to Clay Colloid Chemistry. 2nd Ed. New York: John Wiley & Sons.
WARKENTIN, B. P., G. H. BOLT, and R. D. MILLER
1957. Swelling pressure of montmorillonite. Soil Sci. Soc. Am. Proc. 21:495–97.
YARON, B., and G. W. THOMAS
1968. Soil hydraulic conductivity as affected by sodic water. Water Resources Res. 4:545–52.

2½m-1/84-WC/LAM