

THE EFFECT OF SALT COMPOSITION OF GROUNDWATERS ON THE
RATE OF SALINISATION OF SOILS FROM A WATER TABLE

by

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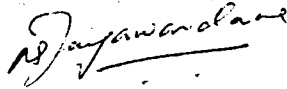
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degree of Doctor of Philosophy

UNIVERSITY OF TASMANIA

HOBART

July, 1977.

This thesis contains no material which has been accepted for the award of any other degree or diploma in any University, and to the best of my knowledge contains no copy or paraphrase of material previously published or written by any other person except where due reference is made in the text of the thesis.

A handwritten signature in dark ink, appearing to read 'N. S. Jayawardane', with a horizontal line drawn underneath the name.

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SUMMARY

The rate of salinisation of soils from a water table has been shown to be related to the concentration of salts in groundwaters and the rate of upward flow of saline groundwater. The maximum rate at which soil can transmit soil solution from a water table to the surface is governed by hydraulic conductivity characteristics and depth to water table. In the present study the effects of chemical composition of groundwaters on the rate of salinisation of three different soils by upward flow were evaluated in soil columns. These effects were related to changes in pore geometry and hydraulic conductivity characteristics of the soils in salt solutions.

The pore geometry, saturated and unsaturated conductivity, and the capacity of the soil to transmit soil solution from a water table to the surface varied little in krasnozem soil exposed to different salt solutions, but marked changes of different magnitudes occurred in the case of alluvial and red brown soils.

With a reduction in cation concentration of salt solution at a given SAR, the inter-aggregate pores of alluvial and red brown soils decreased in size, apparently due to swelling of soil aggregates. The extent of this decrease in pore sizes was greater for solutions of higher SAR. A pore size index was proposed to evaluate quantitatively these changes in pore geometry. Patterns in the volume changes of different sized pores with changes in salt composition were discerned and used to predict the moisture release curves of these soils for other salt solutions and also for soils at slightly different bulk densities in the same salt solutions.

The saturated and unsaturated conductivity at high pressure heads of alluvial and red brown soils decreased markedly with reduction in cation concentration at a given SAR, while unsaturated conductivity at low pressure heads showed only small decreases.

Existing methods for computing hydraulic conductivities from moisture release curves could not accurately predict the conductivities at saturation and high pressure heads in different salt solutions. It was suggested that the discrepancy was partly due to changes in internal pore geometry during desaturation as a result of aggregate shrinkage. These changes were described qualitatively in terms of a theoretical model.

A concept of equivalent salt solutions was also developed and used to predict, fairly closely, the hydraulic conductivities of the soils in different salt solutions. It is proposed that this method could be used to correlate directly flow of different salt solutions in soils under specific boundary conditions.

The computed maximum depths to water table for specific rates of upward flow in alluvial and red brown soils decreased as the cation concentration was reduced at a given SAR. However, marked decreases in maximum depths to water table for specific rates of flow occurred only with solutions of low cation concentration, in the presence of which the values of unsaturated conductivity within a critical range showed marked decreases. This critical range was defined using the data from a flux/unsaturated conductivity ratio method of computing upward flow. The changes in measured rates of upward flow, as the cation concentration was reduced at a given SAR, were generally similar in nature to the changes in computed rates. The application of these results and concepts to field problems and areas for future research have been discussed.

INTRODUCTION

Salinisation of soils by the intrusion of salts and the resultant loss in productivity especially in irrigated lands has been a problem confronting farmers from ancient times and it is a potential hazard in modern irrigation schemes. Capillary rise from saline water tables is one of the processes causing salinisation of agricultural lands and studies to determine the rate of upward flow in soils from water tables at different depths are necessary in designing drainage and agronomic practises for control of salinity.

The movement of salts from saline water tables into the soil above takes place primarily by convection in the upward flow of salt solutions. Thus the rate of salinisation is closely related to the concentration of salts in groundwaters and the rate of upward flow of groundwater, which equals the rate of evaporation under steady flow conditions. Several workers have demonstrated that the rate of evaporation can be limited either by the atmospheric evaporative conditions or the maximum rate at which soils can transmit water to the surface, whichever is less. Previous studies also show that except for shallow depths to water table or humid conditions, the evaporation rate is usually controlled by the capacity of the soil to transmit water to the surface which in turn is governed by the hydraulic conductivity characteristics of the soil and the depth to the water table. Hydrological analysis yields the rates of salinisation of a soil from water tables at specific depths and hence the critical depth below which the water table must be maintained to prevent excessive salinisation can be calculated.

However, in the field, groundwaters with wide variations in salt composition are encountered. The presence of certain salts in the soil solution could change the hydraulic conductivity characteristics of some soils and thus alter the capacity of these soils to transmit soil solution. This would result in changes in the rate of salinisation of soils from a water table. The magnitude of such changes in the rate of salinisation is likely to vary in different soils. The present study attempts to evaluate the effects of salt composition of groundwaters on the rate of salinisation of different soils, and to relate these effects to changes in the pore geometry and conductivity characteristics of these soils in the presence of different salt solutions.

Thus the objectives of the present study are:

- (a) to investigate changes in saturated and unsaturated hydraulic conductivities of soils for salt solutions of increasing sodium absorption ratio and decreasing cation concentration;
- (b) to assess the effects of salt solution composition on the pore geometry of different soils, to relate these changes to saturated and unsaturated conductivities in different salt solutions and to evaluate methods for computing unsaturated conductivities from the moisture release curves of the soils in the presence of the same salt solutions;
- (c) to compute the rate of steady upward flow of salt solutions in soils from groundwaters of different chemical composition using the unsaturated conductivity-pressure head curves of the soils for the same salt solutions and to verify these results experimentally, and
- (d) to propose a new method for computing upward flow from a

water table and to use this method for computing and comparing upward flow from groundwater of different chemical compositions.

II. REVIEW OF LITERATURE

A. Factors affecting the Salinisation of Soils from a Saline Water Table

The decline in productivity of some ancient and recent irrigation schemes has been attributed, at least partly, to secondary salinisation by capillary rise from saline water tables. A number of early studies in the field have shown that the depth to the water table is an important factor determining the extent of salinisation of soils and that deep drainage is necessary to control salinity by lowering the water table. Burgess (1928) found that in fine grained saline and alkali soils in Arizona, U.S.A., the water table had to be lowered to a depth of 240 to 300 cm to prevent salinisation. Field studies have shown that severe salinisation occurred in alluvial soils in Western Australia (Teakle and Burvill, 1945), soils in the Nile Delta (Schoonover *et al.*, 1957), and alluvial soils of the Hawkesbury River floodplain near Sydney, Australia (Collis-George and Evans, 1964) where the depth to the water table was less than 150 to 180 cm, 60 cm and 75 cm respectively.

Polynov (1930) defined the critical depth as "the maximum height above the water table to which salt contained in the groundwater can rise under natural conditions by capillary rise and diffusion". In a detailed review of early studies in the U.S.S.R., and eastern Europe to determine critical depths, Talsma (1963) recognized three groups as follows: (i) field and lysimeter experiments on salinisation of cropped lands (Sukhachev, 1958; Legostaev, 1958); (ii) studies showing the relationship of groundwater depth to the salt content of surface soil layers (Muratova, 1958; Florea and Stoica, 1958); (iii) studies showing the

relationship between the depth to water table and rate of evapotranspiration (Kabaev, 1958; Legostaev, 1958; Kovda, 1961). These studies showed that as the depth to water table increased the rate of salinisation of the soil decreased. It was also found that the critical depth corresponded approximately to an upward flow rate of 0.1 cm/day and that the critical depth to water table varied widely according to the soil and other site characteristics (Suchachev, 1958; Florea and Stoica, 1958; Kovda, 1961).

The importance of soil texture in modifying the critical depth to the water table was shown by the studies of Jackson *et al.* (1956), which indicated that the rate of salinisation of a sandy soil decreased sharply when the water table depth was more than 90 to 120 cm, while for clayey soil at the same location the critical depth was much greater. On the other hand Volobuev (1946) found that the presence of a saline water table at depths less than 120 to 130 cm in a clay soil and less than 130 to 150 cm in a loamy soil led to salinisation. Groenewegen (1959) showed that salt accumulation in surface layers was small in clay soils of very low permeability and increased in loamy soils with higher permeability, but with a further increase in permeability of the soils there appeared to be a decrease in rate of salt accumulation in the surface layers. These studies and results obtained by other workers (Krimgold, 1945; Ravikovitch and Bidner-Bar Hava, 1948) indicated that soils of intermediate texture were more liable to salt accumulation (Talsma, 1963) than either clay soils or sandy soils. Wiegand *et al.* (1966) in a study on salt affected dryland soils, attributed the larger net accumulation of salt in clay soils to faster upward flow of salts from water tables and lower rates of leaching in these soils, when compared to sandy soils.

In a field study on alluvial soil Sukhachev (1958) found that the critical depths for lucerne on stratified and uniform soils were 100 cm and 200 to 240 cm respectively, while for cotton the critical depths for stratified and non-stratified soils were 150 to 180 cm and 280 cm respectively.

An increase in concentration of salts in the groundwater would be expected to lead to a proportionate increase in rate of salt accumulation at the surface, provided that the rate of upward flow of solution was not affected by this change in salt concentration. Kovda (1961) found that when the salt concentration was 1.0 - 1.5% the critical depth to water table was generally around 200 to 250 cm, but if the salt concentration was 0.2-0.3% in groundwaters of the chloride-sulphate type and 0.07-0.1% in groundwater containing alkali (HCO_3^-) the extent of salinisation was generally very low. The capillary rise of different salt solutions in soil columns from a water table has also been studied (Krupkin, 1963; Szabolcs and Lestak, 1967). These studies showed that while 0.1 N solutions of NaCl , MgCl_2 and Na_2SO_4 rose to considerable heights, 0.1 N Na_2CO_3 solution rose only to a limited height above the water table. This was attributed to enhanced clay dispersion at the bottom of the column in the presence of the Na_2CO_3 solution. However, the specific effects of changes in salt solution composition and salt concentration of groundwater on the rate of upward flow and on salinisation of soils seem to have received little attention.

Field studies have shown that vegetation could affect the rate of salinisation from a water table, possibly by changing the moisture distribution pattern within the rooting zone. Verhoeven (1953), Jackson *et al.* (1956) and Van Schaik and Milne (1963) found that in sites with a deep water table greater salt accumulation occurred in

cropped than in fallow plots and this was believed to be caused by greater moisture withdrawal by crops from the surface layers.

Talsma (1963) and Sandoval and Benz (1973) showed that removal of vegetation in soils with a deep water table led to a decrease in the rate of salinisation apparently due to the formation of a dry surface mulch and also due to prevention of moisture withdrawal by crops from the rooting zone. The latter workers also found that in a site with a shallow water table, a soil in which a mulch was created by ploughing in the stubble was salinised almost to the same extent as a bare soil. This was attributed to the difficulty of maintaining a soil mulch by cultivation under the wet conditions and also to less efficient leaching due to the high water table. Hamilton (1972) found that establishing salt tolerant vegetation reduced the extent of salinisation of the surface 30 cm layer in soils with a saline water table, although the specific reasons for these reductions in rate of salinisation were not evaluated. Leo (1963) showed that root systems tended to accelerate salt water movement to the root zones, while fallowed soils tended to accumulate salt near the soil surface.

Studies in the field (Benz *et al.*, 1967; Wiegand *et al.*, 1968; Sandoval and Benz, 1973) have shown that mulches could reduce the rate of salinisation of soils from a saline water table by reducing the evaporation rate and by maintaining the soil in a moist condition so that even a moderate rainfall could leach the salts from the surface layers.

Salinisation of lands from saline water tables has been found to be a greater problem in drier areas, apparently due to higher evaporative conditions, lower annual leaching of salts and higher salinity of groundwaters. Jackson *et al.* (1956) showed in a field study that rainfall, while leaching salts from surface layers, also

raised the water table, as a result of which increased salinisation of surface layers occurred in the following dry months. Although it has been found that diurnal and seasonal temperature variations create temperature gradients in the soil profile which may induce some salt transport to the surface layers (Durand, 1956; Gurr et al., 1952) it has been suggested that this is not an important factor in salinisation of irrigated lands (de Vries, 1958). On the other hand Talsma (1963) found that in sites where the rate of upward flow was low, high atmospheric temperatures could cause excessive drying of the surface leading to the formation of a soil mulch, which then reduced the rate of upward flow from the saline water table.

Thus the early studies showed that the rate of salinisation of soils was affected by factors such as depth to water table, soil profile characteristics, salt concentration of the groundwater, climate and vegetation. However, the relative importance of these factors appeared to vary in different locations and the nature of their interaction was not clearly defined until Philip (1957) and Gardner (1958) proposed methods for computing upward flow in soils from a water table, based on an isothermal steady flow model. Their studies also showed that upward flow from a water table was limited either by the capacity of the soil to transmit water to the surface, governed by the depth to the water table and the hydraulic conductivity properties of the soil, or by the external evaporative conditions, whichever was less. Talsma (1963) in a study of salinity in soils of the Murrumbidgee Irrigation Area, in Australia, used a hydrological approach to compute the potential rate of upward flow of saline groundwaters in different soils. He combined these results with data for the maximum rates of evaporation possible under the climatic conditions existing during different seasons to predict

rates of upward flow in different soils throughout the year. Good agreement between these computed rates of upward flow and the experimentally determined rates of salt accumulation at the surface was obtained. Talsma also found that the critical depth corresponded to an upward flow rate of 0.1 cm/day. This rate being smaller than the potential evaporation rate of any of the seasons, was conducive to the formation of a soil mulch at the surface, thus restricting upward flow.

The principle processes and factors which affect the upward flow of water and salts from a water table and their interactions are illustrated in Figure 1. Field cultural and cropping practices could modify the rate of salinisation of soils due to upward flow from a water table by changing one or more of the factors which affect this upward flow. While many field trials have been carried out to assess the changes in the rate of salinisation of soils due to the use of different agronomic practices, the evaluation of the basic causes responsible for these changes usually requires investigations of a hydrological nature (Hamilton, 1972). Several workers (Visser, 1959; Talsma, 1963; Rijtema, 1965; Hamilton, 1974) have clearly shown that in studies relating to upward flow in soils from a water table, the use of the hydrological approach, taking into account the integrated effects of all factors which could affect this process (Figure 1), provides a means for quantitatively evaluating the specific effects of different factors which affect upward flow of water and salts and hence the rate of salinisation of soils from a water table.

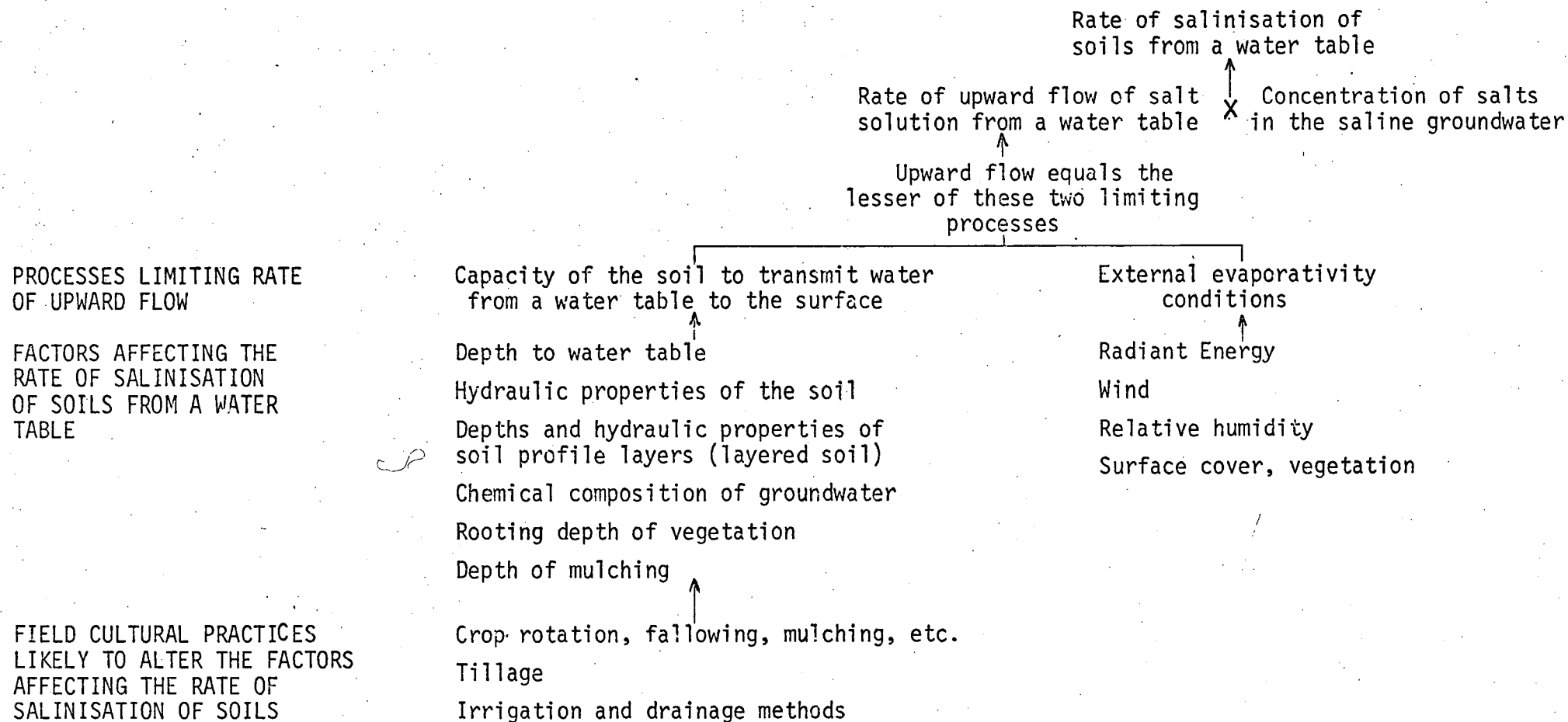


Fig. 1. The processes and factors affecting the rate of salinisation of soils from a saline water table, based on the isothermal, steady flow model.

B. Steady Upward Flow from a Water Table in Soils

King (1899) and Shaw and Smith (1927) found in studies in soil columns that as the depth to the water table was increased, the rate of upward flow decreased. The latter workers also found that there was no measurable upward flow of water in Yolo loam and Yolo sandy loam soil columns when the water table was at a depth exceeding 300 cm. Moore (1939) was the first to measure both the moisture content and pressure head distribution between the water table and the evaporating surface during steady upward flow.

The isothermal, steady, upward flow of water from a water table to the soil surface is given by,

$$q = -k(h) \left(\frac{dh}{dz} + 1 \right)$$

where q is the flux, $k(h)$ is the unsaturated conductivity, h is the pressure head and z is the depth to the water table, with the vertical coordinate taken as positive upwards. Separating the variables and integrating,

$$\begin{aligned} -z &= \int_0^h \frac{1}{1 + q/k(h)} dh \\ -z \text{ max.} &= \int_0^{-\infty} \frac{1}{1 + q/k(h)} dh \dots\dots\dots (1) \end{aligned}$$

It has been shown that if the relationship between the pressure head and unsaturated conductivity is known equation(1) can be solved either by numerical methods (Philip, 1957) or by analytical methods where unsaturated conductivity-pressure head curves can be represented by certain empirical equations (Wind, 1955; Gardner, 1958), to obtain pressure head-depth curves for different rates of upward flow. A number of different empirical equations have subsequently been proposed by different workers to represent experimentally determined unsaturated conductivity-pressure head curves.

The use of these equations has been reviewed in detail by Raats and Gardner (1971, 1974) and grouped into six basic models. The equations were expressed in dimensionless form in order to permit meaningful comparisons between the models. Further details of these models are given in Table 1. Another method for solving equation (1) using the flux/unsaturated conductivity ratio (Jayawardane, 1977) was developed during the present study (Appendix A).

Computed pressure head-depth curves indicate that suction shows a very gradual increase with height above the water table until just below the surface where it increases very rapidly (Gardner, 1960). When evaporation is limited by external conditions the suction at the soil surface is relatively low and only a small increase of this suction is required to markedly increase the evaporation rate. However, as the suction at the soil surface is further increased, the evaporation rate soon reaches a limiting value which cannot be exceeded however high the potential evaporation rate (Gardner, 1958, 1960). This maximum rate of evaporation is determined by the depth to water table and the hydraulic conductivity properties of the soil. Thus the rate of evaporation is limited by the external potential evaporation or the capacity of the soil to move water to the surface, whichever is less. The rate of evaporation is usually limited by the ability of the soil to move water to the surface, except under humid conditions or where a water table occurs at shallow depth.

Gardner (1958) also showed that for unsaturated conductivity-pressure head curves conforming to model E (Table 1), provided that the saturated conductivity/flux ratio is much greater than ten, the maximum rate of upward flow, q_{\max} , from a water table at a depth, z , is given by,

$$q_{\max} = A_n a z^{-n}$$

Table 1. Methods available for computing pressure head profiles during steady upward flow from a water table (based partly on Table 13-1 of Raats and Gardner [1974]).

Methods	Model	Author	Pressure head, h	Empirical equation for unsaturated conductivity	Advantages and Disadvantages
1. ANALYTICAL METHODS	A	Green and Ampt (1911)	$\geq h_1$ $< h_1$	K 0	1. Simplifies computations of upward flow and provides insights into physical aspects of flow. 2. Can be used only for k-h curves which fit one of the many 2 or 3 parameter equations proposed.
	B	Richards (1931)	$\geq h_2$ $< h_2$	$K(1-h/h_2)$ 0	
	C	Gardner (1958)	≤ 0	$K \exp \alpha h$	
	D	Rijtema (1965)	$\geq h_3$ $< h_3$	K $K \exp \beta(h-h_3)$	
	E	Gardner (1958)	≤ 0	$a(S^n+b)^{-1}$	
		Raats and Gardner (1971)	≤ 0	$K[(h/h_{1/2K})^n+1]^{-1}$	
	AE	Whisler (1969)			
	F	Wind (1955)	$\geq h_4$ $< h_4$	K $K(h/h_4)^{-m}$	
		Brooks and Corey (1964)			
2. NUMERICAL METHOD		Philip (1957)			1. Can be used to compute upward flow accurately from any complex k-h curve. 2. Does not provide physical insights into flow.
3. FLUX/UNSATURATED CONDUCTIVITY RATIO METHOD		Appendix A			

where A_n , a and n are constants for a given soil. Values of n of about 2 have been obtained for fine textured soils while coarse textured soils have values of 4 or greater. Cisler (1969) showed that the value of A_n in the above equation was given by $(n/\pi \sin \pi/n)^{-n}$ for both integer and non-integer values of n . In the case of all models in Table 1, except A and B, for any depth of water table, there is a maximum upward flux which approaches zero asymptotically as the depth to the water table becomes greater (Raats and Gardner, 1974). It has been pointed out that these small fluxes may represent an important salinity hazard (Gardner, 1960).

Philip (1957) and Gardner (1958) also extended their analysis to consider the contribution of vapour movement to the rate of evaporation under steady flow conditions. Where the external evaporative demand is greater than the maximum upward flow the soil near the surface dries and within this zone water movement is entirely in the vapour phase while below this zone water movement is in the liquid phase. Gardner (1958) also showed that although movement in the vapour phase would tend to increase the rate of evaporation, the contribution of vapour movement to evaporation would be relatively small and is usually much less than 20 per cent.

A number of studies in the field (Wind, 1955; Visser, 1959; Talsma, 1963) and in soil columns (Gardner and Fireman, 1958; Willis, 1960) have indicated that the theoretical computations of upward flow from water tables based on the isothermal steady flow model show close agreement with the experimentally determined evaporation rates in their soils. However, Schleusener and Corey (1959) and King and Schleusener (1961) found that the evaporation rate decreased as the external evaporative potential was increased, instead of remaining constant after attainment of the limiting rate as predicted by this theory. They suggested that this was due to hysteretic drying of

the surface layers. Hadas and Hillel (1968) also obtained similar results in evaporation experiments but since the decrease in evaporation rate under high external evaporative conditions occurred even though the soil surface was drying monotonically, they suggested that this decrease in evaporation rate could not be attributed to hysteresis. Further, as these decreases occurred under very high external evaporative conditions, (which were much higher than those normally existing under field conditions), it is possible that a downward heat flux (Philip, 1957) could have caused reductions in the evaporation rate. Philip (1957) showed that while temperature gradients had little effect on evaporation from moist soils, these gradients could affect the rate of evaporation as the soils became very dry. Gardner (1958) showed that the presence of a surface mulch would lead to a decrease in the rate of evaporation in soils with a water table and that this decrease was proportional to the depth of the mulch.

The hydrology of highly colloidal soils, which show changes in their bulk volumes during wetting and drying, has been described by Philip (1969a, b). In such soils the total potential includes a component of overburden potential in addition to the moisture and gravitational potentials.

Willis (1960) computed the steady upward flow in a layered soil made up of two layers of different textures by means of an analytical method, using the appropriate conductivity-pressure head function for each layer and matching the pressure head at the boundary of the two layers. He found that the computed rates of upward flow increased as the depth of a coarser sand

layer over a loam layer increased, while increasing the depth of the loam layer over the sand decreased the upward flow. On the contrary, it has been shown by other workers (Hadas and Hillel, 1972; Varallyay, 1974) that increasing the depth of the finer textured layer at the surface increased the rate of upward flow while the increase in depth of the coarser textured layer at the surface decreased the rate of upward flow. Willis (1960) and Hadas and Hillel (1972) found the experimentally determined rates of evaporation in layered soil columns were less than the computed values and it was suggested by the latter authors that this could be due to interlayer resistance in these artificially packed soil columns. Hamilton (1974) found good agreement between computed and measured rates of upward flow in a layered field soil. The large differences in capacities of different soils to transmit water to the surface from water tables at specific depths is due to wide differences in the unsaturated conductivity-pressure head curves of the soils or of the soil layers in stratified soils.

If the presence of different salt solutions causes changes in the hydraulic conductivity properties of a given soil, this could be expected to lead to changes in the ability of the soil to transmit solutions from the water table to the surface. The magnitude of these changes in upward flow would depend on the relative changes in the conductivity properties of the soil in the presence of the different salt solutions.

C. Movement of Salts in Soils

The two main processes causing salt movement in soils are convection of dissolved salts with the mass flow of soil solution and molecular diffusion due to concentration gradients (Gardner, 1965; Peck, 1971; Boast, 1973).

Movement of salt by molecular diffusion takes place from zones of higher salt concentration towards zones of lower salt concentration. The rate of diffusion is approximately proportional to the difference in concentration. However, molecular diffusion in soils is a very slow process. For instance, Peck (1971) showed that molecular diffusion would move more salt than would convection only where the soil solution flow was less than 10^{-9} m/sec.

During salt movement by convection, the macroscopic rate of movement of salt is equal to the product of the macroscopic rate of movement of soil water and the concentration of salt in that soil water. However, the bulk movement of the soil solution is made up of many different microscopic rates of flow. Thus there is a faster rate of flow in the larger pores and also a faster rate of flow in the centre of individual pores than near the pore walls which results in a dispersion of salt in the soil. This process is referred to as hydrodynamic dispersion. In soils where there is a net movement of soil solution, convection could interact with hydrodynamic dispersion to enhance the diffusion process (Gardner, 1965). Other processes of lesser importance affecting salt movement in soils include salt seiving (Kemper, 1960; Krupp et al., 1972), absorption into the exchange complex resulting in less rapid transport of salts subject to such sorption (Biggar and Nielsen, 1963) and storage in dead end pores (Philip, 1968).

In a field study Doering et al. (1964) estimated downward movement of salt near an evaporating surface at between 15 to 40 per cent of the upward movement of salt. Gardner (1965) in reviewing the data from earlier studies concluded that downward diffusion was not sufficiently rapid to carry nitrate more than a few inches below the point where the evaporation was occurring. A number of studies of the extent of salt accumulation near an evaporating surface in conjunction with upward movement of the soil solution have found good correlation between calculated rates of salt accumulation in the surface layers, assuming convection is the main factor and ignoring the effects of molecular diffusion, and measured rates of salt accumulation (Richards et al., 1956; Talsma, 1963).

During the movement of salt solutions through soils, changes could occur in both chemical composition and salt concentration. As soil solution evaporates at or near the soil surface salt accumulation leads to concentration of soil solution near the evaporating zone until the solubility limits of the salts are reached. Varallyay (1968) showed diagrammatically the changes in composition of groundwater as it moved from the water table to the soil surface in the Danube Valley area. The concentration of soil solution as it moved to the surface resulted in a precipitation of salts in the reverse order of their solubility. Hence the maximum concentration of CaCO_3 , MgCO_3 , NaHCO_3 and NaCl were found respectively at decreasing depths toward the surface. The concentration of the soil solution due to evaporation near the soil surface results in an increase in SAR value of this soil solution, in proportion to the square root of this concentration.

In a field study on salinisation of a glacial till soil by upward flow from a saline water table, van Schaik and Milne. (1963) found that the increase in salt concentration at the surface was accompanied by a marked increase in SAR value and he also showed that this was partly due to a precipitation of calcium near the surface.

D. Salt Solution Composition Parameters

The changes in hydraulic conductivity of certain soils in the presence of dissolved salt may be caused by swelling and dispersion of clays. When sodium ions are present in the soil solution they may be absorbed within the exchange complex of the soil. Further if the salt concentration of this solution is sufficiently low, this could lead to enhanced swelling and dispersion of the clay, resulting in decreases in hydraulic conductivity. While several parameters have been suggested to characterise the composition of salt solutions in relation to their effect on the hydraulic conductivities of soils, the cation concentration expressed in meq/l and the sodium absorption ratio of the solution have been widely used in recent studies (McNeal and Coleman, 1966; Thomas and Yaron, 1968).

The sodium absorption ratio (SAR) is defined as,

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

where all concentrations are expressed in meq/l. The use of this ratio was first proposed on an empirical basis (U.S. Salinity Laboratory staff, 1954) from experimental results for several soils in the USA, which showed a relationship between the sodium absorption ratio and the exchangeable sodium percentage.

A theoretical basis for the use of this ratio was provided by Babcock (1963).

E. Effect of Salt Solutions on Swelling of Clays in Soils

Swelling and dispersion of clays have been recognized as the major factors causing the decrease in hydraulic conductivity of soils to salt solutions. The initial swelling of dry clays which occurs in a stepwise manner until an inter-particle spacing of approximately 19 Å is reached, is caused by absorption of hydration shells by exchangeable cations. This process is essentially completed at soil suctions of 27 bars (Quirk, 1953; Emerson, 1963). Further swelling of clays is attributed to formation of diffuse double layers, and this process has been mathematically described by Gouy (1910). This theory has been further developed and used to describe swelling of clays (Schofield, 1946; Bolt, 1956). The diffuse double layer theory has also been applied to clays with mixed cations (de Haan, 1964; Collis-George and Bozeman, 1970).

In X-ray studies of crystalline swelling, Norrish (1954) found that sodium montmorillonite in 0.3 M sodium chloride showed extensive swelling and an interlayer spacing of more than 43 Å. He also showed that calcium montmorillonite had a fixed basal spacing of 19 Å in solutions ranging from 1.7 M calcium chloride to distilled water, Na-vermiculite had an interlayer spacing of 14.8 Å even in dilute solutions, and muscovite crystals did not swell at all.

Electron-microscopic studies (Aylmore and Quirk, 1962) showed that in soils clay crystals existed in regions of near parallel arrangement referred to as domains which were several microns in thickness. Hence clay swelling in many soils could also be caused by interaction between the crystals within the domains and such intercrystalline swelling in sodium saturated clays has also been

shown to be sensitive to both electrolyte concentration and applied suction (Aylmore and Quirk, 1962). While the swelling of domains is generally unidirectional, due to the random orientation and distribution of domains in the soil (Aylmore and Quirk, 1960) the swelling of clays and soil aggregates is largely isotropic (Fox, 1964).

Several workers have attempted to verify theoretical predictions of the swelling of clays according to double layer theory under varying conditions of electrolyte concentration, type of exchangeable cations and applied suction.

Satisfactory agreement between theoretical and experimental swelling pressure versus film thickness curves was obtained by Bolt (1956) and Warkentin *et al.* (1957) for sodium montmorillonite in 10^{-3} M to 10^{-4} M sodium chloride solutions, provided prior compression of twenty atmospheres was applied to force the clay into parallel orientation and reduce the void space between the crystals. Warkentin *et al.* (1957) also found that with pre-compressed calcium montmorillonite the swelling pressure versus film thickness curves were displaced to lesser film thicknesses than expected from the theory, assuming a clay surface area of $800 \text{ m}^2 \text{ g}^{-1}$. Several workers (Aylmore and Quirk, 1959, 1960; Blackmore and Miller, 1961) have suggested that calcium montmorillonite formed tactoids or quasi-crystals and hence could show only limited intra-crystalline swelling. Blackmore and Miller (1961) found from X-ray studies that as the constraint was increased, the number of clay particles per tactoid also increased.

The swelling of sodium Willalooka illite, which shows only inter-crystalline swelling, was found by Aylmore and Quirk (1962) to be sensitive to both electrolyte concentration and pressure and hence was very similar to the intra-crystalline swelling of montmorillonite. Aylmore and Quirk (1966) found that the swelling of

kaolinite clays saturated with sodium or with calcium was almost identical and neither system exhibited any marked sensitivity to electrolyte concentration.

These studies have shown clearly that the amount of swelling is highly dependent on the previous compression or dehydration history of the sample (Warkentin *et al.*, 1957; Aylmore and Quirk, 1959), the nature of the clay and the exchangeable cations present. It has also been shown that in calculating water absorption by clays using the diffuse double layer theory, allowance has to be made for water present in the gel structure (Schofield, 1946; Aylmore and Quirk, 1962; Quirk, 1968). In addition, the constraints to swelling of clays in soils due to the presence of mechanical and chemical bonding needs to be considered.

Two models have been proposed to describe the distribution of sodium and calcium ions in mixed-ion clays. According to the mixed-ion model used by some workers (de Haan, 1964) each species of cation is distributed over all mineral surfaces in proportion to its exchangeable cation percentage. Other workers (McNeal *et al.*, 1966; Shainberg *et al.*, 1971) favour the demixed-ion model, according to which in the case of sodium-calcium clays, a demixing of ions into calcium regions and sodium regions could occur with only the sodium regions exhibiting concentration dependent interlayer swelling. McNeal (1970) showed that most of the limited data on swelling of clays in mixed-ion solutions agreed more closely with theoretical predictions from the demixed-ion model than with the mixed-ion model.

McNeal *et al.* (1966) studied the effects of salt solutions on the macroscopic swelling of clays. They found that the macroscopic swelling of clays obtained from soils containing expanding

minerals, as the electrolyte concentration was reduced in calcium chloride solution, was small. But in the presence of mixed sodium-calcium solutions, the electrolyte concentration at which the clays show specific extents of swelling increased as the SARs of these solutions were increased. Similar results were obtained by Rowell *et al.* (1969) in studies on the effect of salt solutions on swelling of oriented clay aggregates of a brown earth soil. McNeal *et al.* (1966) also found that the measured macroscopic swelling of clay pads showed a close correlation with the interlayer swelling values calculated from the demixed-ion model after accounting for the solution taken up by the gel structure of the soil. The agreement between the theoretical and experimental swelling values was further improved by assuming that the first increment of sodium was ineffective in causing swelling. Shainberg and Otoh (1968) suggested that the first increments of sodium added to a calcium clay were largely concentrated on the external surfaces of the clay crystals and hence did not affect the basic structure of the clay tactoids. Further additions of sodium into the exchange complex (over 20 per cent) resulted in the breakdown of the tactoids.

F. Effect of Salt Solutions on Hydraulic Conductivity of Soils

The decrease in hydraulic conductivity and deterioration of structure in the presence of sodium in the percolating solution or on the exchange complex of soils, and the effect of calcium in increasing the hydraulic conductivity (Fireman and Magistad, 1945; U.S. Salinity Laboratory Staff, 1954) was recognized in early studies. Fireman and Bodman (1939) and Bodman and Fireman (1950) showed that satisfactory permeability can be maintained in a soil which is 30 per cent saturated with sodium, by using a solution with high

salt concentration. Emerson (1954) defined the concentration at which the hydraulic conductivity of sodium saturated soils dropped essentially to zero as the critical concentration, and used this value as an index of soil structure.

Quirk and Schofield (1955) introduced the concept of threshold concentration, defined as the minimum concentration of electrolyte necessary in the soil solution to prevent a decrease in permeability of more than 10 - 15 per cent. In a detailed study on Sawyers soil at Rothamsted Experimental Station, they found that the threshold concentration for this soil when saturated with sodium was higher than when saturated with calcium. When the soil was saturated to varying degrees with sodium the threshold concentration increased with sodium saturation. A method for classifying irrigation waters into permeability classes was suggested by Doneen (1961) on the basis of his work on a high montmorillonite Yolo soil whose permeability was reduced by 25 per cent when the electrolyte concentration was less than 3.5, 6.5 and 10.0 meq/l at ESP values of 0.6, 3.4 and 8.0 respectively.

McNeal and Coleman (1966) studied the effect of salt solutions on several soils with different clay mineralogy. They found that with a reduction in electrolyte concentration or increase in SAR, the decrease in hydraulic conductivity was most pronounced for soils high in 2 : 1 layer silicates, while soils high in kaolinite and sesquioxides were virtually unaffected by changes in salt composition. The most labile hydraulic conductivities were exhibited by soils containing the most montmorillonite. These studies also showed that the chemical properties of the solution alone were not sufficient to characterise the relative hydraulic conductivities of different soils in the presence of different salt solutions, especially where

the clay mineralogy of the soils show marked differences.

Yaron and Thomas (1968) found that in three soils whose respective mineralogy was predominantly montmorillonite, montmorillonite-illite and illite-kaolin, appreciable reduction in hydraulic conductivity for solutions where electrolyte concentration was 34.5 meq/l occurred when the ESP values exceeded 10, 27 and 21 respectively, while in solutions of electrolyte concentration of 11.3 meq/l the marked decreases in hydraulic conductivity occurred at ESP values of 8, 13 and 18 respectively. In a study on vermiculitic soils, Rhoades and Ingvalson (1969) showed that a 25 per cent decrease in hydraulic conductivity to solutions with electrolyte concentration between 5 - 20 meq/l did not occur at ESP values less than 50. This was attributed to these soils having an appreciable fraction of their exchangeable sodium sorbed on relatively large, semi-expanding vermiculite particles, this fraction of exchangeable sodium being inactive in its effects on soil permeability. Naghshineh-pour *et al.* (1970) found that while the increase in SAR and decrease in electrolyte concentration led to rapid decrease in conductivity of two montmorillonitic soils, a soil with a clay fraction primarily composed of tabular halloysite did not show appreciable decreases in hydraulic conductivity except in solutions of the highest SAR (SAR 38) and lowest salt concentration (10.5 meq/l). The hydraulic conductivity of another soil containing tabular halloysite and also having a high iron oxide content remained virtually unchanged in all salt solutions. Studies carried out by El-Swaify and Swindale (1968) showed that in soils high in kaolin and iron oxide, the hydraulic conductivity did not change much as the electrolyte concentration of solutions was decreased and the sodium content was increased.

Factors which contribute to the stabilisation of soil structure, such as organic matter and sesquioxides (Russell, 1971) could affect the stability and conductivity characteristics of soils to salt solutions. Emerson (1954) and Dettmann and Emerson (1959) found that the permeability to high sodium, low salt solutions was much higher in a soil which had supported a permanent pasture for several years than in soils under arable cropping, apparently due to the effectiveness of organic matter as a cementing agent. Several workers (El-Swaify and Swindale, 1968; McNeal *et al.*, 1968; Naghshineh-pour *et al.*, 1970) found that in soils with high contents of sesquioxides, hydraulic conductivity changed only slightly in high sodium, low salt solutions. McNeal *et al.* (1968) also showed that the partial removal of the free iron-oxides led to a decrease in stability of these sesquioxide rich soils, and a reduction in conductivity to high sodium, low salt solutions occurred. El Rayah and Rowell (1973) showed that a deposit of sesquioxides around a soil aggregate reduced swelling and increased permeability to sodium chloride solutions of low salt concentration until the coating was broken by the swelling pressure of the clay in the aggregates. Aluminium oxides have been shown to be more important than iron oxides in increasing the structural stability of soils (Deshpande *et al.*, 1968) especially in soils in which the iron oxides exist as discrete nodules and hence do not contribute to the stability of these soils (Greenland *et al.*, 1968). Thus in addition to the amounts of these cementing agents present, their nature and distribution are important factors in increasing the stability of soils to salt solutions.

The above studies show that, in general, in soils containing chiefly montmorillonitic clays, the hydraulic conductivity was markedly decreased in high sodium, low salt solutions, while in

soils with predominantly kaolinitic clays the hydraulic conductivity was only slightly affected. It was also seen that the presence of sesquioxides and organic matter, depending on their nature and distribution, could increase the stability of soils to salt solutions.

The effects of other soil properties on the hydraulic conductivity of soils in low salt, high sodium solutions have also been studied. McNeal et al. (1968) found that in soils having similar clay mineralogy the decrease in hydraulic conductivity with reduction of salt concentration at a given SAR was more pronounced for soils with higher clay content.

The type of cations present in salt solutions also affects the saturated conductivity of soils. Quirk and Schofield (1955) showed that the concentration of electrolyte causing a 10 to 15 per cent decrease in hydraulic conductivity increased as the cation saturating the exchange complex was changed progressively from calcium to magnesium to potassium and then to sodium. McNeal et al. (1968) found that the hydraulic conductivity was lower for sodium-magnesium solutions than for sodium-calcium solutions of comparable SAR and salt concentration values.

Martin and Richards (1959) showed that in soils packed into cylinders when moist, very little changes in hydraulic conductivity occurred as the exchangeable hydrogen content and exchangeable calcium content was markedly increased, but increasing the exchangeable sodium, potassium and ammonium resulted in marked decreases in conductivity. Increase in exchangeable hydrogen, however, tended to increase the dispersive action of sodium, potassium and ammonium. Ammonium caused a greater decrease in hydraulic conductivity in an acid soil, than in a base-saturated soil. In further studies,

Martin et al. (1964) showed that the same amount of sodium caused a greater reduction in hydraulic conductivity in acid than in neutral or alkaline soils, and this was attributed to the decreases in cation exchange capacity with increasing soil acidity. Hence the same amount of sodium in an acid soil represents a higher exchangeable sodium percentage than in a neutral or alkaline soil.

The permeability to water of Shepparton soil columns saturated with calcium and magnesium showed only a negligible decrease with time (Bakker and Emerson, 1973). With sodium present, the ESP required to produce a given decrease in flow rate was reduced when magnesium was the complementary cation rather than calcium.

When a soil that is initially non-sodic is irrigated with sodic water the hydraulic conductivity decreases until the soil profile reaches equilibrium with the permeating solution (Yaron and Thomas, 1968). This decrease in hydraulic conductivity was characterised by a curve whose shape varied with the soil and the salt composition and concentration of the permeating solution. Thomas and Yaron (1968) found that for a given salt concentration of irrigation water, an increase in SAR resulted in a greater decrease in hydraulic conductivity of the soil.

The decrease in conductivity of some soils for salt solutions has been attributed by some workers (Burgess, 1928; Gardner, 1945; Bodman and Fireman, 1950) largely to clay dispersion causing a blocking of conducting pores. Other workers (Carman, 1939; Smith and Stallman, 1955; Koenigs, 1961) found that clay swelling in salt solutions was the main cause of the decrease in conductivity. It has also been proposed that decreases in hydraulic conductivity of most soils in salt solutions occur in two stages, with swelling causing a weakening of bonds between clay particles, followed by particle

dispersion and translocation (McNeal and Coleman, 1966; Rowell et al., 1969; Quirk, 1971). McNeal et al. (1966) found close correlation between hydraulic conductivity decreases in low salt, high sodium solutions and predicted and experimental swelling of clays in the same salt solutions. They found that limited reversibility of clay swelling in soils made reversibility of permeability decreases a doubtful index of the type of permeability decreases that had occurred. Rowell et al. (1969) found a close correlation between cation concentration at which marked decreases in hydraulic conductivity occurred and the cation concentration at which marked swelling of clay pads took place. In the same soil, turbidity of percolating solutions during conductivity experiments occurred at lower cation concentrations. However, earlier studies (Bodman and Harradine, 1938) have shown that clay dispersion could occur within the soil column even though a clear percolate flowed from the bottom of the column.

Shainberg et al. (1971) and Shainberg and Caiserman (1971) found that addition of a small amount of exchangeable sodium to a clay saturated with calcium, had only a moderate effect on swelling and did not decrease the hydraulic conductivity of this clay to water, under conditions where the clay was not free to move. But where the clay was free to move, there were marked decreases in hydraulic conductivity on addition of small amounts of exchangeable sodium. They suggested that when about 10 per cent of the exchangeable calcium was replaced by sodium, most of the adsorbed sodium was on external surfaces of the tactoids and hence there was no additional swelling of the clay. But this addition of sodium led to increases in the electrophoretic mobility of the clay (Bar-On et al., 1970) and where the clay was free to move, hydraulic conductivity decreases could occur.

In scanning electron microscope observations on columns of a sandy soil and a clayey soil, Chen and Banin (1975) found that clay dispersion was responsible for hydraulic conductivity decreases in these soils at low ESP from 0 to 20 per cent.

Although the relative importance of clay swelling and clay dispersion in causing decreases in hydraulic conductivity varies in different soils, both these processes appear to be enhanced by high sodium, low salt solutions.

Rowell et al. (1969) showed that the cation concentration at which clay dispersion occurred when the soil was at a given ESP, was increased as the mechanical stress applied to the soil was increased. These results also emphasise the need to regulate the extent of mechanical disturbance to which the soil is subjected during wetting, leaching with specific salt solutions and also during conductivity measurements, in relation to the specific field conditions to which these results are applied.

One of the problems in reviewing and comparing the results of different workers on the effects of salt solutions on hydraulic conductivity of soils is the use of different soil fractions in these studies. The use of the entire fraction below 2mm, which is a common practice, presents difficulties as these fractions are likely to contain a fair amount of dusty clay material. The presence of this fine material could lead to greater dispersion in high sodium, low salt solutions than if a soil core or a packed soil column of a size fraction which excludes the finer size fractions were used. Further, since the dusty material is largely an artefact of the process of soil grinding and the amount present also varies with the type of grinding and the soil used, this could be responsible for some of the variations in the results of different workers.

Recent developments in applying double layer theory to swelling of mixed-ion clays (de Haan, 1964; Collis-George and Bozeman, 1970) could encourage the search for theoretical methods for predicting hydraulic conductivities of soils to different salt solutions from soil characteristics. However, the complexity of the factors determining the swelling of clays in soils in the presence of salt solutions and their resultant effects on the pore geometry and hydraulic conductivity, as illustrated in Figure 2, present practical difficulties in making such predictions. For instance, the method proposed by Lagerwerff *et al.* (1969) for predicting hydraulic conductivity changes in the presence of salt solutions, based on the double layer theory and Kozeny-Carman equation, appears to have limitations in use (McNeal, 1974). However, several semi-empirical methods have been proposed. McNeal (1968) devised a method for predicting hydraulic conductivity for salt solutions using the measured conductivity of the soil in a high salt, high sodium solution and in a solution of a lower salt concentration. Calculated interlayer swelling values of montmorillonite were used as a frame of reference for these predictions. Yaron and Thomas (1968) proposed a semi-empirical method for predicting the hydraulic conductivity decreases expected from the use of sodic waters. Rowell *et al.* (1969) predicted hydraulic conductivity changes in salt solutions from changes in porosity of clay pads due to swelling of clay in the same solutions. Although satisfactory agreement was found between the saturated conductivities predicted from these semi-empirical methods and the measured saturated conductivities, further testing of these methods for a wider range of soils is necessary.

SATURATED AND UNSATURATED HYDRAULIC CONDUCTIVITY OF SOILS

- 1. Pore geometry of soils in salt solutions
- 2. Relation between pore geometry and hydraulic conductivity

PORE GEOMETRY OF SOILS IN SALT SOLUTIONS

- 1. Initial pore geometry of the dry soil
- 2. Swelling of aggregates causing changes in pore geometry

SWELLING OF SOIL AGGREGATES

- 1. Swelling of clay matrix
- 2. Skeletal grain structure of sand and silt fraction
(and the distribution and orientation of clay within the soil fabric)
- 3. Cementing agents such as organic matter, sesquioxides, etc. (nature, amounts and distribution in the soil)

SWELLING OF CLAY MATRIX

- 1. Crystalline swelling of clay
- 2. Amounts of different clay minerals
- 3. Formation of domains and quasicrystals
- 4. Formation of void or gel structures

CRYSTALLINE SWELLING OF CLAYS

- 1. Type of clay and its crystal structure
- 2. Exchange cations
- 3. Electrolyte concentration of solution
- 4. Suction or confining pressure

Fig. 2. A generalized scheme for interaction of factors affecting the swelling of clays and soil aggregates and their effects on saturated and unsaturated hydraulic conductivities of soils.

The effect of salt solutions on unsaturated conductivity and diffusivity has also been studied. Gardner et al. (1959) showed that with increase in ESP and decrease in electrolyte concentration of salt solutions there was a decrease in mean-diffusivity of Pachappa sandy loam and Yolo loam soils. Thus when the ESP was above 25, as the electrolyte concentration was reduced from 300 to 3 meq/l, the mean-diffusivity decreased one thousand-fold. They also found that although the reduction in diffusivity for high sodium, low salt solutions occurred mainly at higher water contents, significant reductions also occurred at lower water contents corresponding to suctions of 15 bars.

Christenson and Ferguson (1966) found that an artificial soil containing 17.6 per cent dickite had a higher diffusivity for 0.015 N CaCl_2 than for distilled water at all water contents when the soil had an ESP of 5.0. When the soil had an ESP of 20.3, the diffusivity for 0.015 N CaCl_2 solution was higher only at high water contents. An artificial soil with 17.6 per cent montmorillonite showed a higher diffusivity in 0.015 N CaCl_2 than in distilled water only at high water contents when the soil had an ESP of 5.5. In this soil with ESP increased to 13.6, the diffusivity for water and CaCl_2 solutions was similar. These results were explained in terms of changes in diffuse double layers caused by the presence of CaCl_2 solution. Kijne (1967) showed that the treatment of a clay loam soil with Krilium and a polyvinyl alcohol caused an increase in diffusivity over the whole range of water contents while the diffusivity of a loamy sand soil showed only small changes when treated with these soil conditioners.

Three soils with ESP values between 5 and 20 were found by Sharma (1972) to have greater diffusivity and unsaturated conductivity at high water contents in the presence of water

treated with gypsum than in the presence of untreated water. However, at lower water contents the difference in diffusivity of soils to gypsum treated and untreated water was small.

Thus previous studies show that the saturated and unsaturated conductivities of some soils are altered, to varying extents, by the presence of certain salt solutions. Therefore, in such soils it could be expected that the upward flow from water tables could be changed according to the chemical composition of groundwaters and the magnitudes of such changes are likely to vary in different soils. The present study investigates the changes in rates of upward flow in different soils, as the chemical composition of the groundwater is altered and relates these changes to changes in conductivity properties and pore geometry of the soils in different salt solutions.

III. MATERIALS AND EXPERIMENTAL METHODS

A. Soil Materials

Three soils located in north and south-eastern regions of Tasmania, Australia, were selected for this study. Soil samples were collected from the subsoils of:

(a) a krasnozem soil formed on basalt near Burnie in north-western Tasmania, mapped as Burnie clay loam (Loveday and Farquhar, 1958). The soil material taken from a depth of 20 to 32 inches had a reddish brown (2.5YR 4/6) colour and a field grading of clay. Pedality was strong medium angular blocky and moist consistency was very friable.

(b) an alluvial soil on a fairly broad alluvial plain near Sorell in south-eastern Tasmania, mapped as such by Loveday (1957). The soil material taken from 24 to 36 inches depth had a brownish black (10YR 3/2) colour and a field grading of sandy clay loam. Pedality was moderate medium sub-angular blocky.

(c) a red brown soil formed on basalt near Sorell in south-eastern Tasmania. This soil was mapped as Stoneleigh clay loam (Loveday, 1957). The soil material, taken from a depth of 22 to 34 inches, had a dark reddish brown (5YR 3/3) colour and a field grading of clay loam. Pedality was moderate medium sub-angular blocky.

B. Preparation of Soils and Salt Solutions

The soils were air dried, thoroughly mixed and divided into two equal portions. One portion was stored. The remaining portion was gently broken by hand, crushed and sieved using a rotating crusher and sieve, to obtain a fraction between 2.0 and 0.2 mm in diameter. This fraction consisted largely of small aggregates and contained very little dusty material. This sieved soil

material was again thoroughly mixed and subsampled by a process of mixing, quartering, remixing opposite quarters, quartering and so on to obtain eight subsamples of each soil which were stored for future use. When taking soil from these subsamples for use in the experiments, the same process of mixing and quartering was employed, to ensure as far as possible uniformity of the soil material used in the experiments.

Further, the procedures of packing, wetting, draining and the experimental methods were standardized and the exact procedure was followed always in order to minimise the possible variation in pore geometry of the packed soil samples due to differences in experimental handling of the soils (Loveday, 1974). All experiments were conducted in a constant temperature room at $22^{\circ} \pm 1^{\circ} \text{C}$.

The solutions of specific salt composition and concentration used in the experiments were prepared by dissolving the required quantities of reagent grade sodium chloride and calcium chloride in deionized, deaerated water (Appendix C). The solutions also contained 20 ppm of mercuric chloride to suppress any bacterial activity.

C. Physical and Chemical Characteristics of the Soil Materials

The physical and chemical characteristics of the soils were determined using the procedures described in "Methods of soil analysis" (Black et al., 1965): Determinations of particle size analysis by the pipette method, organic matter content by the Walkley and Black method, free iron oxide content by the diphenylamine sulphonate method, soil reaction using a glass electrode, cation exchange capacity by the sodium saturation method, and total exchangeable bases were carried out.

The particle density of the soils was measured using a pycnometer and the moisture characteristic at low energy was determined by keeping fragmented soil samples in a vacuum desiccator containing saturated salt solutions of known relative humidity. The clay fractions extracted from the soils were used to determine the minimum cation concentrations in solutions of specific SARs in which the clay flocculated, using the method described by Collis-George and Smiles (1963).

D. Moisture Release Curves of Soils for Salt Solutions

The moisture release curves of the soils in the presence of salt solutions of different composition were determined using a modified Haines apparatus (Figure 3). The apparatus consisted of a sintered glass buchner funnel with a porous plate (porosity 3) of 5 cm diameter, on which the soil was packed to a depth of 2 cm. A wire mesh positioned on the top of the packed soil served as a confining plate (Bridge *et al.*, 1970). This was designed to prevent changes in bulk density of the soil due to the swelling of the soil aggregates at the surface and also to minimise the disturbance of the soil column during leaching. The confining plate was supported and held in place by gluing it to the end of an inverted funnel, whose stem passed through and was glued to a large rubber stopper covering the sintered funnel. A thick walled rubber tube connected the sintered funnel to a 50 ml burette. A side tap T, opened only during the leaching of the soil, was provided at the base of the burette. A flexible tube from the stem of the inverted funnel to the top of the burette prevented loss of water through evaporation and also permitted equilibration of the air pressure in the sintered funnel and the burette.

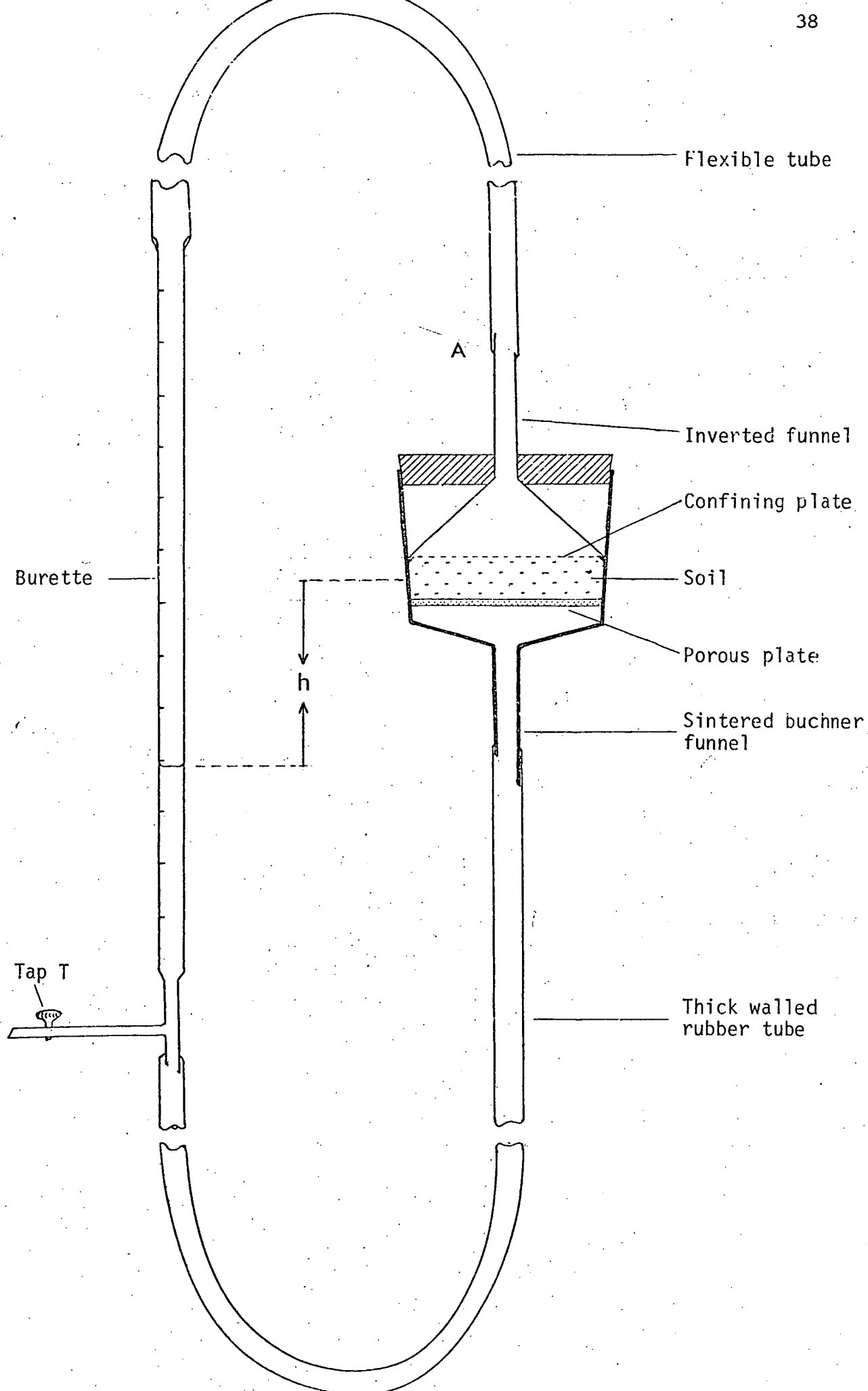


Fig. 3 . Modified Haines apparatus for measuring the moisture release curves of soils in salt solutions.

The sintered funnel, rubber tube and the burette were filled with a solution of SAR 40 and highest cation concentration of 640 meq/l, through the pre-saturated sintered plate. With the plate at a suction of 60 cm, the 2.0 to 0.5 mm fractions of the krasnozem, alluvial and red brown soils were packed on the plate to bulk densities of 0.87 ± 0.02 , 1.01 ± 0.02 and 1.05 ± 0.02 g/cm³ respectively.

The soil was wetted for a few hours under suctions of 60 and 30 cm of water and then saturated by raising the level of the solution in the burette. The soils were twice drained to 60 cm suction and reflooded to allow the soil to pack. The soil was reflooded and the same salt solution was siphoned into the buchner funnel from a mariotte bottle through connection A. With tap T open, the saturated soil was leached with approximately 20 pore volumes of this salt solution (McNeal and Coleman, 1966) to ensure chemical equilibrium between the soil and the leaching solution. All tube connections were sealed with a rubber sealant. The moisture release curve was determined by increasing the suction in small increments and recording the volume of outflow of the solution into the burette and the final equilibrium suction. A small predetermined correction for plate drainage was applied to these readings.

The soil was again saturated and then leached with the solution of the same SAR and next highest cation concentration and the moisture release curve was determined using the same procedure. This process was continued until the moisture release curve of the soil in equilibrium with the salt solution at SAR 40 and the lowest cation concentration was determined. The soil was then sampled to determine the moisture content at the highest suction applied.

The experiment was then repeated with a new soil column for each of the other series of salt solutions at SAR values of 20, 10 and 0. The soil was first leached with the solution of highest cation concentration at each of these SARs and then with progressively lower cation concentrations, and the moisture release curves were determined as described.

Moisture release curves were also determined for soil columns of 2.0 to 0.2 mm fractions of the alluvial and red brown soils packed to higher bulk densities of 1.13 and 1.20 g/cm³ respectively for the same series of salt solutions. These curves were measured over a pressure head range of 0 to -160 cm of water using an experimental procedure similar to that described for the previous experiment. In this series of experiments the sintered funnel was replaced by the end-cap, fitted with a ceramic plate of porosity 4 (which was used in the unsaturated conductivity experiments). The soil was packed on this plate, inside a perspex tube of 4.5 cm diameter and 6 cm in height. The soil was confined at the top by a wire mesh held in place by a hose clamp and the perspex tube was clamped to the end-cap using brass screws.

E. Saturated Conductivity of Soils for Salt Solutions

The hydraulic conductivity of the soils for different salt solutions was measured using the apparatus shown in Figure 4 with soil columns of 2.0 to 0.2 mm size fractions of the krasnozem, alluvial and red brown soils packed to bulk densities of 1.00 ± 0.01 , 1.13 ± 0.01 and 1.22 ± 0.01 g/cm³ respectively. A perspex tube of 5.04 cm diameter and 12 cm long, fitted with a wire mesh 5 cm from the bottom of the tube, was inverted and filled with the soil.

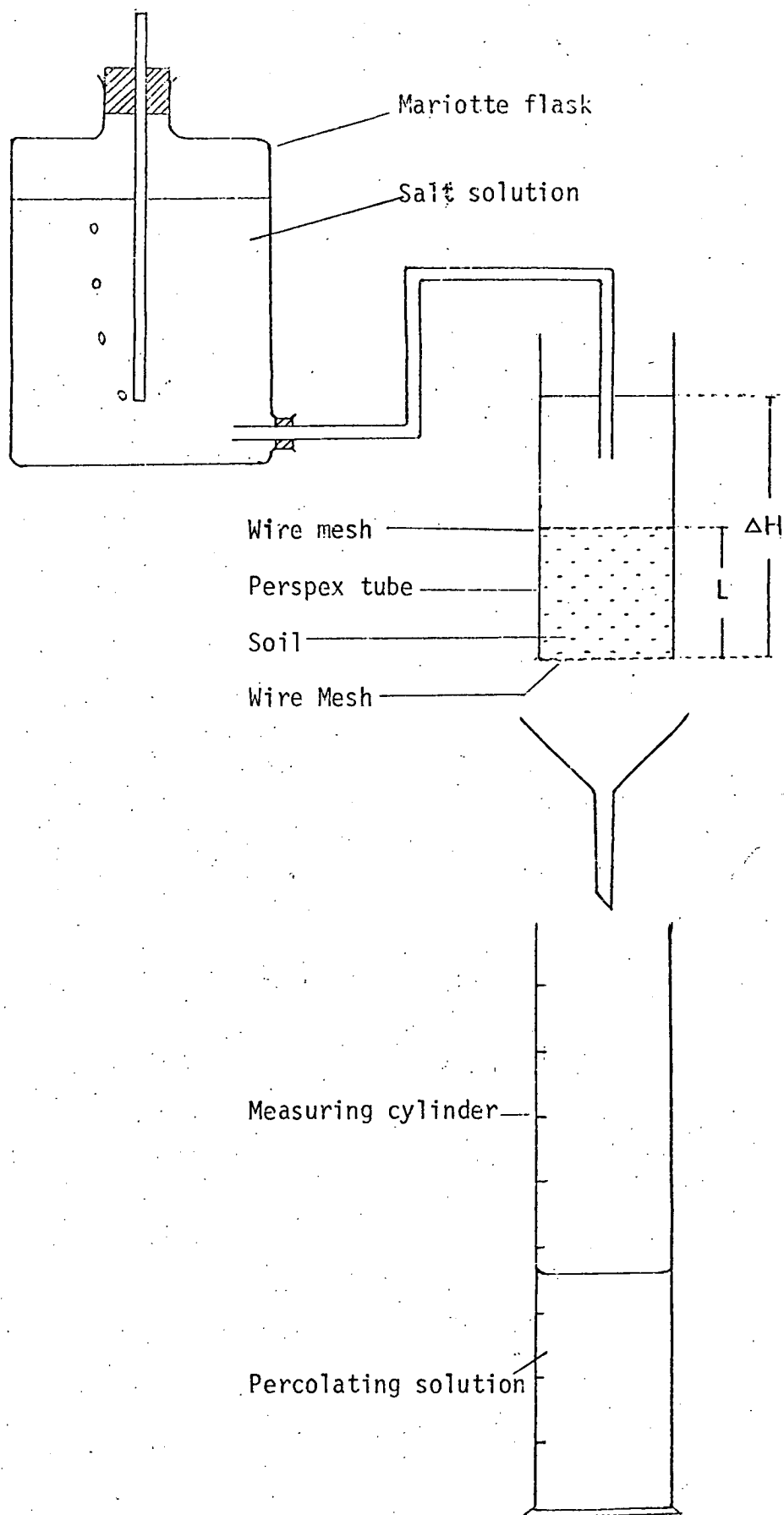


Fig. 4. Permeameter apparatus used in determining the saturated hydraulic conductivity of soils for salt solutions.

Another wire mesh was placed over the soil and secured onto the end of the tube using a hose clamp. Carbon dioxide was passed through the soil to displace the air. The tube was turned over and the soil slowly wetted through a filter paper and then saturated by immersion in a salt solution of SAR 40 and the highest cation concentration of 640 meq/l for 48 hours. The saturated soil was leached with 20 pore volumes of the same salt solution siphoned into the top of the perspex tube from a 10 litre mariotte flask. A constant head of water was established in the upper part of the tube to give a hydraulic gradient of two. The rate of outflow was measured and the saturated hydraulic conductivity was calculated using the relationship

$$K = (Q/At)(L/\Delta H)$$

where Q is the outflow in time t, A is the area of cross section and $(\Delta H/L)$ is the hydraulic gradient.

The soil was then leached with the solution of the same SAR and the next highest cation concentration until the soil reached chemical equilibrium with the salt solution. Twenty pore volumes of the salt solution were used in the case of the kraznosem soil and 24 hours of leaching was used in the case of the other soils. The saturated conductivity of the soils in the solution of lower cation concentration was measured as before. This process of successively leaching with solutions of lower salt concentration at the same SAR and determining the saturated conductivity was continued until conductivity for the solution of lowest cation concentration was measured.

The experiment was repeated with new soil columns for alluvial and red brown soils with series of salt solutions of SAR 20 and 10 and different cation concentrations using the same experimental

procedure. The experiment was also repeated with soil columns of alluvial soil packed to a bulk density of 1.22 g/cm^3 , for a series of salt solutions of SAR 40, 20 and 10 and different cation concentrations.

F. Unsaturated Conductivity of Soils at High Pressure

Heads for Salt Solutions

The unsaturated hydraulic conductivity of soils to salt solutions was determined by applying a constant hydraulic head across the unsaturated soil column and measuring the resultant steady flow of solution using the method and type of apparatus (Figure 5) described by Klute (1965). A brass pressure cell, 5 cm in internal diameter and 5 cm in height was used (Figure 6). This cell had an opening for the air pressure tube on one side and on the opposite side two openings for tensiometers positioned 1 cm from each end of the cell. Each tensiometer consisted of a 4 cm long brass tube with a porous plate of porosity 4 cemented to the end. A brass olive on each tensiometer tube close to the porous plate provided an airtight connection when the tensiometer nut was screwed tightly against the pressure cell. Two end-caps fitted over the ends of the pressure cell. A porous plate (porosity 4) was cemented to the inside wall of each end-cap. Two tubes opened into the space between the end-cap and the porous plates. A groove was machined onto the inside edges of each end-cap for fitting an O-ring to provide an airtight seal when the end-caps were tightly clamped to the two ends of the pressure cell by means of 4 brass screws.

The experiment was carried out on soil columns of the 2.0 to 0.2 mm size fractions of the krasnozem, alluvial and red brown soils packed

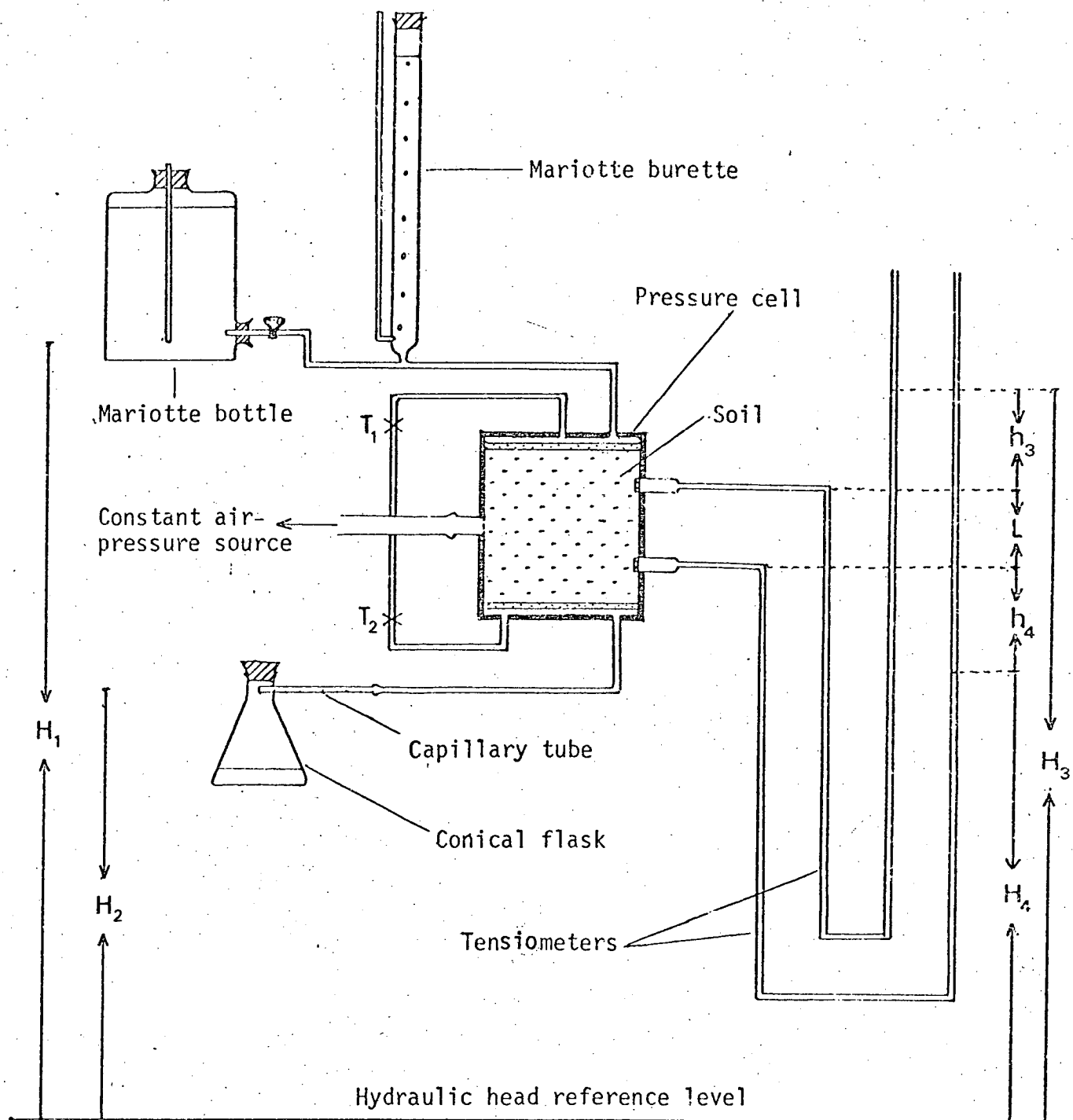


Fig. 5. Schematic diagram of apparatus for the steady state method of measuring unsaturated conductivity of soils.

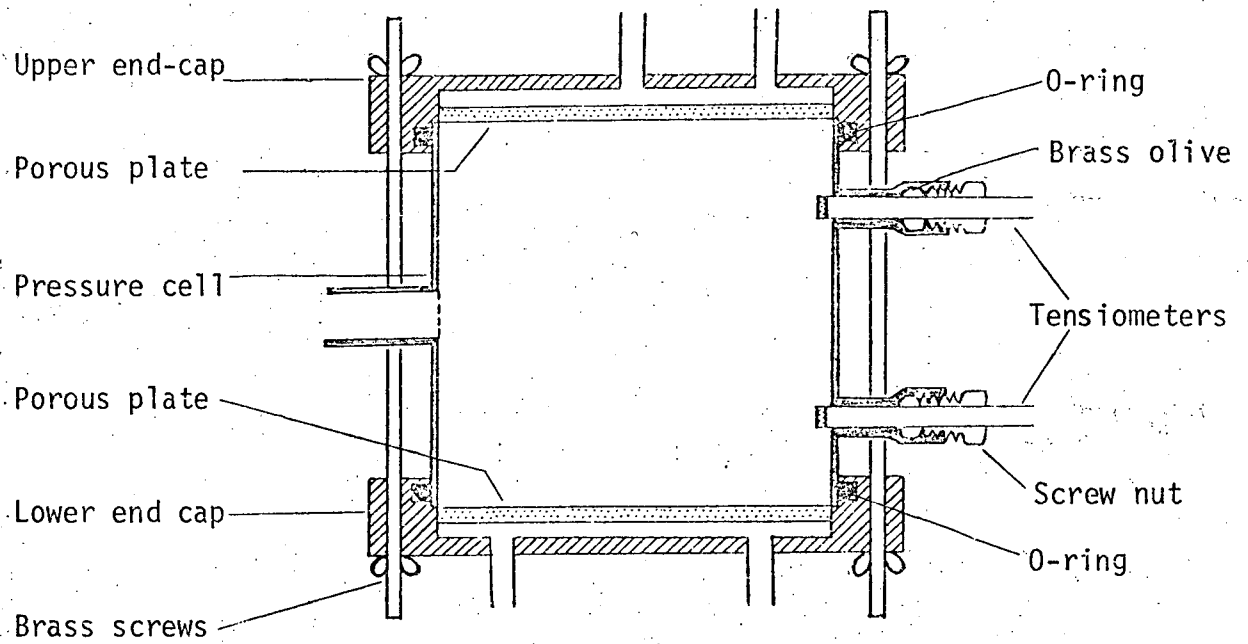


Fig. 6. Pressure cell assembly used in determining unsaturated conductivity of soils.

to bulk densities of 1.00 ± 0.01 , 1.22 ± 0.01 and 1.22 ± 0.01 g/cm³ respectively.

Tensiometers, end-caps and the solution supply and removal systems were filled with the salt solution to be used in the experiment. The pressure cell was attached to the lower end-cap. With the porous plate of this end-cap at a suction of 60 cm of water, the soil was packed into the cell. The upper end-cap was placed in position and the air in the cell was replaced by carbon dioxide. The two end-caps were then tightly clamped over the pressure cell, to make the cell airtight. The tensiometers and the solution supply and removal systems were connected and the soil wetted initially under suction. The soil was saturated by raising the level of the solution supply system, and then leached with at least 20 pore volumes of the salt solution to ensure that the soil was in chemical equilibrium with the salt solution.

A flow system was established such that the solution supplied at a hydraulic head, H_1 , to the upper porous plate, flowed through the plate, the soil column and lower porous plate into the space below the lower plate, where a hydraulic head, H_2 , was maintained by adjusting the location of the drip point. The solution outflow was collected in a flask and measured by weighing or alternately by observing the movement of an air bubble in a horizontal capillary tube attached to the drip point. The taps, T_1 and T_2 , were always kept closed except when it was necessary to flush out the air from the end-cap spaces above and below the porous plates. The air collecting in the tensiometers was removed by flushing through a capillary tube which opened close to the porous plate and which was kept closed when the tensiometer was in use. Initially

an air pressure of 20 cm of water was applied. A constant hydraulic head was established and maintained until steady flow was attained, indicated by equality of inflow and outflow rates and time-invariant readings of the tensiometers. The volume of outflow, Q , that occurred in time, t , during steady flow, the pressure heads, h_3 and h_4 , and the hydraulic heads, H_3 and H_4 , were recorded.

The moisture content of the soil column was reduced gradually by increasing in increments the gas pressure, P , in the cell to 30, 50, 80, 100, 120, 160 cm of water. At each of these applied gas pressures, after steady flow was obtained, the rate of steady outflow, pressure heads and hydraulic heads were measured. From this data, the unsaturated hydraulic conductivity ($k[h]$) at a specific pressure (h) was calculated using

$$k[h] = (Q/At) (L/(H_3 - H_4))$$

where $h = P + (h_3 + h_4)/2$, A is the cross-sectional area of the soil column, and L is the distance between the tensiometers.

The unsaturated conductivities at different pressure heads were determined for krasnozem soil in salt solutions of SAR 20 and cation concentrations of 160 and 2.5 meq/l, and for alluvial soils in salt solutions of SAR 20 and cation concentrations of 160, 40 and 10 meq/l. Unsaturated conductivities were determined for red brown soils in salt solutions of SAR 20 and cation concentrations of 160, 40 and 10 meq/l and also in solutions of SAR 40 and cation concentrations of 80 and 40 meq/l. Separate soil columns were used for each solution.

G. Moisture Release Curves and Unsaturated Conductivity of Soils at Low Pressure Heads for Salt Solutions

The moisture release curves were determined on the soil columns

used in the previous experiment. One end-cap of the pressure cell was replaced by a confining plate and the other end-cap was replaced by an end-cap fitted with a plate of one atmosphere bubbling pressure. The tensiometers were removed and all openings on the sides of the cell were closed. The soil was rewetted, resaturated and leached with the same salt solutions used in the previous experiment. The cell was placed in a pressure chamber with one outlet of the end-cap closed and the other outlet connected to the outflow tube (Figure 7). The air pressure was raised in stages to 10, 20, 30, 50, 80, 100, 120, 160 and 200 cm of water and the outflow at each of these air pressures was measured to give the moisture release curve of the soil for that salt solution.

The unsaturated conductivity at low pressure heads was determined using the one-step outflow method (Gardner, 1962; Doering, 1965). The air pressure in the cell was raised from 200 to 500 cm of water and the outflow was measured at regular time intervals until the outflow ceased. The diffusivity, $D(\theta)$, of the soil to salt solutions at pressure heads between -200 to -500 cm of water was calculated using

$$D(\theta) = - \frac{4L^2}{\pi^2(\theta - \theta_f)} \frac{d\theta}{dt}$$

where L is the length of the soil column, θ_f the final equilibrium water content, and θ the water content when the instantaneous outflow rate is $d\theta/dt$. The unsaturated conductivity $k(\theta)$ at a specific water content was calculated using

$$k(\theta) = -D(\theta) \frac{d\theta}{dS}$$

where $d\theta/dS$ is the slope of the water content versus suction relationship at the water content corresponding to $D(\theta)$.

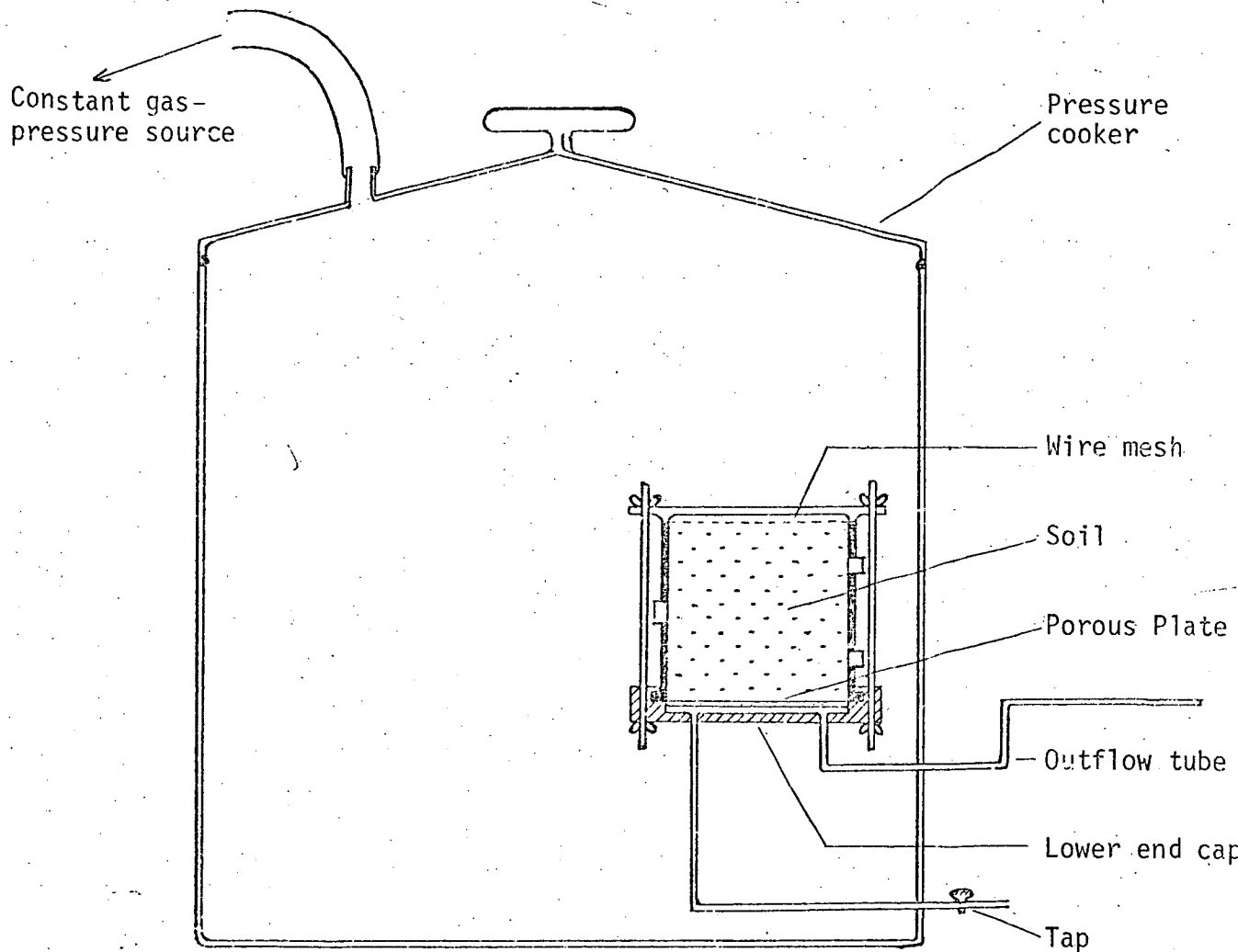


Fig. 7. Diagram of the apparatus for the outflow method of measuring the unsaturated conductivity of soils for salt solutions at low pressure heads.

H. Rates of Evaporation from Soils with Water Tables of Different Salt Solution Composition

A soil column packed inside a perspex tube of 4.5 cm internal diameter and 100 cm in length was used in this experiment. The 2.0 to 0.2 mm size fraction of soil was packed to bulk densities of 1.00 ± 0.01 , 1.22 ± 0.01 , 1.22 ± 0.01 g/cm³ for krasnozems, alluvial and red brown soils respectively. Uniform packing and bulk densities were obtained by pouring in weighed amounts of soil to fill measured short lengths of the soil column through a smaller tube, while rotating and tapping the sides of the column. A confining plate was secured by a hose clamp over the top of the column. The bottom of the soil column was wetted through a filter paper with a salt solution of SAR 20 and concentration of 160 meq/l. The same solution was then introduced from a mariotte burette through a flexible tube connected to the base of the column. The soil column was slowly wetted and saturated by raising the level of the solution in the soil column. The soil was then leached with the same salt solution until chemical equilibrium was reached.

The soil column was then placed on a turntable (Figure 8) and positioned so that the top of the column fitted into a hole in the circular wire frame covered with aluminium foil. The foil ensured that only the top of the soil column was exposed and prevented the column below the surface receiving heat during the evaporation experiment. The turntable was rotated at a speed of 2 rpm. The four 150 watt lamps positioned above the columns to provide heat and the fan which provided the ventilation were switched on. The speed of the fan and the distance of the lights above the soil column were adjusted to give external evaporation rates which slightly exceeded the maximum rate of upward flow. This external

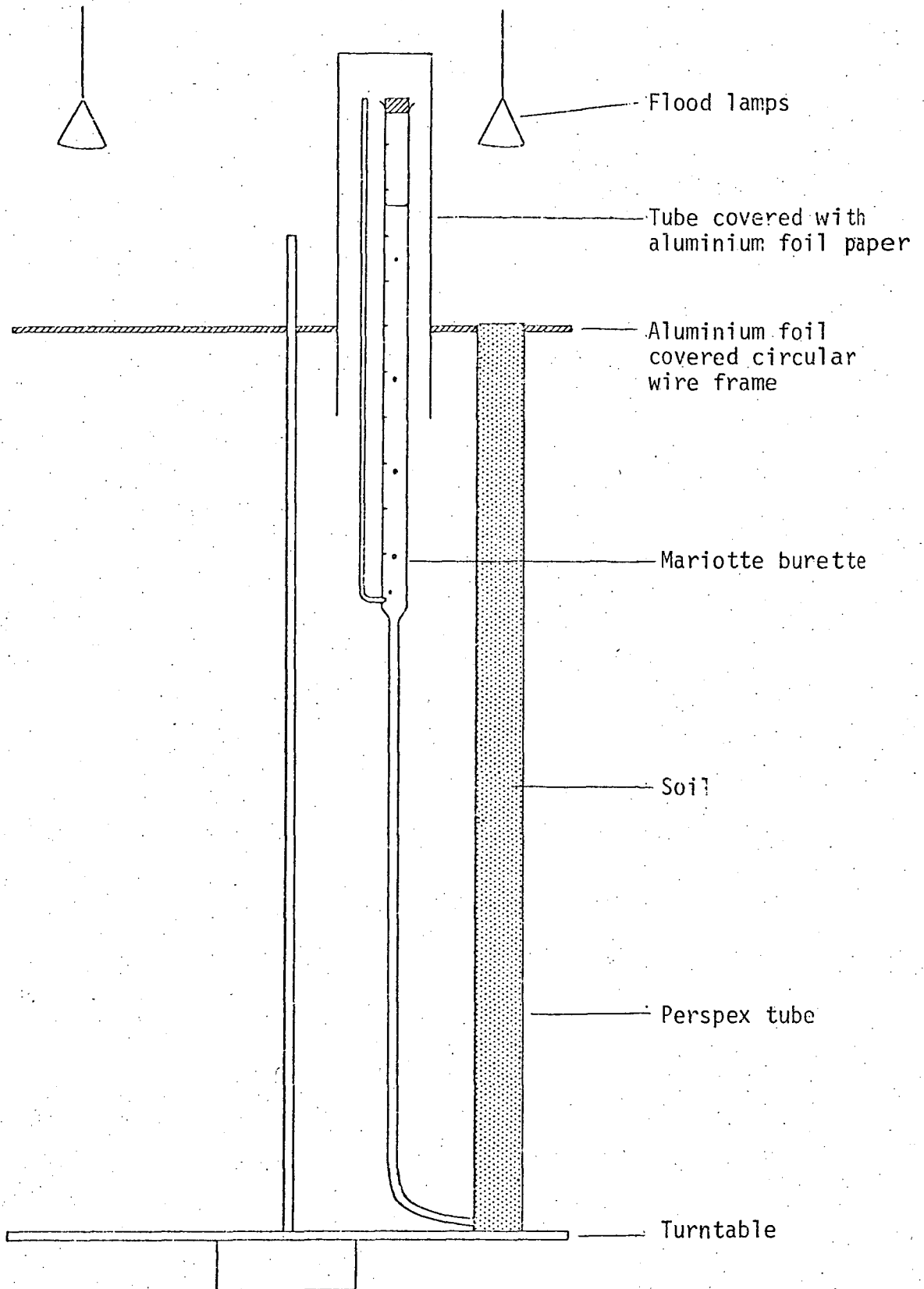


Fig. 8. Diagram of the apparatus used in determining the rates of upward flow in soil columns from a saline water table.

evaporation rate was determined by recording the rate of inflow from a mariotte burette into a tube filled with water, the top of which was at the same level as the soil surface in the columns. The water table was positioned at a depth of 30 cm and the confining plate was removed. A record of the rate of flow of solution from the mariotte burette into the soil column was kept. After steady flow was attained the rate of upward flow to the surface in the soil column was calculated. The rate of evaporation from a specific depth to water table was measured over a period of one week. The depth to the water table was increased by lowering the mariotte burette to 40, 45, 50, 60, 80 and 100 cm and the steady-state evaporation rate from each of these depths was similarly measured. The soil surface was scraped before each measurement and a two centimeter segment of the soil column was removed after two successive measurements of steady upward flow, in order to minimise the possible effects on evaporation rates of salt accumulation near the surface.

The alluvial and red brown soil columns were then rewetted from the base of the column and leached with salt solution of SAR 20 and the next highest cation concentration of 40 meq/l till chemical equilibrium with the salt solution was reached. The steady evaporation rates from different depths to water table were then measured using the same experimental procedure. The soil was then rewetted and leached with salt solution of SAR 20 and cation concentration of 10 meq/l till chemical equilibrium was attained and the steady evaporation rates from different depths to water table were determined in a similar manner. The krasnozem soil column was rewetted and leached with solution of SAR 20 and cation concentration of 2.5 meq/l and the steady evaporation rates from different depths to water table were similarly measured.

IV. RESULTS AND DISCUSSION

A. Physical and Chemical Characteristics of Soils

Some physical and chemical characteristics of krasnozems, alluvial and red brown soils used in the present study are given in Table 2.

All soils contained more than 35 per cent clay. The minimum cation concentrations of salt solutions at specific SARs in which the clay extracted from these soils flocculated, are given in Table 3. It was found by Collis-George and Smiles (1963), that the minimum cation concentration at which clay flocculated correlated closely with the concentration at which significant decreases in permeability occurred, in the absence of factors conferring a structural stability on soil aggregates. However, it was observed that the flocculated clay of alluvial and red brown soils was swollen and it is possible that hydraulic conductivity decreases could occur in these soils at higher cation concentrations in solutions of specific SARs due to clay swelling. On the other hand, it is also possible that the cementing agents present in the soils, which were largely removed in extracting the clay for this experiment, could reduce or inhibit clay swelling and clay dispersion and hence prevent decreases in hydraulic conductivity; at least until cation concentrations much lower than the values for minimum cation concentrations for flocculation given in Table 3, were reached.

In a preliminary study, the swelling of premoistened 1 to 2 mm sized aggregates of these soils placed in different salt solutions was observed under a microscope. While the aggregates of alluvial and red brown soil showed very marked swelling in low salt, high sodium solutions, the aggregates of krasnozems soil showed little change in their size in different salt solutions.

Table 2. Physical and chemical characteristics of soils

	KRASNOZEM	ALLUVIAL	RED BROWN
Particle size analysis - Sand (2.0-0.02mm)	11.1	52.9	32.1
Silt (0.02-0.002mm)	20.2	9.1	25.6
Clay (<0.002mm)	66.9	36.9	40.6
Soil pH	5.9	7.1	6.9
Organic carbon %	1.26	1.40	0.81
Free iron oxides %	6.76	0.82	5.21
Cation exchange capacity (meq/100 g soil)	18.27	43.32	24.57
Exchangeable bases (meq/100 g soil)	3.57	29.79	23.81

Table 3. Minimum cation concentration (meq/l) of solutions of SAR 10, 20, 40 and 60 in which clay flocculation occurred.

SOIL	SAR			
	10	20	40	60
Krasnozem	6.5	13.5	26	29
Alluvial	5.5	10.5	17	35
Red Brown	8.5	20.5	45	60

B. Effect of Salt Solution Composition on Moisture Release Curves of the Soils

(i) Moisture release curves

The moisture release curves of krasnozem, alluvial and red brown soils in the presence of a series of salt solutions of SAR 0, 10, 20, and 40 and different cation concentrations between 640 and 2.5 meq/l are given in Figures 9 to 19. The changes in shape of these curves show the effects of salt solution composition on the structural stability (Childs, 1940) and pore geometry of the soils, which in turn could be expected to relate to the changes in saturated and unsaturated conductivity at high pressure heads in these soils in the presence of the same salt solutions.

The equivalent pore neck radii of the most common pore of the soils for specific salt solutions, obtained from the differential of the moisture release curve and the relationship between the radius of a capillary tube and the matric potential are given in Table 4. Childs (1940) and Collis-George and Laryea (1972) used the ratio of the radii of necks of the common pore in soils subject to quick and slow wetting as an index of the structural stability of the soil, in resisting a breakdown of soil aggregates on quick wetting. In the present study it was observed that in some salt solutions there was a visually observable swelling of the soil aggregates into the inter-aggregate pores, although the aggregates themselves did not breakdown into smaller aggregates. A pore size index defined as the ratio of the radius of the common pore neck of the soil in solutions of lower cation concentration to the radius of the common pore neck in the solution of the highest cation concentration at the same SAR, was used in the present study as a quantitative measure of the relative changes in size of common pores of a soil in different salt solutions.

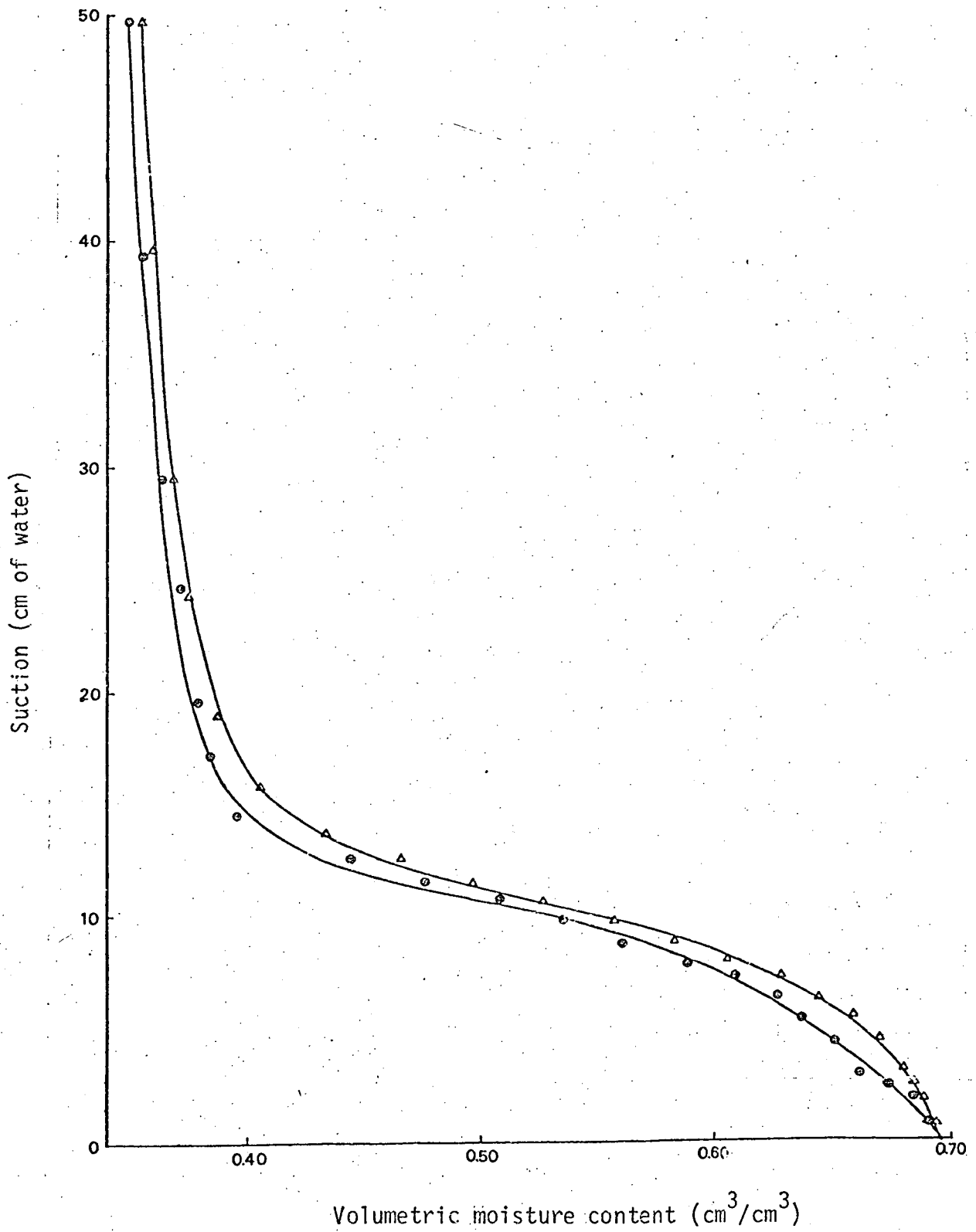


Fig. 9. Moisture release curves of krasnozem soil for salt solutions of SAR 0 and cation concentrations (meq/l); ● 160, ▲ 2.5. Bulk density of soil, 0.88 g/cm³.

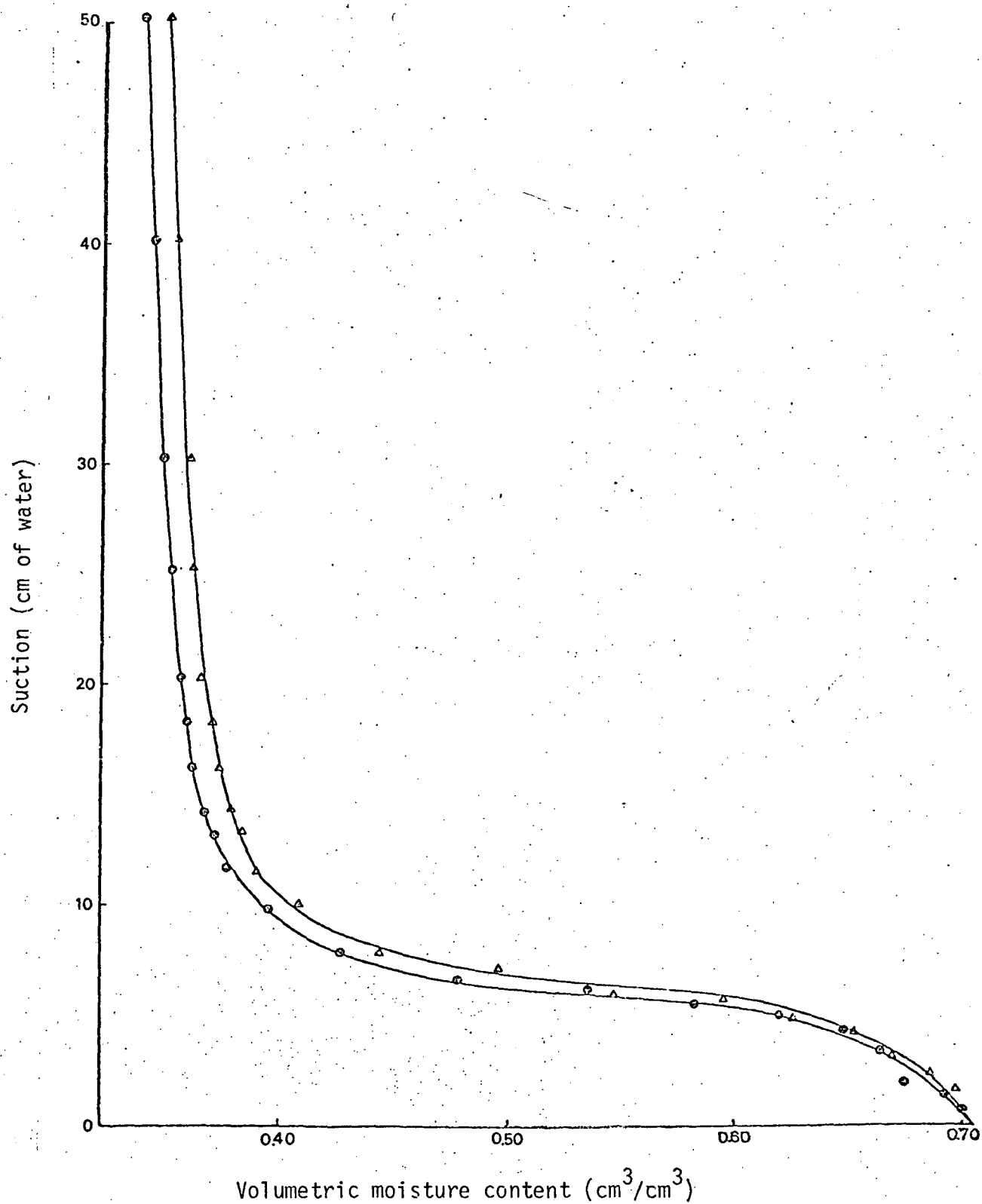


Fig. 10. Moisture release curves of krasnozem soil for salt solutions of SAR 20 and cation concentrations (meq/l); \circ 160, \triangle 2.5. Bulk density of soil, 0.85 g/cm³.

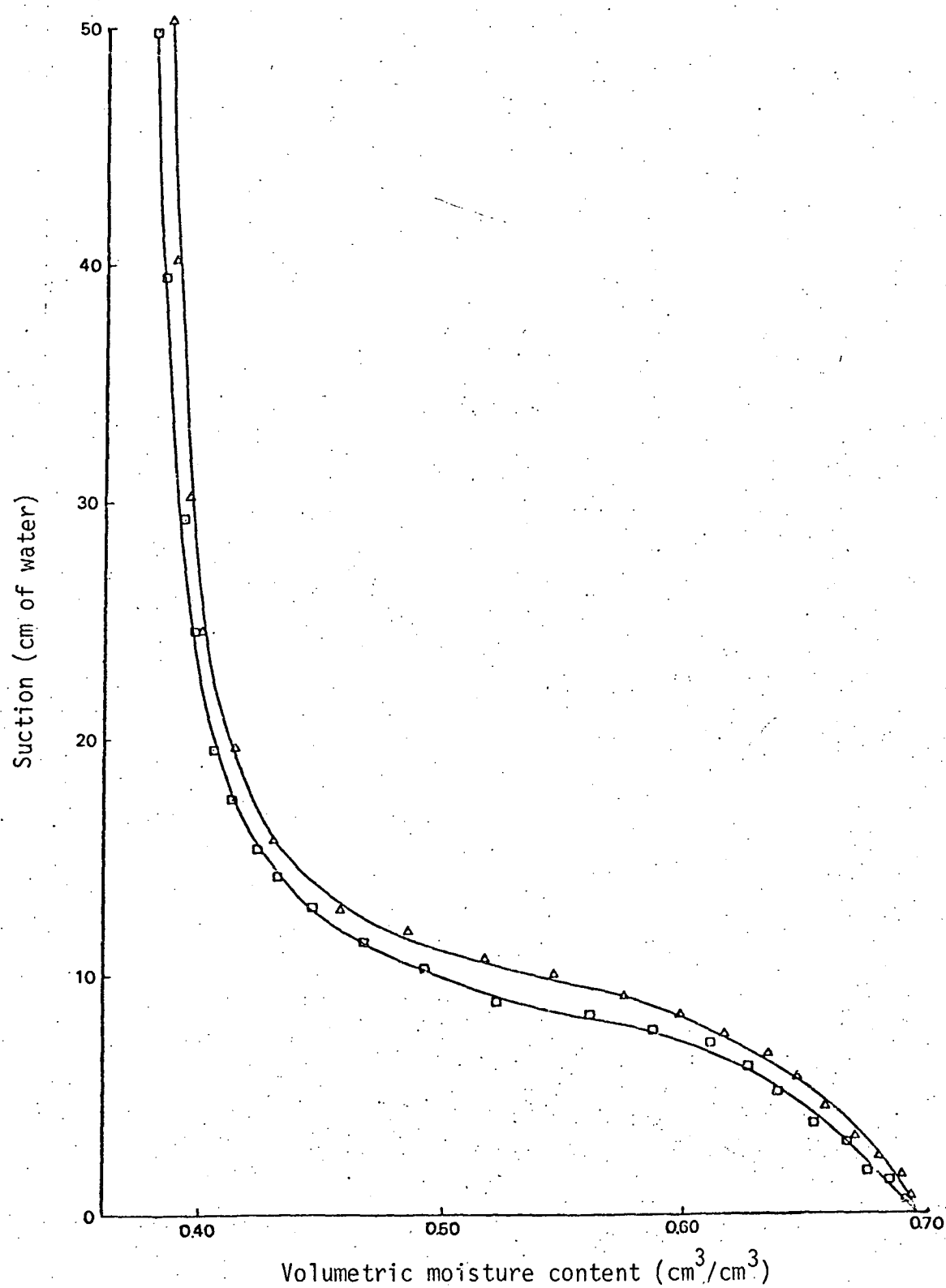


Fig. 11. Moisture release curves of krasnozem soil for salt solutions of SAR 40 and cation concentrations (meq/l); \square 640, \triangle 2.5. Bulk density of soil, 0.89 g/cm^3 .

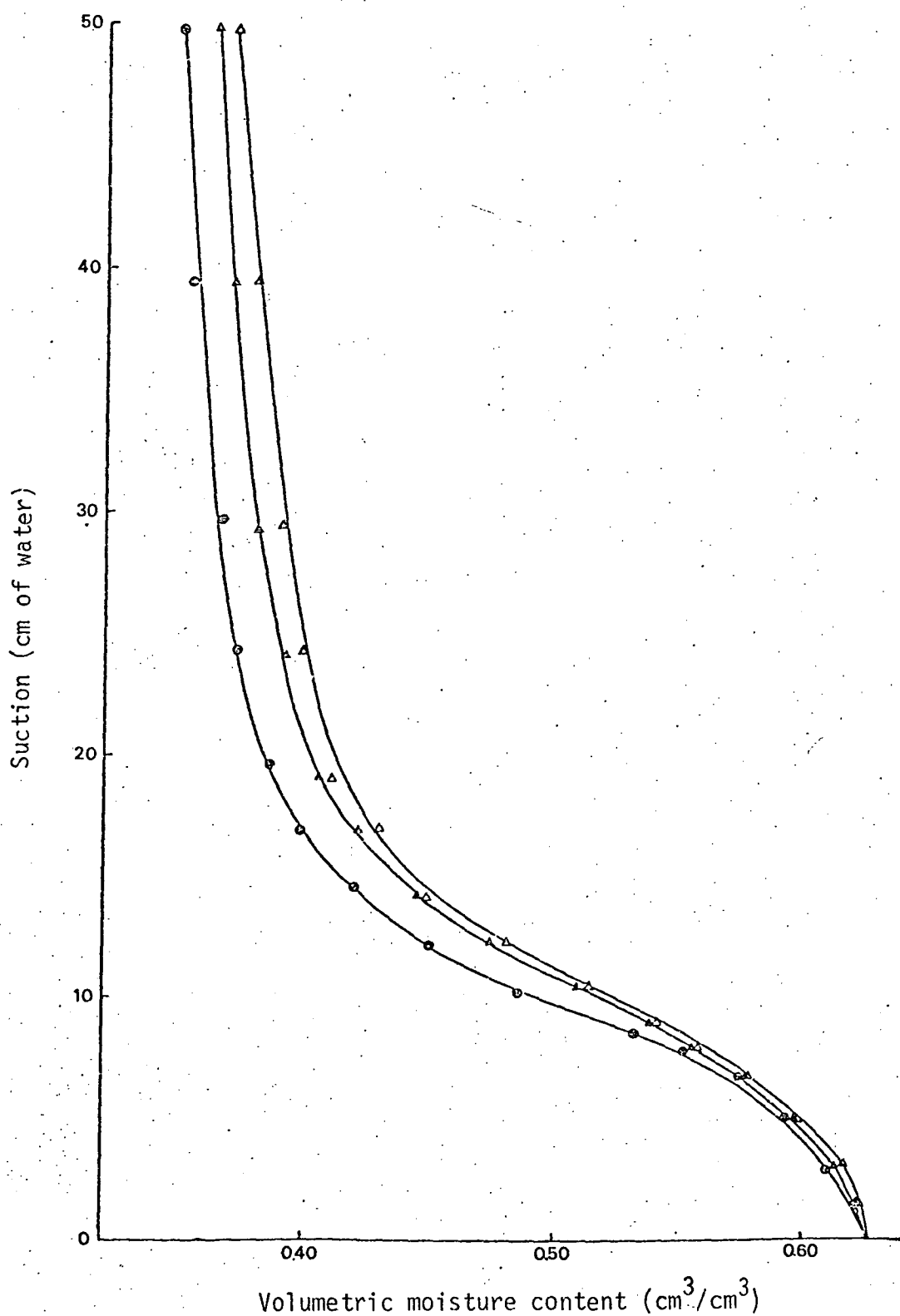


Fig. 12. Moisture release curves of alluvial soil for salt solutions of SAR 0 and cation concentrations (meq/l); \odot 160, \triangle 10, \triangle 2.5. Bulk density of soil, 1.00 g/cm³.

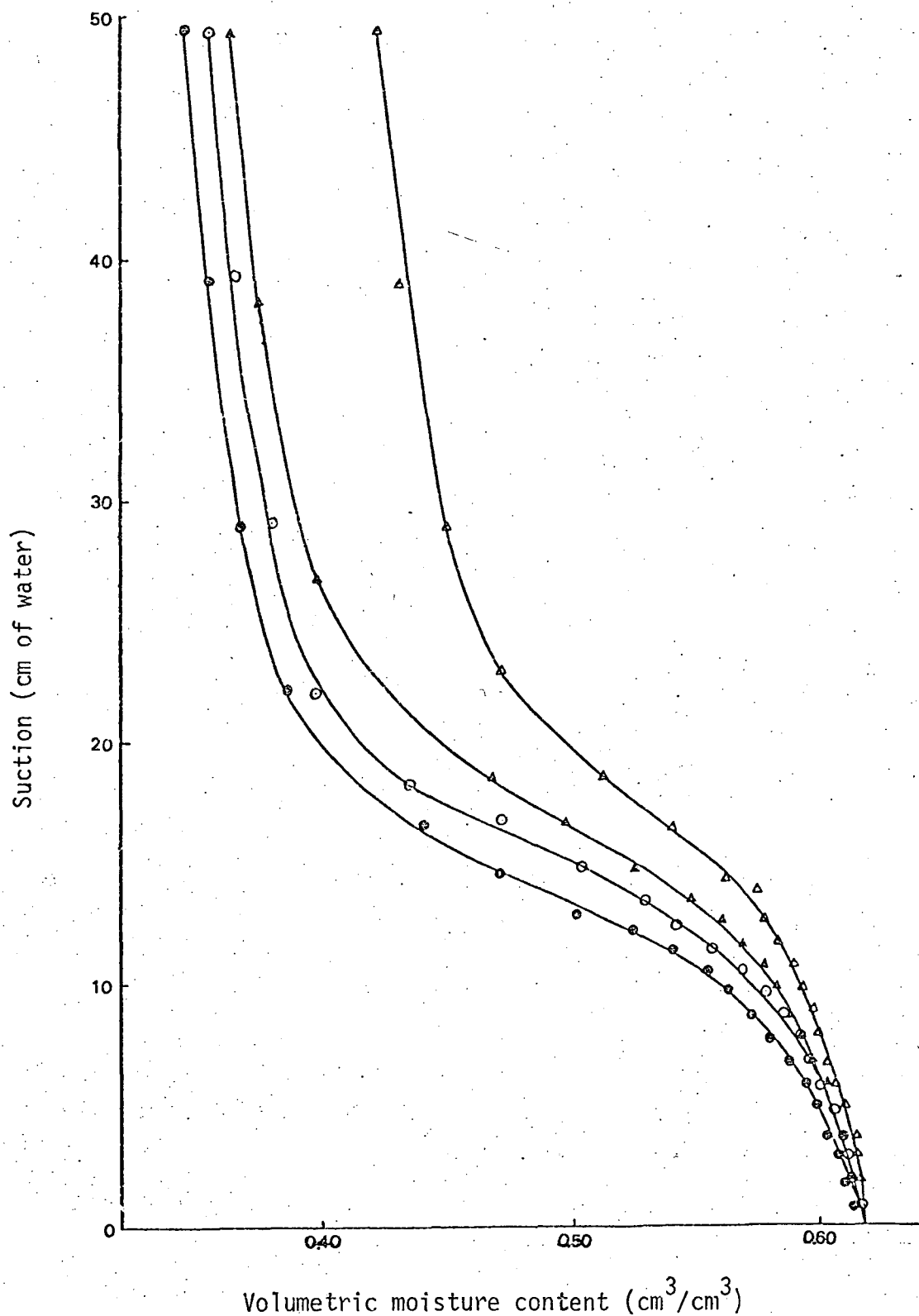


Fig. 13. Moisture release curves of alluvial soil for salt solutions of SAR 10 and cation concentrations (meq/l); ● 160, ○ 40, ▲ 10, △ 2.5. Bulk density of soil, 1.02 g/cm³.

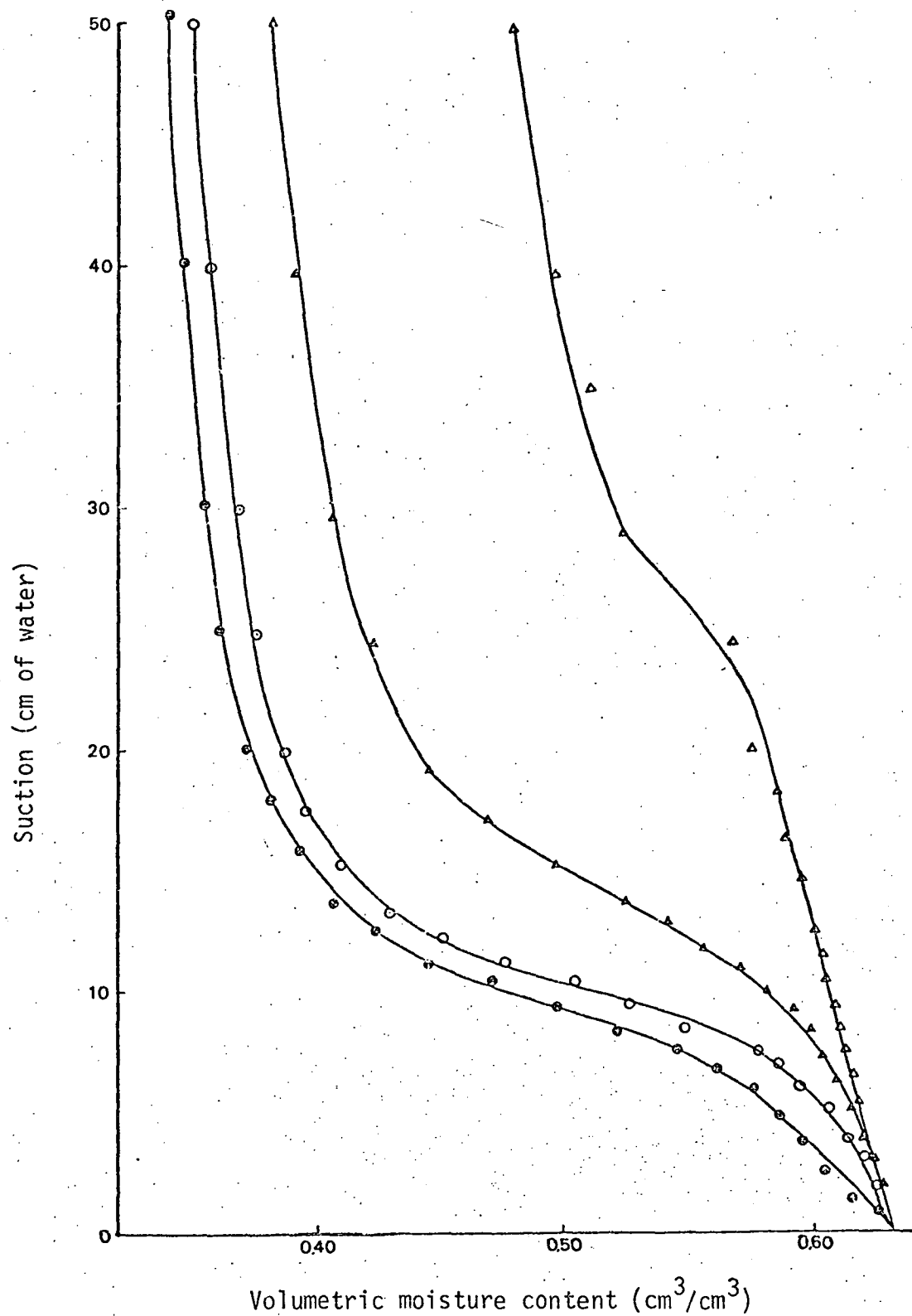


Fig. 14. Moisture release curves of alluvial soil for salt solutions of SAR 20 and cation concentrations (meq/l); ● 160, ○ 40, ▲ 10, △ 2.5. Bulk density of soil, 1.00 g /cm³.

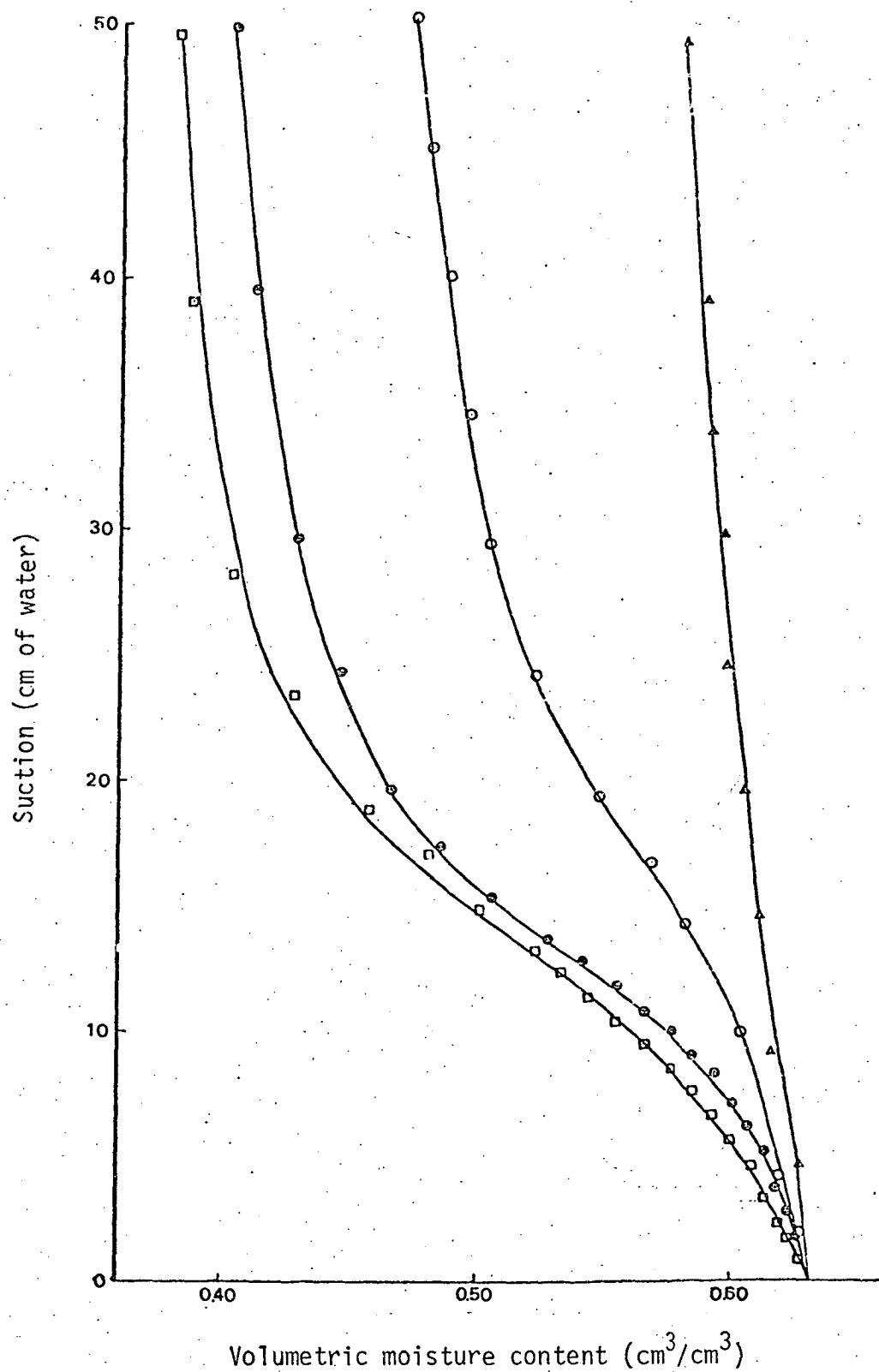


Fig. 15. Moisture release curves of alluvial soil for salt solutions of SAR 40 and cation concentrations (meq/l); \square 640, \odot 160, \circ 40, \triangle 10. Bulk density of soil, 1.00 g/cm^3 .

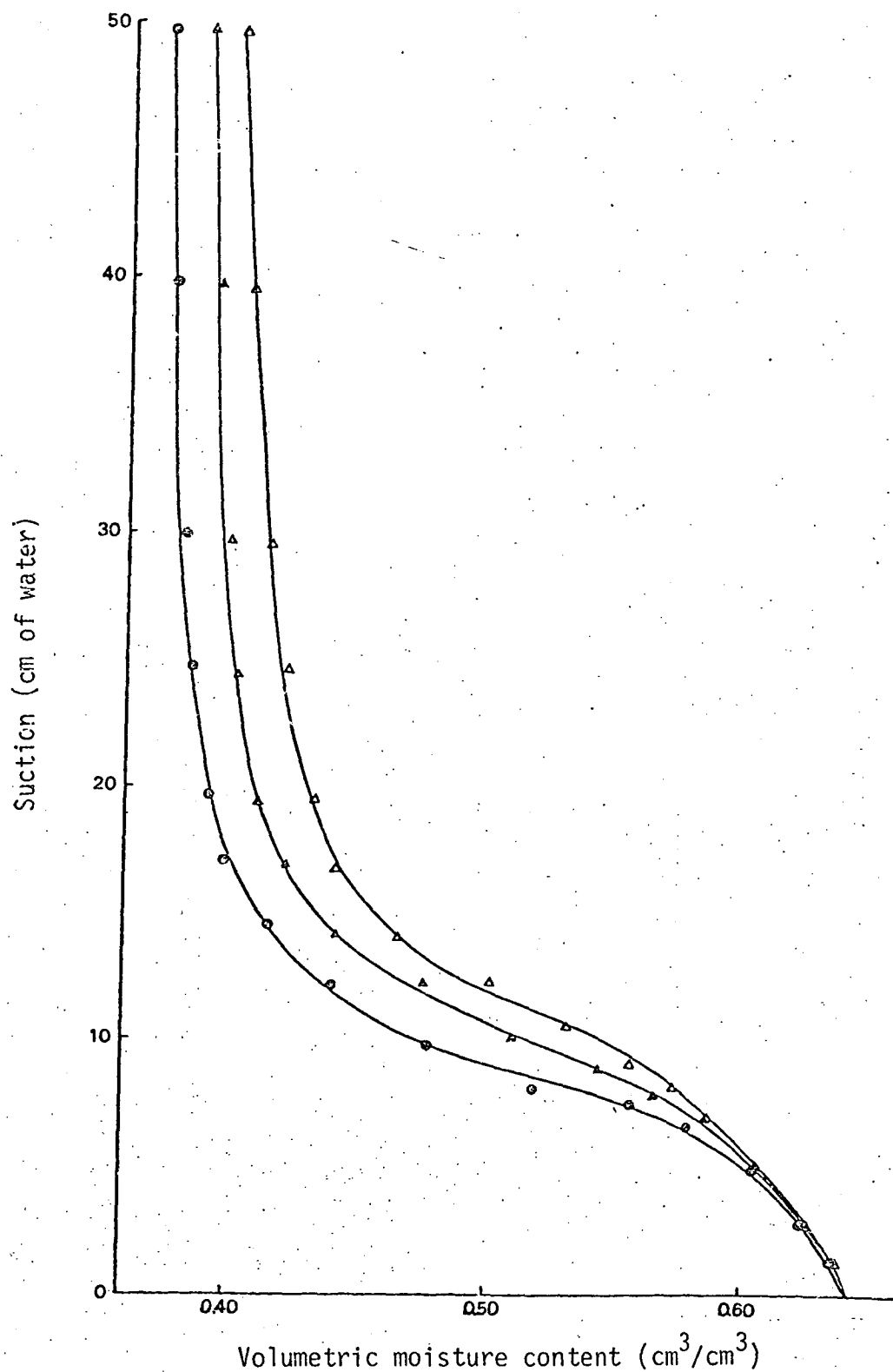


Fig. 16. Moisture release curves of red brown soil for salt solutions of SAR 0 and cation concentrations (meq/l); \circ 160, \triangle 10, \triangle 2.5. Bulk density of soil, 1.04 g/cm³.

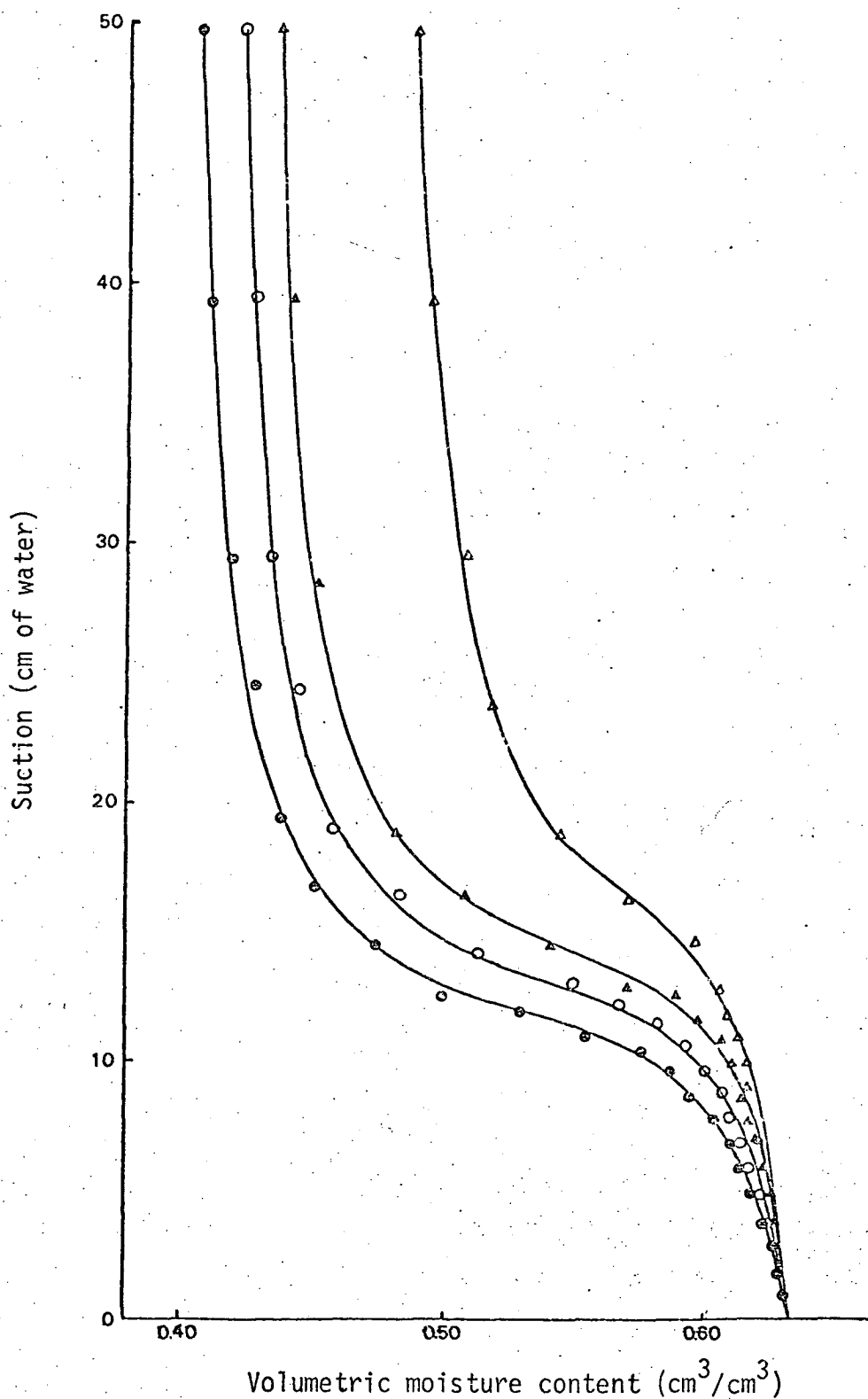


Fig. 17. Moisture release curves of red brown soil for salt solutions of SAR 10 and cation concentrations (meq/l); ● 160, ○ 40, ▲ 10, ▼ 2.5. Bulk density of soil, 1.07 g/cm³.

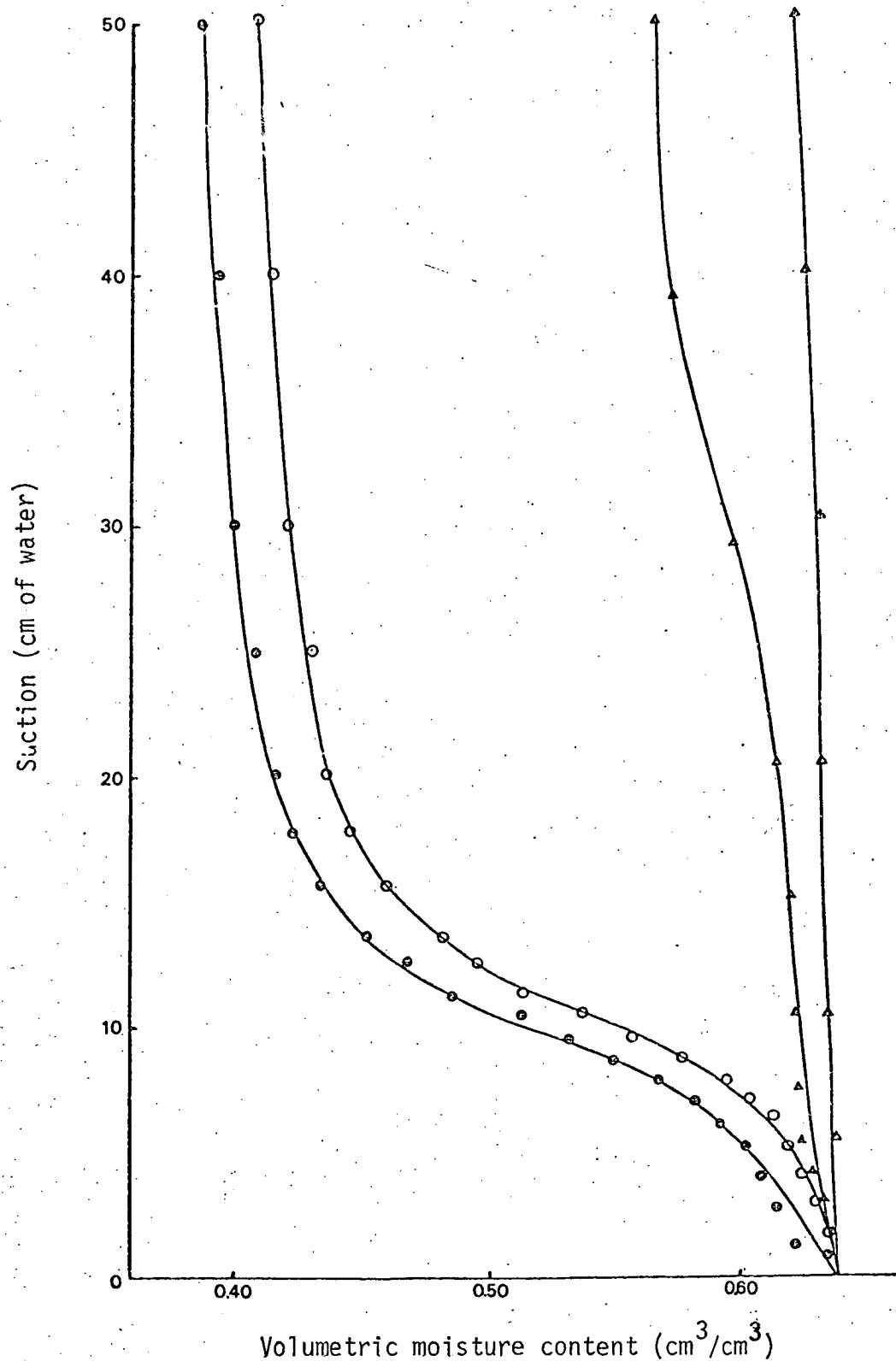


Fig. 18. Moisture release curves of red brown soil for salt solutions of SAR 20 and cation concentrations (meq/l); \odot 160, \circ 40, \triangle 10, \triangle 2.5. Bulk density of soil, 1.06 g/cm^3 .

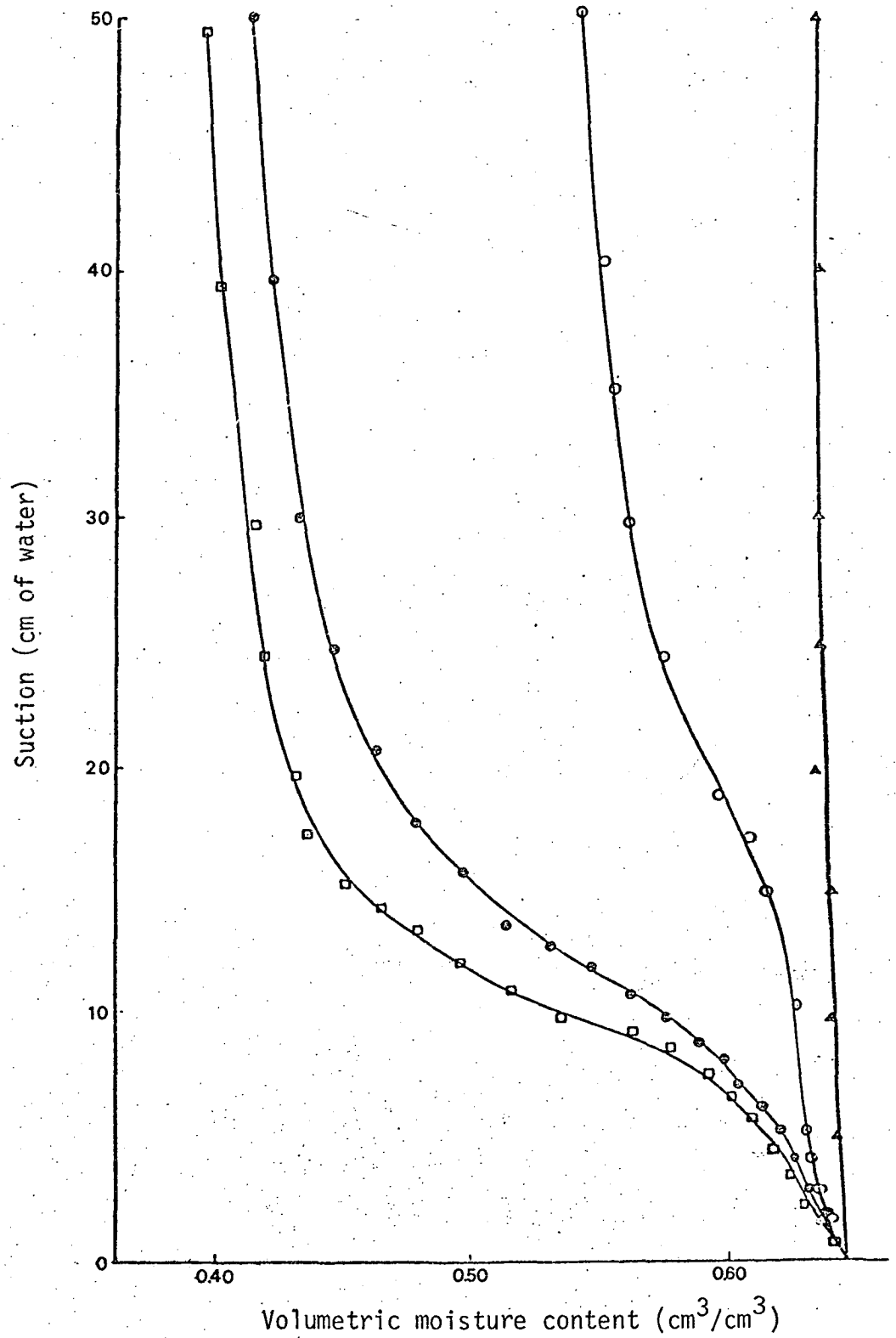


Fig. 19. Moisture release curves of red brown soil for salt solutions of SAR 40 and cation concentrations (meq/l); \square 640, \bullet 160, \circ 40, \triangle 10. Bulk density of soil, 1.04 g/cm^3 .

In the presence of salt solutions of low sodium absorption ratio (SAR 0) and high cation concentration (160 meq/l) the moisture release curves of all soils showed a distinct inflexion at high pressure heads due to a substantial pore volume occupied by pores with relatively large necks draining within a narrow pressure head range (Figures 9, 12 and 16). With solutions of SAR 0 and low cation concentrations (10 and 2.5 meq/l) all three soils showed little visual evidence of swelling and only small changes in their moisture release curves. The persistence of large inter-aggregate pores in this low sodium solution even when the cation concentration was very low indicated the high stability and low swelling of soil aggregates in such solutions. Accordingly, the pore size index remained high even at low cation concentrations in this low sodium solution (Table 5).

With solutions of high sodium absorption ratio (SAR 40) and high cation concentration (640 meq/l) all soils were structurally stable as indicated by their moisture release curves (Figures 11, 15 and 19) showing a distinct inflexion at low pressure heads.

When leached with solutions of SAR 40 and lower cation concentrations the krasnozem soil, predictably showed very little change in moisture release curves and the pore size index remained high (Table 5).

In the case of the alluvial and red brown soils, as the cation concentration of this SAR 40 solution was reduced, a shift of the moisture release curves to the right (Figures 15 and 19) indicated a reduction in size of the necks of the inter-aggregate pores, caused by a visually observable swelling of the soil aggregates into this pore space. Hence, as the salt concentration was reduced in this high sodium solution, the pore size index fell (Table 5).

With both alluvial and red brown soils, as the cation concentration

Table 4. Radii of necks of common pore (mm) in krasnozem, alluvial and red brown soils packed to bulk densities of approximately 0.87, 1.01 and 1.05 g/cm³ respectively, for different salt solutions.

SOIL	KRASNOZEM			ALLUVIAL				RED BROWN			
CATION CONCENTRATION (meq/l)	SAR			SAR				SAR			
	0	20	40	0	10	20	40	0	10	20	40
640	-	-	0.173	-	-	-	0.126	-	-	-	0.161
160	0.144	0.150	-	0.160	0.109	0.165	0.118	0.177	0.128	0.158	0.129
40	-	-	-	-	0.094	0.138	0.085	-	0.113	0.135	0.077
10	-	-	-	0.146	0.088	0.110	<0.025	0.152	0.104	0.049	<0.025
2.5	0.136	0.144	0.146	0.143	0.085	0.056	<0.025	0.130	0.091	<0.025	<0.025

Table 5. Pore size indices of krasnozem, alluvial and red brown soils, packed to bulk densities of approximately 0.87, 1.01 and 1.05 g/cm³ respectively, for different salt solutions.

SOIL	KRASNOZEM			ALLUVIAL				RED BROWN			
CATION CONCENTRATION (meq/l)	SAR			SAR				SAR			
	0	20	40	0	10	20	40	0	10	20	40
640	-	-	1.00	-	-	-	1.00	-	-	-	1.00
160	1.00	1.00	-	1.00	1.00	1.00	0.97	1.00	1.00	1.00	0.80
40	-	-	-	-	0.86	0.83	0.68	-	0.88	0.86	0.48
10	-	-	-	0.91	0.81	0.67	<0.12	0.86	0.81	0.31	<0.15
2.5	0.95	0.96	0.84	0.89	0.78	0.34	<0.12	0.74	0.71	<0.16	<0.15

was reduced from 640 to 160 meq/l the loss of structural stability was small, but when the cation concentration was reduced to 40 meq/l a marked loss of stability occurred. However, the rate of decrease of the pore size index, as the cation concentration was reduced in this SAR 40 solution, was much greater for the red brown soil than for the alluvial soil (Table 5). Further, as the cation concentration of this solution was reduced, progressively the volumes of the most common pores in both soils decreased (Appendix D) and these common pores also tended to drain over a larger pressure head range (Figures 15 and 19). With solutions of SAR 40 and cation concentration of 10 meq/l, the moisture release curves of alluvial and red brown soils did not have an inflexion at pressure heads greater than -60 cm of water and they also showed pronounced swelling and loss of shape and visual identity of aggregates.

At intermediate values of sodium absorption ratio (SAR 20 and SAR 10) and high cation concentration (160 meq/l) all soils showed a stable structure. But as the salt concentration was reduced in the SAR 20 solution, while the krasnozem soil retained its stability (Figure 10), the alluvial and red brown soils showed a shift of their moisture release curves to the right (Figures 14 and 18) indicating a decrease in size of inter-aggregate pores in these soils. These changes were qualitatively similar to those observed as the cation concentration was reduced in solutions of SAR 40. However, in both alluvial and red brown soils, equivalent losses in stability indicated by similar decreases in the pore size index (Table 5), equivalent shifts of the moisture release curves to the right and visible swelling occurred at relatively lower cation concentrations in this solution of SAR 20 than in the SAR 40 solution. Similarly, with salt solutions of SAR 10, as the

cation concentration was reduced, the moisture release curves of alluvial and red brown soils showed a progressive shift to the right (Figures 13 and 17), but the equivalent losses in stability in this SAR 10 solution occurred only at relatively much lower cation concentrations (Table 5) than for the SAR 20 and SAR 40 solutions.

The moisture release curves of alluvial and red brown soils for different salt solutions (Figures 12 to 19) could also be used to follow the changes in pore geometry of these soils as the SAR was increased at a given cation concentration, by plotting together the appropriate curves taken from the different figures. In general the resultant changes in moisture release curves, as the SAR was increased at a given cation concentration, appear to be qualitatively similar to the changes in moisture release curves caused by a reduction in cation concentration at a given SAR, namely a shift of the moisture release curves to the right indicating a reduction in size of pores due to swelling of the soil aggregates. This is illustrated by comparing the moisture release curves of red brown soil in solutions of cation concentration of 10 meq/l and SARs of 0, 10, 20 and 40. Such a comparison is possible only where the moisture release curves in the high salt solutions at all SARs are similar. In this respect, the use of the pore size index appears to have an advantage since it measures the changes in common pore size in lower salt solution relative to the size of common pore in salt solutions of the highest cation concentration at the same SAR. In this way it could account, at least partly, for small differences in the initial pore geometry in the high salt solutions, due to variations in packing. The relative shapes of the moisture release curves indicate that at least some of the

common pores in the low salt solutions were originally the common pores in the solutions of the highest cation concentration (e.g. Figure 17).

The values of the pore size indices and the relative changes in moisture release curves in salt solutions indicate that the loss of stability as the cation concentration was reduced or the SAR was increased, was greater for the red brown soil than for the alluvial soil, while the krasnozem soil was very stable in all salt solutions.

Previous studies of Quirk and Schofield (1955) and McNeal and Coleman (1966) have demonstrated that the hydraulic conductivities of soils containing smectite clays showed a pronounced decrease, as the cation concentration was reduced in solutions of a given SAR. These results are consistent with the data from the present study in which the moisture release curves of the alluvial and red brown soils show a shift to the right, indicating a decrease in size of pores with reduction in cation concentration at a given SAR. The similarity in the shape of the moisture release curves of each soil in the highest salt solution at each SAR was in accord with the previous studies of McNeal and Coleman (1966) which showed that the hydraulic conductivities of a soil in such high salt solutions were similar, and varied by less than 10% from an average value. This also shows that by careful subsampling, packing and wetting the confined soil column slowly, separately packed soil columns of uniform bulk density (Collis-George and Bridge, 1973) and similar pore geometry could be obtained. It is also likely that during the process of leaching to obtain chemical equilibrium with salt solutions prepared from de-aerated water, most of the entrapped air was removed and thereby the variations in moisture release curves due to incomplete saturation were minimised.

(ii) Changes in volumes of pores of specific size ranges

The changes in volumetric moisture contents or undrained porosity at specific pressure heads in alluvial and red brown soils as the cation concentration was reduced in solutions with a fixed SAR are shown in Figures 20 to 25. The data for plotting these curves were derived from the moisture release curves of the soils in the same salt solutions. These curves can be used to obtain the moisture release curves in salt solutions of SAR 40, 20 and 10 and any intermediate value of cation concentration from 2.5 to 160 meq/l. Similarly, the moisture release curves at intermediate values of SAR can be obtained from curves showing the changes in volumetric moisture content at specific pressure heads, as the SAR was increased in solutions of a given cation concentration. Combining these two sets of curves it is possible to obtain, by means of graphical interpolation, the moisture release curves of the soils at any intermediate values of SAR from 0 to 40 and cation concentration from 2.5 to 160 meq/l.

These curves in Figures 20 to 25 may also be used to follow the changes in volume of pores of any particular size in the alluvial and red brown soils as the salt concentration is reduced in a solution of a given SAR. For any given salt solution the volumes of pores which drain within a specific pressure head range, corresponding to a particular pore size range, are given by the distance between the specific curves for the two pressure heads which defines the limits of the range. The volumes of all pore groups initially showed little change with reductions in cation concentration of solutions at a given SAR, until a threshold concentration was reached. The volumes of the largest pore group and some of the other large pore groups showed pronounced decreases with a further reduction in cation concentration of the solution

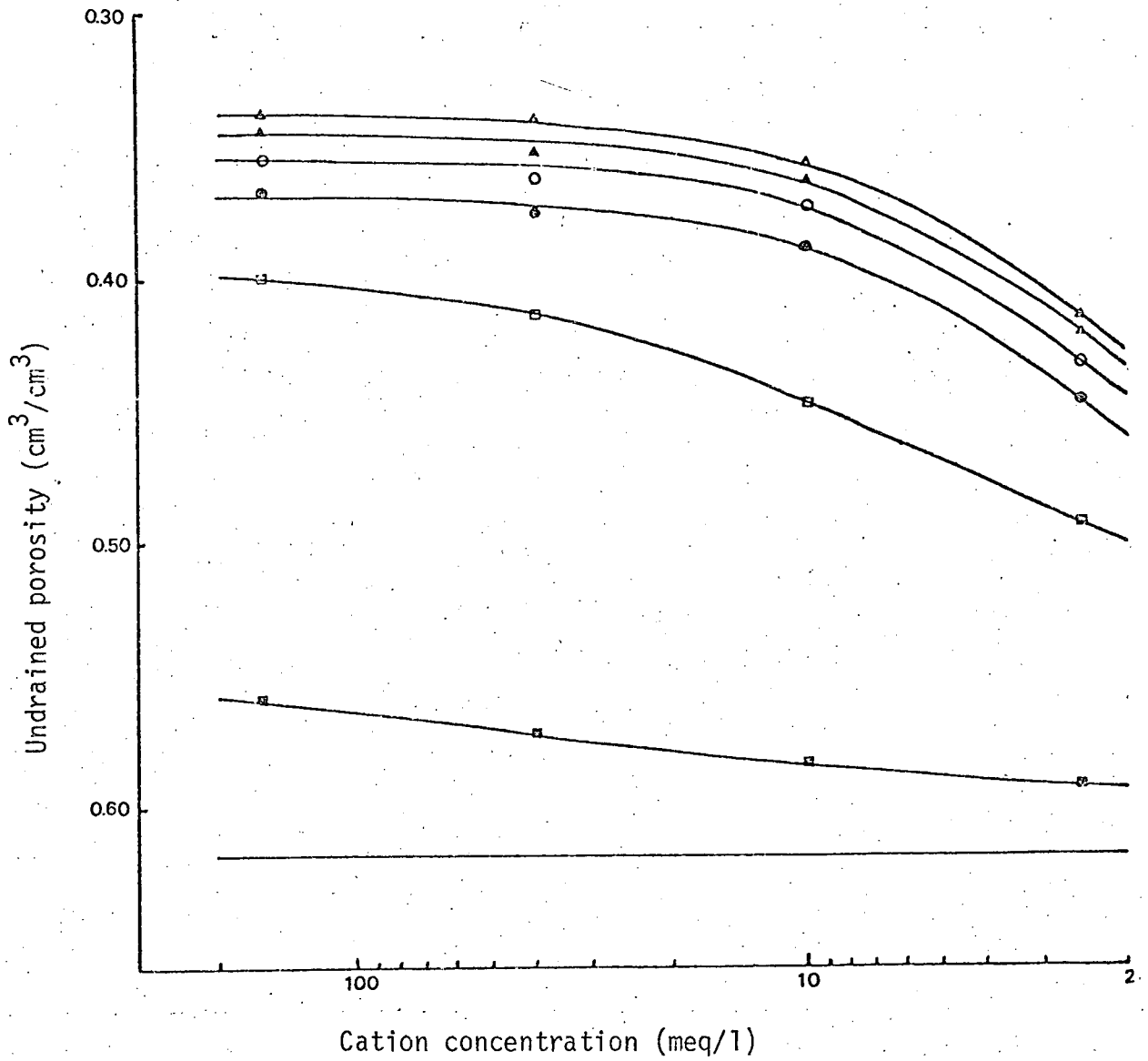


Fig. 20. Undrained porosity at specific pressure heads of ■-10, □-20, ●-30, ○-40, ▲-50 and △-60 cm of water in alluvial soil for salt solutions of SAR 10 and different cation concentrations. Bulk density of soil, 1.02 g/cm^3 .

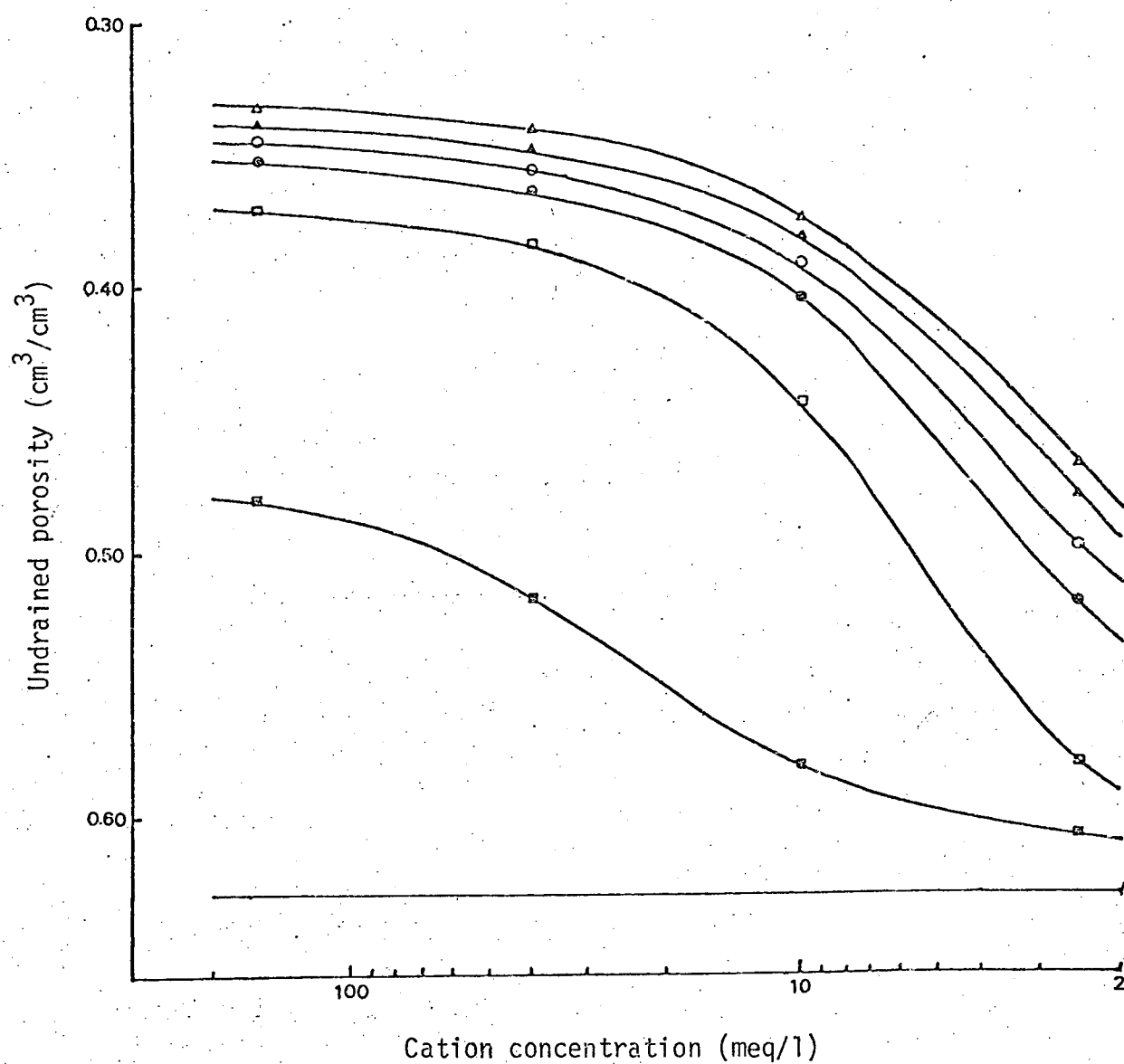


Fig. 21. Undrained porosity at specific pressure heads of ■-10, □-20, ●-30, ○-40, ▲-50 and △-60 cm of water in alluvial soil for salt solutions of SAR 20 and different cation concentrations. Bulk density of soil, 1.00 g/cm^3 .

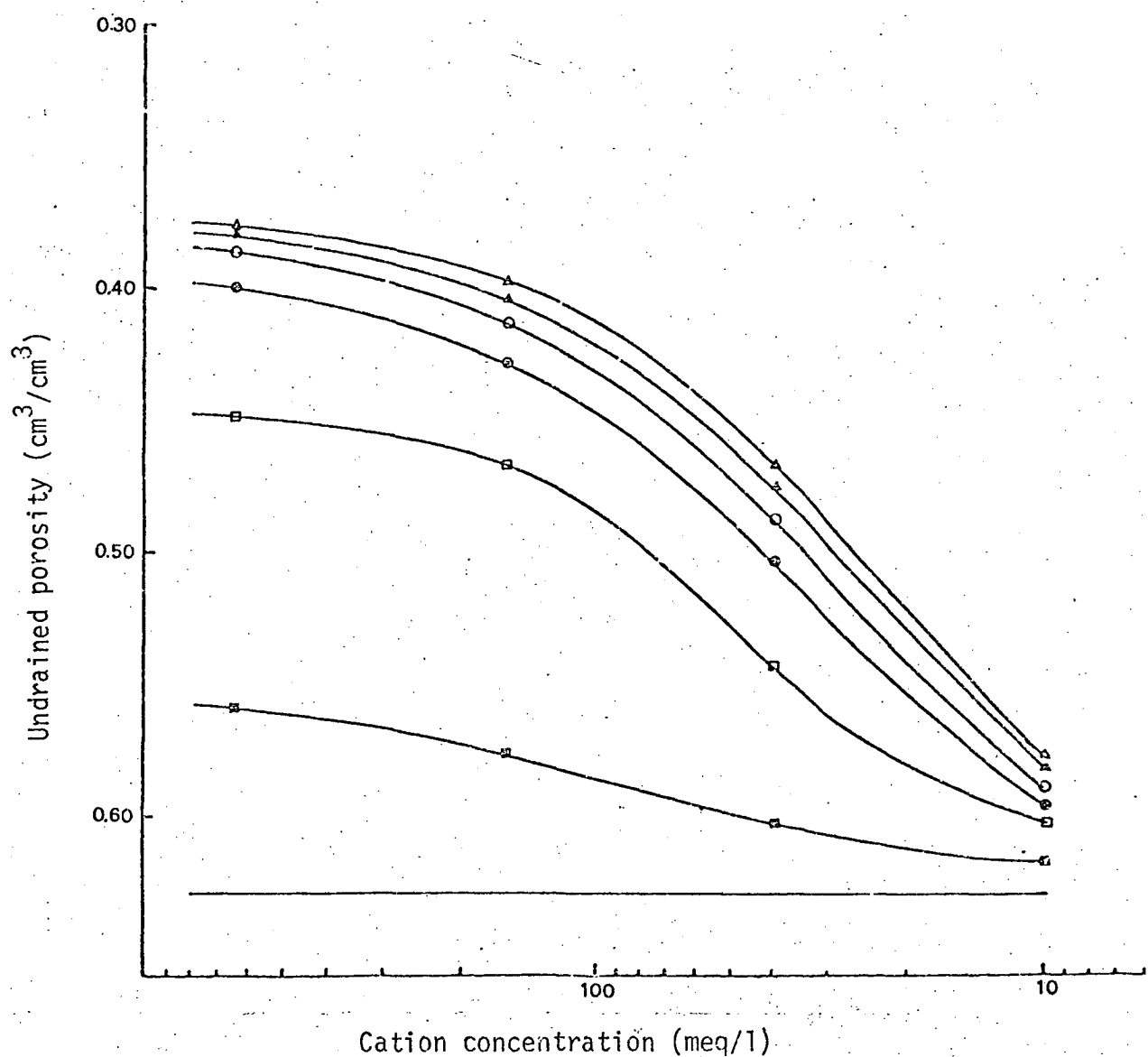


Fig. 22. Undrained porosity at specific pressure heads of \blacksquare -10, \square -20, \odot -30, \circ -40, \triangle -50 and \triangle -60 cm of water in alluvial soil for salt solutions of SAR 40 and different cation concentrations. Bulk density of soil, 1.00 g/cm^3 .

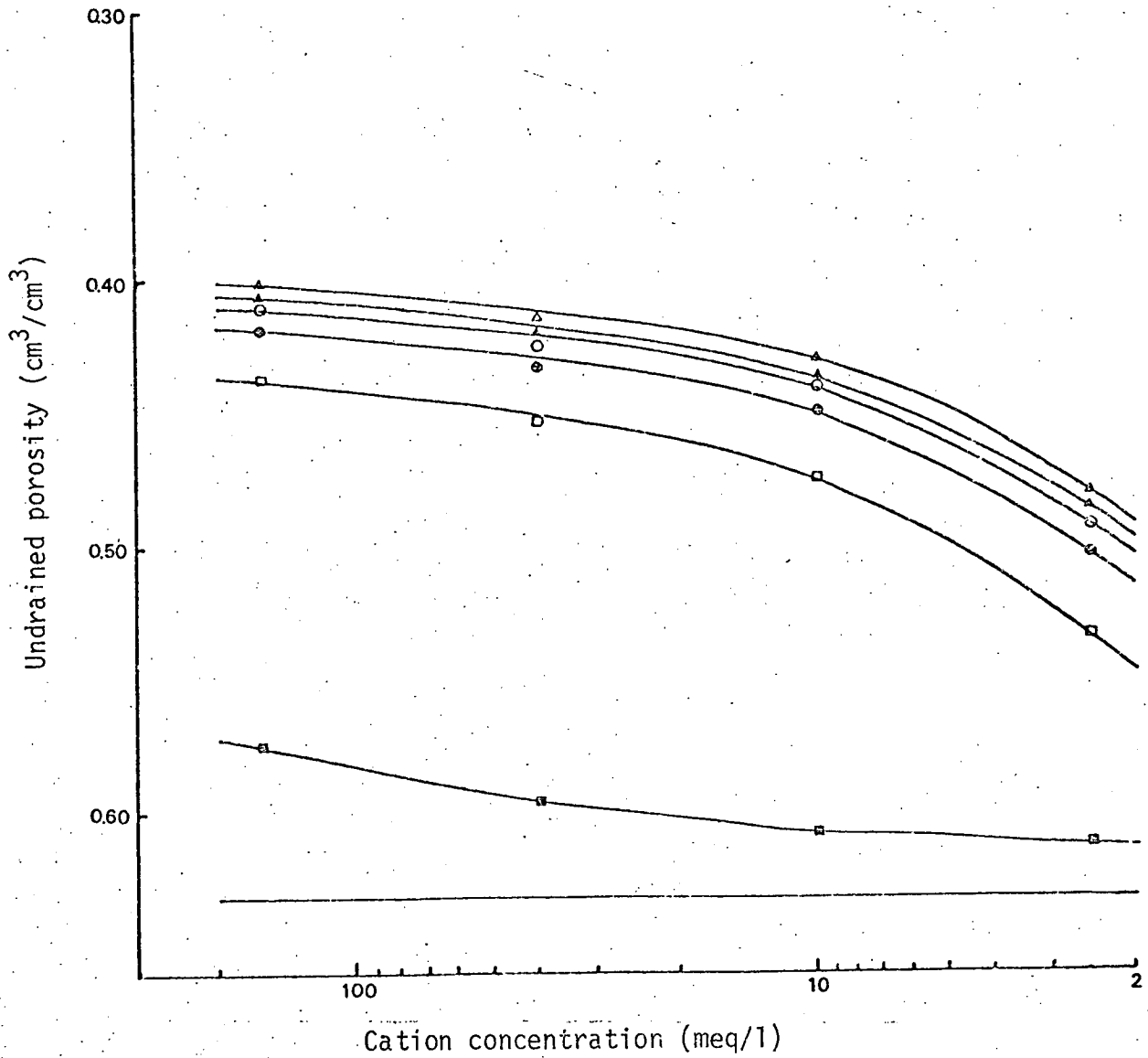


Fig. 23. Undrained porosity at specific pressure heads of ■-10, □-20, ⊙-30, ○-40, △-50 and ▲-60 cm of water in red brown soil for salt solutions of SAR 10 and different cation concentrations. Bulk density of soil, 1.07 g/cm^3 .

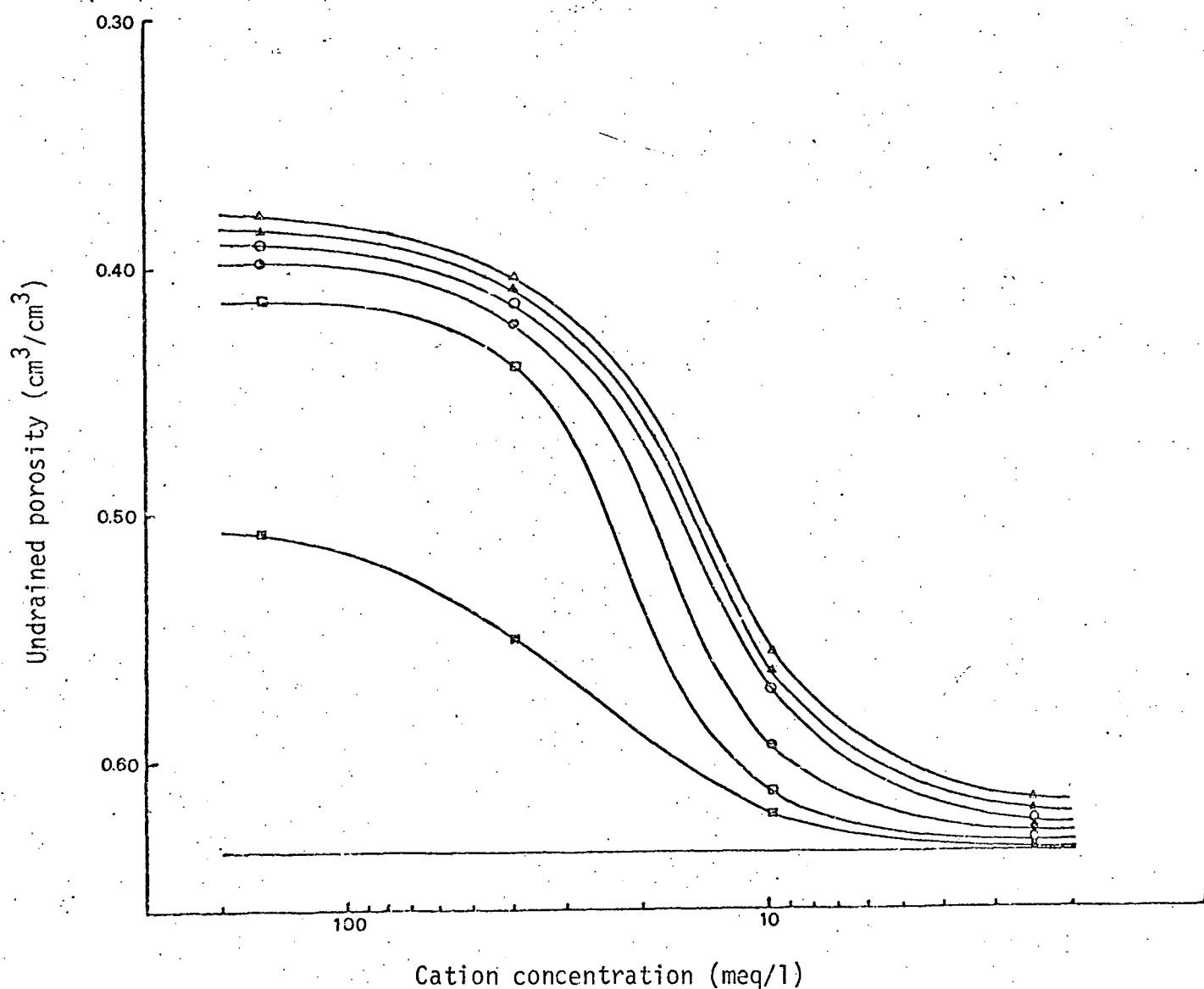


Fig. 24. Undrained porosity at specific pressure heads of ■-10, □-20, ●-30, ○-40, ▲-50 and △-60 cm of water in red brown soil for salt solutions of SAR 20 and different cation concentrations. Bulk density of soil, 1.06 g/cm^3 .

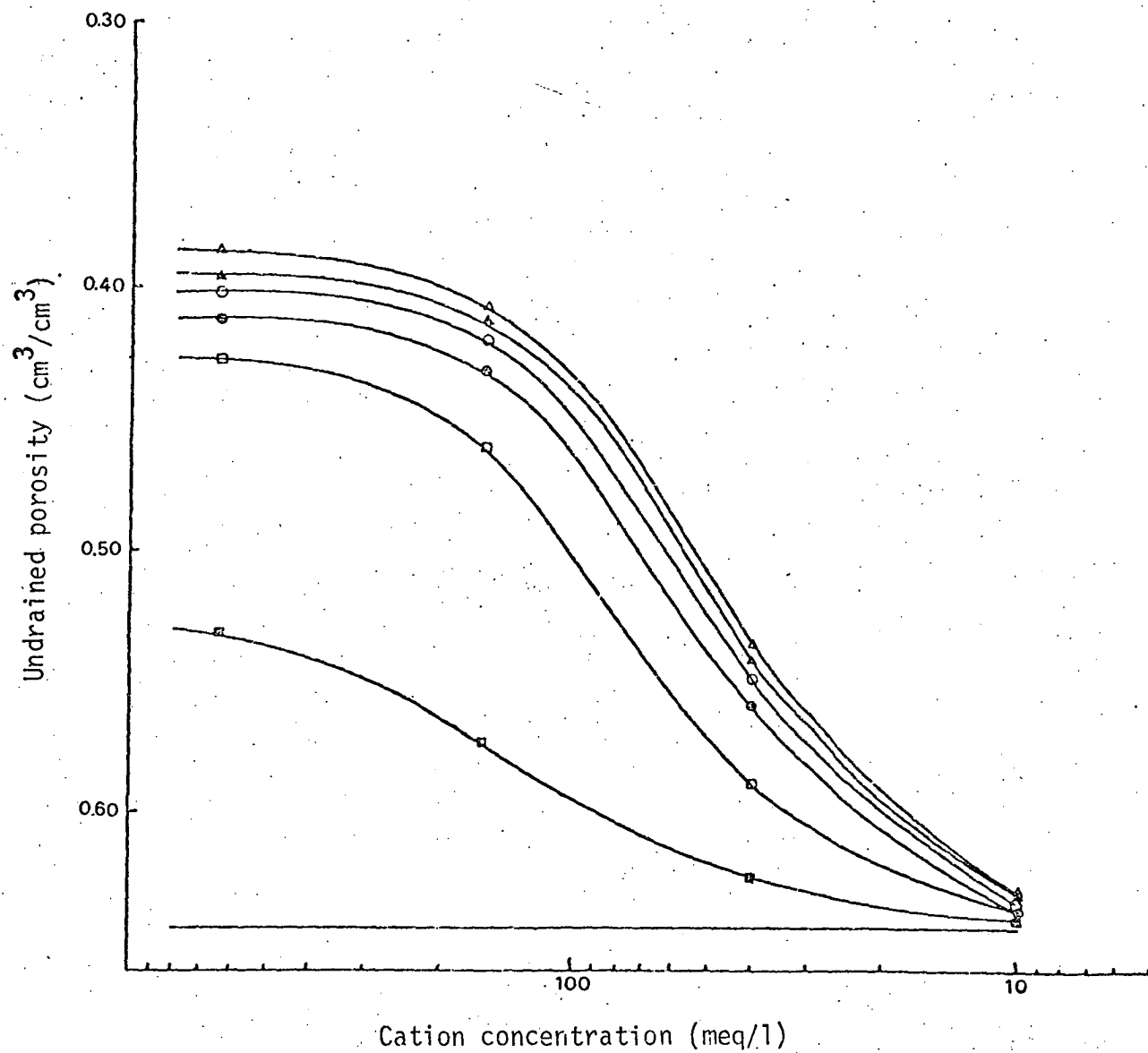


Fig. 25. Undrained porosity at specific pressure heads of ■-10, □-20, ●-30, ○-40, ▲-50 and △-60 cm of water in red brown soil for salt solutions of SAR 40 and different cation concentrations. Bulk density of soil, 1.04 g/cm^3 .

at any given SAR. On the other hand, although no experimental data is available, it can be theoretically envisaged that the volume of the smallest pore group will continue to show increases as the cation concentration is reduced below the threshold concentration, due to reduction in radii of larger sized pores, and similar changes are likely in other small sized pore groups. The changes in volumes of medium sized pore groups of the alluvial and red brown soil in different salt solutions showed some variability but in general they initially tended to show small increases as the cation concentration was reduced below the threshold concentration but with further reductions in cation concentration the volumes of these pore groups were very markedly decreased.

These patterns of change in volumes of large and medium sized pore groups were further examined using the moisture release data for the alluvial and red brown soils (Tables 6 and 7) packed to higher bulk densities of 1.13 and 1.20 g/cm³ respectively. A similar pattern of change of volume of these pores, where the large sized pore groups show a continuous decrease in volume and the medium sized pore groups initially show an increase to a maximum and then a decrease, was observed as the cation concentration of a solution of given SAR was reduced. It may also be seen that for both soils, as the average diameter of the particular intermediate sized pore group became smaller, the cation concentration at which the maximum pore volume was reached became lower.

These observations on the changes in volumes of pore groups may be explained in the following manner. The swelling of soil aggregates into pores, which in the high salt solution constitute a pore group of specific size, will reduce the diameter of some of these pores to groups of relatively smaller size pores. The

resultant decrease in volume of this specific group of pores will be balanced, to varying degrees, by an increase in volume of this pore group due to swelling of aggregates into pores which in high salt solutions were originally in a still larger size group. The relative magnitudes of these two processes will determine the nature and magnitude of the volume changes of that pore group, as the cation concentration is reduced in a solution of a given SAR. Hence we obtain the general pattern of changes in volumes of large and medium size pores. The data in Tables 6 and 7 include a few values which tend to differ from this general pattern, possibly due to variations in packing causing a distribution of pores different from the normal distribution in the solution of the highest cation concentration at a given SAR.

In this discussion of the changes in volume of specific size pores in different salt solutions, it must be borne in mind that the pores draining between specific pressure heads will have an average radius corresponding to the average pressure head applied. But the same pore will have a different radius when the soil is subjected to other pressure heads, if the soil aggregates show shrinkage on drying. Hence, in such soils the curves indicating the pore size distribution at different pressure heads will vary from each other and also from the moisture release curve (Appendix B) of the soil. This is likely to have a marked effect on the hydraulic conductivity characteristics of the soil, which will be examined in detail in later sections of this study.

In order to make meaningful quantitative comparisons of the effects of salt solutions on the pore geometry of a soil packed to different bulk densities, it is necessary to take into account

the differences in the initial pore size distribution in the presence of a high salt solution in which the soil is stable. This could be achieved by expressing the volumes of pores of a given pore size range in salt solutions with a lower cation concentration, as a percentage of the pore volume of the same size range in the presence of solutions of the highest cation concentration at the same SAR (Tables 8 and 9). In this analysis the changes in the total volume of pores larger than specific radii has been considered. It is seen that for both alluvial and red brown soils packed to a higher bulk density, all the pore groups analysed, except the largest sized pore group, show a proportionately bigger percentage decrease in volume in the low salt solution at a given SAR, than the same soil packed to a lower bulk density in the presence of the same low salt solution. Apparently the greater amount of clay per unit volume of soil packed to a high bulk density results in greater swelling which thus leads to greater percentage decrease in volume of specific pore groups. However, although the total amount of swelling per unit volume is less in soils of lower bulk density, apparently most of this swelling of the soil aggregates is accommodated in the more abundant pores of the largest size group, resulting in a relatively greater percentage decrease in volumes of these large sized pores in the soil packed to a lower bulk density.

Although the bulk density was changed from 1.01 to 1.13 g/cm³ in the alluvial soil and from 1.05 to 1.20 g/cm³ in the red brown soil, the values for the fractional decrease of volumes of pores of specific sizes in low salt solution for each soil packed to two different bulk densities was generally well within 0.1 of each other (Tables 8 and 9). It can be expected that for smaller bulk density changes the differences in values will be

Table 6. Moisture release data of alluvial soil for salt solutions of SAR 40, 20 and 10 and different cation concentrations. Bulk density of soil, 1.13 g/cm^3 .

SAR	Suction	Solution Outflow (cm^3)						
		Cation Concentration (meq/l)						
		640	160	80	40	20	10	2.5
40	0 - 10	4.20	3.40	2.58	1.92	1.06	..	0.30
	10 - 20	6.32	5.04	4.22	3.12	1.12	..	
	20 - 30	4.32	3.96	3.48	2.24	1.26	..	
	30 - 40	1.42	1.26	<u>1.42</u>	1.22	<u>1.30</u>	..	0.28
	40 - 60	0.92	<u>1.14</u>	<u>1.16*</u>	1.08	1.00	..	0.40
	60 - 80	0.46	<u>0.68</u>	<u>0.72</u>	0.58	<u>0.80*</u>	..	0.30
	80 - 100	0.36	<u>0.48</u>	0.42	<u>0.54</u>	<u>0.66*</u>	..	0.44
	100 - 120	0.26	<u>0.34</u>	<u>0.38</u>	<u>0.40</u>	<u>0.46*</u>	..	
	120 - 160	0.40	<u>0.66</u>	0.56	<u>0.76</u>	<u>0.78*</u>	..	1.04
20	0 - 10		4.12	..	3.16	2.50	1.82	
	10 - 20		6.18	..	6.14	6.02	4.24	2.72
	20 - 30		3.82	..	<u>4.48*</u>	3.18	<u>4.08</u>	
	30 - 40		1.34	..	<u>1.42</u>	<u>1.70*</u>	1.32	4.52
	40 - 60		1.14	..	1.06	<u>1.34</u>	<u>1.46*</u>	1.28
	60 - 80		0.64	..	<u>0.66</u>	<u>0.74*</u>	0.64	0.66
	80 - 100		0.44	..	<u>0.44</u>	<u>0.50</u>	0.48	<u>0.54*</u>
	100 - 120		0.24	..	<u>0.30</u>	<u>0.32</u>	<u>0.34</u>	
	120 - 160		0.46	..	0.42	<u>0.60</u>	<u>0.64</u>	<u>1.08*</u>
10	0 - 10		3.78	..	3.10	..	2.72	1.84
	10 - 20		5.88	..	5.72	..	5.26	4.24
	20 - 30		4.32	..	<u>4.88*</u>	..	3.46	3.02
	30 - 40		2.28	..	1.96	..	<u>2.26</u>	1.64
	40 - 60		1.52	..	<u>1.54</u>	..	<u>1.62*</u>	1.52
	60 - 80		0.76	..	<u>0.76</u>	..	0.70	<u>0.74</u>
	80 - 100		0.40	..	<u>0.46</u>	..	<u>0.48*</u>	0.40
	100 - 120		0.34	..	<u>0.38</u>	..	<u>0.40*</u>	0.36
	120 - 160		0.48	..	0.42	..	<u>0.50*</u>	0.48

*Cation concentrations at which the specific sized pores reach their maximum volume. The values underlined indicate where the volume of these specific sized pores has increased as the cation concentration was reduced from the previous higher concentration at the same SAR.

Table 7. Moisture release data of red brown soil for salt solutions of SAR 40, 20 and 10 and different cation concentrations. Bulk density of soil, 1.20 g/cm³.

SAR	Suction	Solution Outflow (cm ³)						
		Cation Concentration (meq/l)						
		640	160	80	40	20	10	2.5
40	0 - 10	2.21	1.52	0.96	0.52	0.16		
	10 - 20	1.79	1.44	0.76	0.48	0.18		
	20 - 30	1.04	<u>1.22*</u>	<u>1.22</u>	0.48	0.20		
	30 - 40	1.02	<u>1.12</u>	<u>1.70*</u>	0.44	0.16		
	40 - 60	0.70	<u>0.84*</u>	0.80	0.46	0.24		
	60 - 80	0.58	<u>0.70</u>	0.62	<u>1.18*</u>	0.30		
	80 - 100	0.32	<u>0.46*</u>	0.32	<u>0.42</u>	0.20		
	100 - 120	0.26	0.22	<u>0.26</u>	<u>0.28*</u>	0.20		
	120 - 160	0.32	<u>0.38</u>	0.34	<u>0.42</u>	<u>0.64*</u>		
20	0 - 10		2.14	..	1.34	1.00	0.28	0.16
	10 - 20		2.96	..	2.90	1.28	0.36	
	20 - 30		1.82	..	1.50	<u>1.90*</u>	0.26	
	30 - 40		0.74	..	0.42	<u>0.90</u>	<u>1.04*</u>	0.18
	40 - 60		0.70	..	<u>0.76</u>	<u>0.86*</u>	0.72	0.34
	60 - 80		0.44	..	<u>0.54*</u>	0.46	0.42	0.34
	80 - 100		0.24	..	<u>0.28</u>	<u>0.28</u>	<u>0.28</u>	0.10
	100 - 120		0.18	..	<u>0.22</u>	<u>0.22</u>	<u>0.26*</u>	
	120 - 160		0.24	..	<u>0.30</u>	<u>0.30</u>	<u>0.34*</u>	0.20
10	0 - 10		2.74	..	2.00	..	1.48	1.18
	10 - 20		2.04	..	<u>2.22*</u>	..	1.78	1.06
	20 - 30		1.76	..	1.58	..	<u>1.78*</u>	1.06
	30 - 40		1.28	..	1.24	..	0.92	0.84
	40 - 60		1.10	..	0.90	..	<u>0.92</u>	0.62
	60 - 80		0.70	..	0.54	..	0.46	<u>0.52</u>
	80 - 100		0.30	..	0.26	..	<u>0.34</u>	<u>0.46*</u>
	100 - 120		0.20	..	<u>0.22</u>	..	<u>0.26</u>	<u>0.28*</u>
	120 - 160		0.26	..	<u>0.34</u>	..	0.32	<u>0.36*</u>

*Cation concentrations at which the specific sized pores reach their maximum volume. The values underlined indicate where the volume of these specific sized pores has increased as the cation concentration was reduced from the previous higher concentration at the same SAR.

Table 8. Relative volumes* of pores of specific diameter ranges in alluvial soil in different salt solutions. Soils packed to bulk densities of 1.01 and 1.13 g/cm³.

Bulk Density (gm/cm ³)		1.01 ± 0.02							1.13						
SAR	Pore diameter (μm)	Cation Concentration (meq/l)							Cation concentration (meq/l)						
		640	160	80	40	20	10	2.5	640	160	80	40	20	10	2.5
40	> 300	1.00	0.73	0.54	0.35	0.24	0.15		1.00	<u>0.81</u>	<u>0.62</u>	<u>0.46**</u>	0.26
	> 150	1.00	0.91	0.73	0.47	0.27	0.14		1.00	0.81	0.65	0.48	0.21	..	0.03
	> 100	1.00	0.87	0.74	0.55	0.33	0.15		1.00	0.84	0.70	0.49	0.24
	> 75	1.00	0.87	0.77	0.57	0.36	0.17		1.00	0.84	0.72	0.53	0.30	..	0.04
	> 50	1.00	0.92	0.82	0.64	0.42	0.20		1.00	0.87	0.75	0.56	0.34	..	0.06
20	> 300		1.00	..	0.72	0.50	0.31	0.15		1.00	..	<u>0.77</u>	<u>0.61**</u>	<u>0.45**</u>	..
	> 150		1.00	..	0.95	0.88	0.71	0.19		1.00	..	0.91	0.83	0.59**	<u>0.27</u>
	> 100		1.00	..	0.96	0.91	0.81	0.40		1.00	..	0.98	0.83	0.72	..
	> 75		1.00	..	0.97	0.92	0.83	0.47		1.00	..	0.99	0.87	0.75	0.47
	> 50		1.00	..	0.97	0.93	0.85	0.54		1.00	..	0.98	0.89	0.78	0.52
10	> 300		1.00	..	0.76	..	0.59	0.44		1.00	..	<u>0.82</u>	..	<u>0.72**</u>	<u>0.49</u>
	> 150		1.00	..	0.93	..	0.82	0.57		1.00	..	0.92	..	0.83	<u>0.63</u>
	> 100		1.00	..	0.97	..	0.92	0.68		1.00	..	0.98	..	0.82	0.65
	> 75		1.00	..	0.97	..	0.93	0.71		1.00	..	0.97	..	0.85	0.66
	> 50		1.00	..	0.97	..	0.94	0.72		1.00	..	0.97	..	0.87	0.69

*The volume of pores of specific diameter in salt solutions of lower cation concentration at a given SAR, expressed as a fraction of their volume in a solution of the highest cation concentration at the same SAR.

**The differences in the corresponding sets of values in the soil packed to the two bulk densities is larger than 0.10. The values underlined show an increase of more than 0.02 over the corresponding values in the soil packed to lower bulk density.

Table 9. Relative volumes* of pores of specific diameter ranges in red brown soil in different salt solutions. Soils packed to bulk densities of 1.05 and 1.20 g/cm³.

Bulk Density 9gm/cm ³)		1.05 ± 0.02							1.20						
SAR	Pore diameter (µm)	Cation concentration. (meq/l)							Cation concentration (meq/l)						
		640	160	80	40	20	10	2.5	640	160	80	40	20	10	2.5
40	> 300	1.00	0.63	0.35	0.16	0.08	0.04		1.00	<u>0.69</u>	<u>0.44</u>	<u>0.24</u>	0.08		
	> 150	1.00	0.84	0.54	0.25	0.12	0.04		1.00	0.74	0.43**	0.25	0.09		
	> 100	1.00	0.91	0.67	0.36	0.16	0.04		1.00	0.83	0.59	0.30	0.11		
	> 75	1.00	0.93	0.70	0.39	0.19	0.04		1.00	0.88	0.60	0.32	0.12		
	> 50	1.00	0.91	0.74	0.41	0.20	0.05		1.00	0.91	0.66	0.36	0.14		
20	> 300		1.00	..	0.66	0.39	0.11	0.02		1.00	..	0.63	<u>0.47</u>	0.13	..
	> 150		1.00	..	0.89	0.46	0.10	0.02		1.00	..	0.84	<u>0.45</u>	<u>0.13</u>	0.04
	> 100		1.00	..	0.90	0.62	0.18	0.03		1.00	..	0.83	0.61	0.13	..
	> 75		1.00	..	0.91	0.68	0.27	0.05		1.00	..	0.81	0.67	0.26	0.05
	> 50		1.00	..	0.90	0.70	0.32	0.08		1.00	..	0.83	0.71	0.32	0.09
10	> 300		1.00	..	0.64	..	0.41	0.34		1.00	..	<u>0.73</u>	..	<u>0.54**</u>	<u>0.43</u>
	> 150		1.00	..	0.91	..	0.81	0.51		1.00	..	0.89	..	<u>0.69**</u>	0.47
	> 100		1.00	..	0.93	..	0.86	0.60		1.00	..	0.89	..	0.77	0.51
	> 75		1.00	..	0.92	..	0.86	0.63		1.00	..	0.90	..	0.76	0.53
	> 50		1.00	..	0.94	..	0.87	0.64		1.00	..	0.89	..	0.78	0.54

*The volume of pores of specific diameter in salt solutions of lower cation concentration at a given SAR, expressed as a fraction of their volume in a solution of the highest cation concentration at the same SAR.

**The differences in the corresponding sets of values in the soil packed to the two bulk densities is larger than 0.10. The values underlined show an increase of more than 0.02 over the corresponding values in the soil packed to lower bulk density.

smaller. On the basis of these results, it is tentatively suggested that the experimentally determined values of fractional decreases in volumes of pores of specific sizes in a soil packed to a given bulk density (Tables 8 and 9) as the salt concentration is reduced at a given SAR, could be used as a frame of reference to compute the moisture release curves in the presence of the same salt solutions for the same soil packed to another bulk density, provided that the bulk density differences are not too large. In making these computations, the pore size distribution of the soil packed to this new bulk density in the presence of a high salt solution should be known.

Although the data in Tables 8 and 9 also indicate that the fractional decrease in the largest sized pore group in specific low salt solutions in a particular soil packed to two different bulk densities sometimes show differences exceeding 0.1, these larger differences occur in the part of the moisture release curves in which drainage of substantially large pore volumes takes place over a narrow pressure head range. Therefore, these variations may not be very significant, as differences of similar magnitude can occur at these high pressure heads in duplicate soil columns of the same bulk density in the same salt solution due to slight differences in packing and errors in experimental measurements. Whether a variation of the magnitude of 0.1 is too large to be acceptable will depend on the proposed uses of the moisture release curves computed by this method. In most studies, particularly in the field, where very accurate determinations of moisture release curves are generally not necessary on account of the large variations in field soil characteristics, differences of this magnitude may not be significant, especially in view of the

advantages of this method for such studies. However, it must be stressed that this method has been tested for only two soils and three series of salt solutions and further evaluation of the general use of this method would be necessary.

(iii) Equivalent salt solutions

A concept of equivalent salt solutions was developed during this study. It appears to have a number of important applications to studies which involve flow of salt solutions in soils. The theoretical basis for this concept may be explained in the following manner. Previous studies have shown that the extent of clay swelling increases with a reduction in cation concentration at any given SAR value. For a given extent of swelling produced by reductions in cation concentration to specific values in solutions of high SAR, there are at progressively lower SARs, progressively lower cation concentrations which produce the same extent of swelling. These combinations of SAR and cation concentrations producing a given extent of swelling would constitute an equivalent salt solution series. Similarly there would be other series of equivalent salt solutions producing other extents of swelling. Therefore, if the initial pore geometry in the high salt solution at each SAR to which the soil is stable is identical, and since the introduction of equivalent salt solutions is likely to result in an equivalent change in pore geometry, the soil in equivalent salt solutions could be expected to have identical moisture release curves.

If the soils have identical moisture release curves in the presence of equivalent salt solutions they could theoretically be expected to have the same saturated hydraulic conductivity as well as identical unsaturated conductivity-pressure head curves for such

solutions. Hence the flow of equivalent salt solutions in a given soil under any specific boundary conditions could also be expected to be identical.

Therefore, if the values of the equivalent salt solutions are known, and the changes in saturated and unsaturated conductivities of a soil as the cation concentration is reduced at any given SAR are also known, it should be possible to predict values for saturated and unsaturated conductivity for solutions of any other combination of SAR and cation concentration, using the values of the equivalent salt solution as a frame of reference. Using a similar procedure it could also be possible to directly correlate flow of equivalent salt solutions under the same boundary conditions. Applying this concept to the present study, if the capacity of a soil to transmit salt solutions from a water table to the surface is studied in detail for a series of salt solutions of decreasing cation concentrations at a given SAR, it should be possible to extrapolate these results to other series of salt solutions of decreasing cation concentrations at other SARs, using the values for the equivalent salt solutions as a frame of reference. This theoretical concept of equivalent salt solution was tested using the values of saturated and unsaturated conductivities for salt solutions measured during the present study.

The "equivalent salt solutions for swelling" may be defined as different combinations of SAR and cation concentrations of salt solutions in which soils have similar moisture release curves.

The "equivalent salt solutions for swelling" of alluvial and red brown soils can be obtained by comparing the moisture release curves of these soils in different salt solutions (Figures 12 to 19).

However, instead of the moisture release curves, the values of pore size indices were used to obtain the equivalent salt solutions.

Theoretically it could be argued that if the initial pore geometry of the soil is the same, the equivalent salt solutions, while producing identical changes in the size of the common pore, could also be expected to produce identical changes in all other pores. Under such conditions the value of pore size index could be expected to characterise the shape of the entire moisture release curve. However, in practice, duplicate columns of a given soil packed to the same bulk density are likely to show some variations in initial pore geometry. The pore size index, as it measures the relative swelling in lower salt solutions, could partly account for small differences in initial pore geometry.

The moisture release data of alluvial and red brown soils packed to bulk densities of 1.13 and 1.20 g/cm³ respectively for different salt solutions (Tables 6 and 7) were plotted and the pore size indices were obtained using the procedure described previously. The changes in values of pore size indices of these soils, as the cation concentration was reduced in solutions of given SARs of 40, 20 and 10, were plotted (Figures 26 and 27). These curves were then used to obtain the equivalent salt solutions of alluvial and red brown soils (Figures 28 and 29) in which the pore size indices fell to specific values. (Figures 28 and 29 could also be used to obtain the "equivalent salt solutions for swelling" of these soils for intermediate values of SAR.) The use of these values of "equivalent salt solutions for swelling" of alluvial and red brown soils for predicting changes in hydraulic conductivities in salt solutions will be evaluated in later sections.

The values of pore size indices of alluvial and red brown soils, packed to bulk densities of 1.13 and 1.20 g/cm³ respectively in the presence of specific salt solutions (Figures 26 and 27) show a

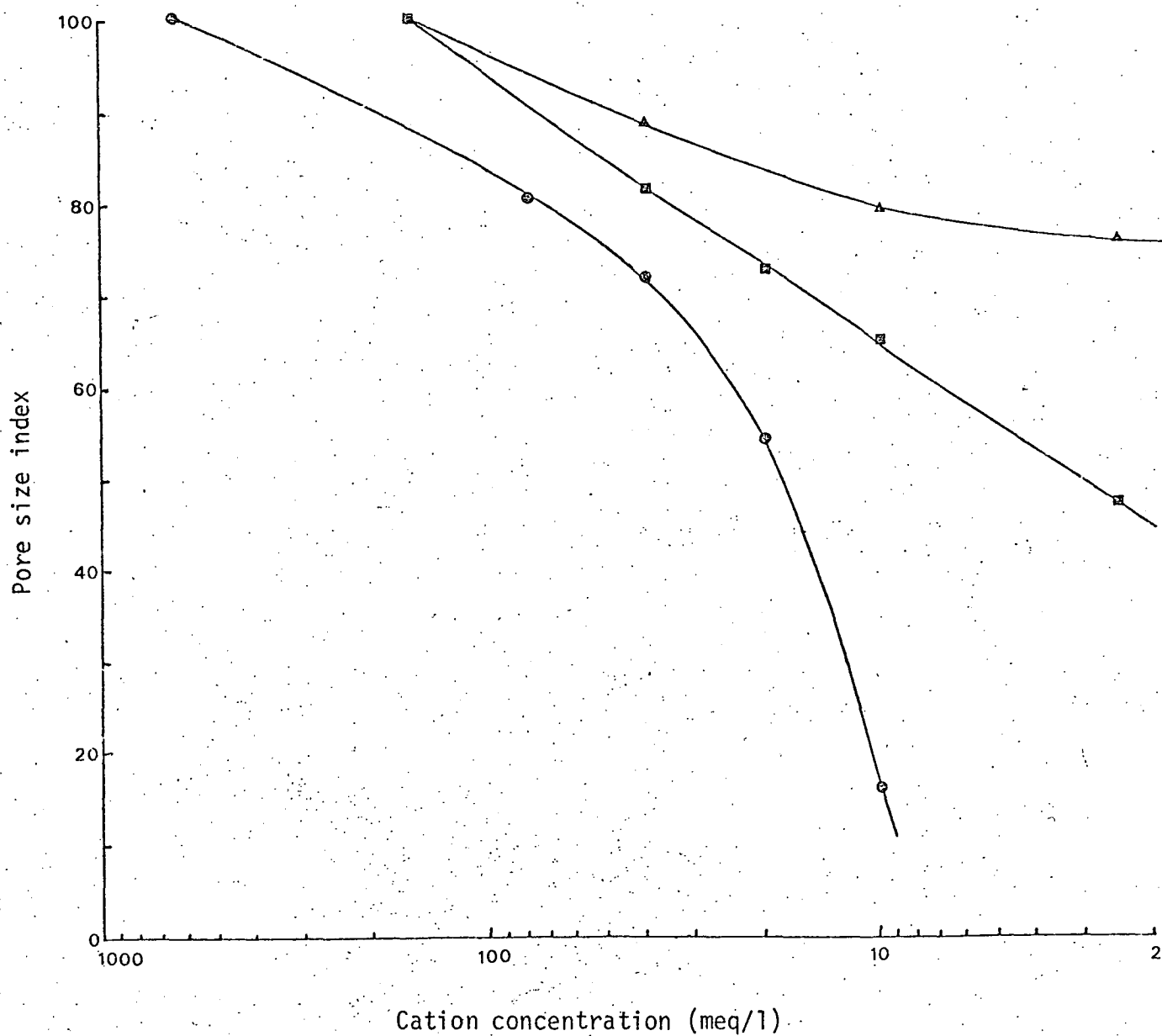


Fig. 26. Pore size indices of alluvial soil for salt solutions of SAR \odot 40, \square 20, \triangle 10, and different cation concentrations. Equivalent salt solutions producing a specific decrease in pore size index could be read from these curves by drawing lines parallel to the horizontal axis. Bulk density of soil, 1.13 g/cm^3 .

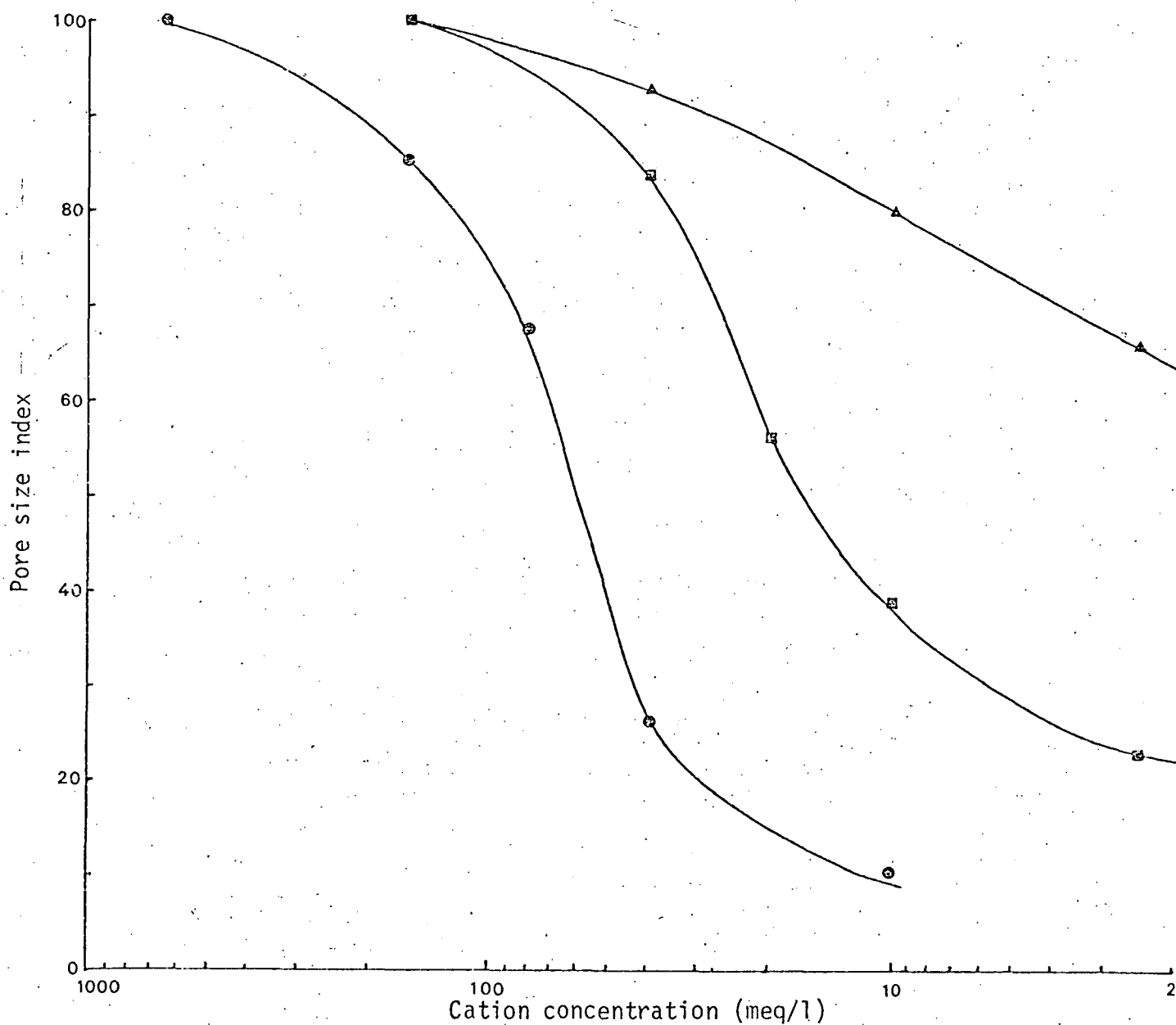


Fig. 27. The pore size indices of red brown soil for salt solutions of SAR \circ 40, \square 20, \triangle 10 and different cation concentrations. Equivalent salt solutions producing a specific decrease in pore size index could be read from these curves by drawing lines parallel to the horizontal axis. Bulk density of soil, 1.20 g/cm^3 .

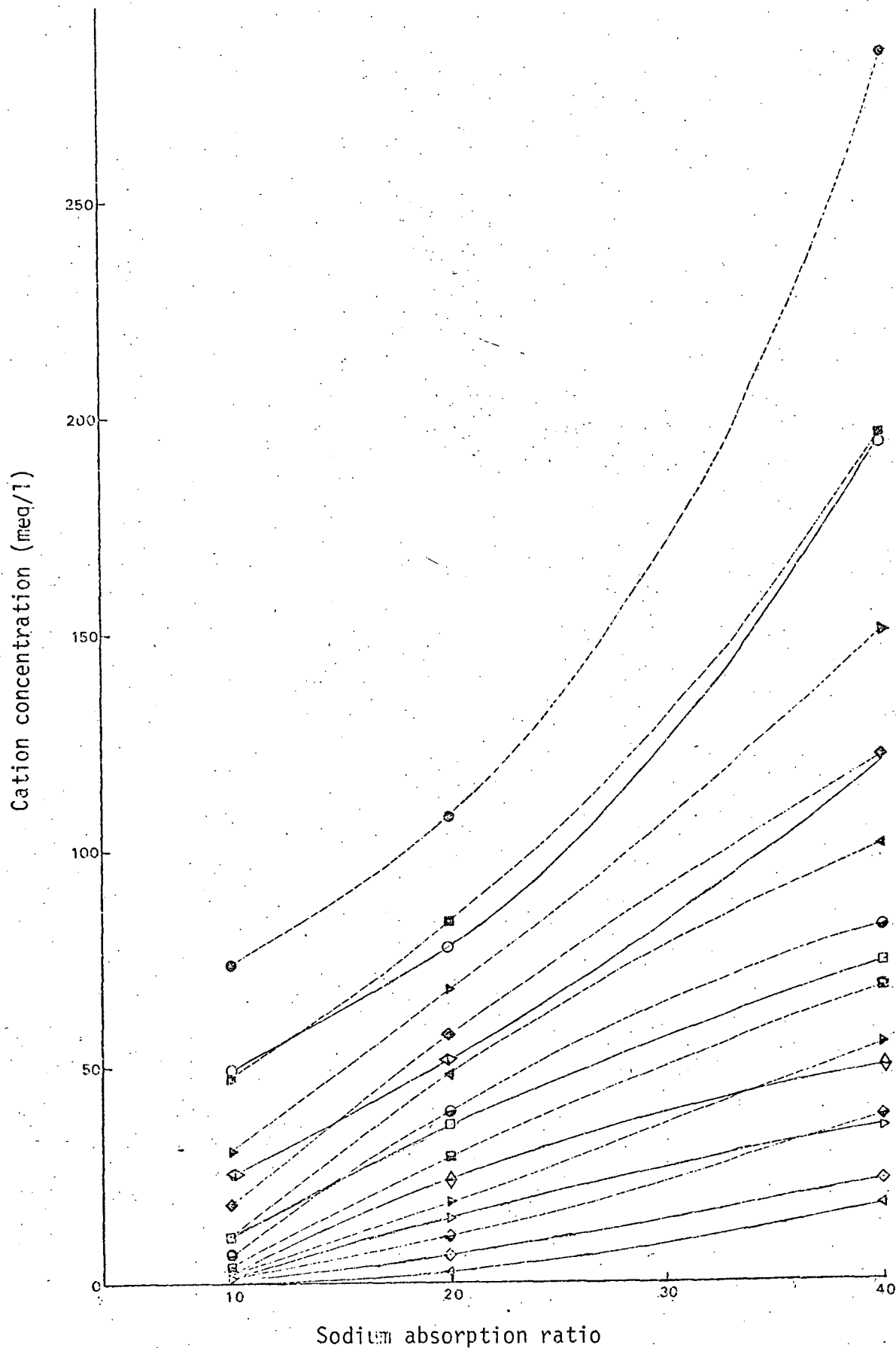


Fig. 28. Equivalent salt solutions for swelling and for saturated hydraulic conductivity in alluvial soil. Equivalent salt solutions for swelling causing a decrease in pore size index to ○ 0.9, ◐ 0.85, ◑ 0.8, ◒ 0.75, ◓ 0.7, ◔ 0.6, ◕ 0.5. Equivalent salt solutions causing a decrease in saturated conductivity to ⊙ 90%, ⊠ 80%, ⊡ 70%, ⊢ 60%, ⊣ 50%, ⊤ 40%, ⊥ 30%, ⊦ 20%, ⊧ 10% of its value in the highest salt solutions. Bulk density of soil, 1.13 g/cm³.

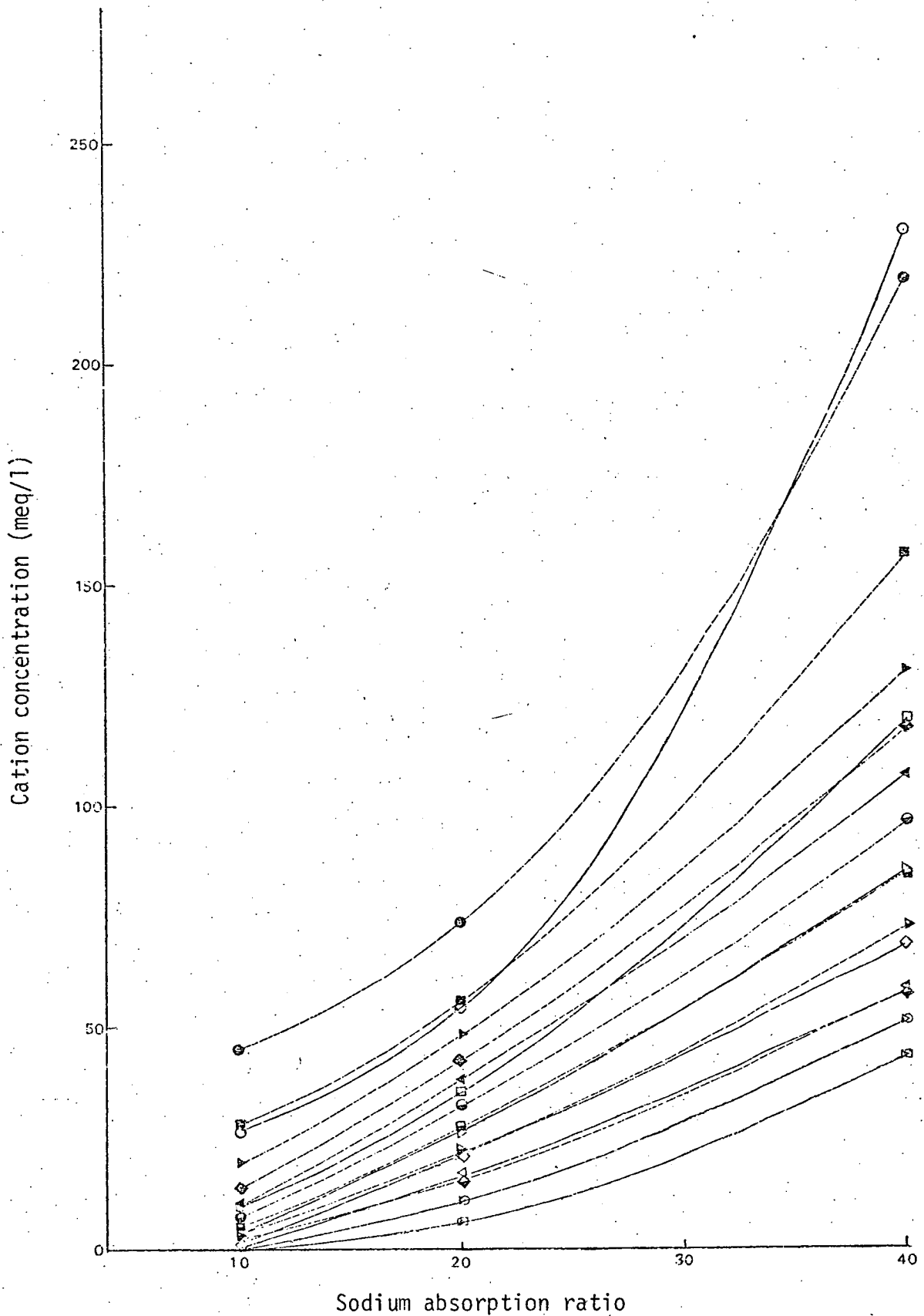


Fig. 29. Equivalent salt solutions for swelling and for saturated hydraulic conductivity in red brown soil. Equivalent salt solutions for swelling causing a decrease in pore size index to \circ 0.9, \square 0.8, \triangleright 0.7, \diamond 0.6, \triangleleft 0.5, \boxtimes 0.4, \boxminus 0.3. Equivalent salt solutions causing a decrease in saturated conductivity to \odot 90%, \blacksquare 80%, \blacktriangleright 70%, \blacklozenge 60%, \blacktriangleleft 50%, \odot 40%, \blacksquare 30%, \blacktriangleright 20%, \blacklozenge 10% of its value in the highest salt solutions. Bulk density of soil, 1.20 g/cm^3 .

general similarity to the corresponding values of pore size indices of these soils packed to lower bulk densities (Table 5). Theoretically it is possible for a given soil packed to markedly different bulk densities to have different pore size indices in specific low salt solutions, but these indices when plotted as described earlier could give the same equivalent salt solutions for that particular soil. However, some of the differences in values of pore size index observed for a given soil, packed to two different bulk densities, for a specific salt solution, could be partly due to experimental inaccuracies and variation in determining moisture release curves in different soil columns and in deriving the pore size indices from these curves.

A comparison of values in Figures 28 and 29 shows that the values of "equivalent salt solutions for swelling" for alluvial and red brown soils differ, even after allowing for the relatively greater swelling in the red brown soil.

C. Effect of Salt Solution Composition on Saturated Conductivity of the Soils

(i) Changes in saturated conductivities

The krasnozem soil packed to a bulk density of 1.00 g/cm^3 showed a saturated hydraulic conductivity of 2237 cm/day for a salt solution of SAR 20 and cation concentration of 160 meq/l (Table 10). When the cation concentration was reduced to 2.5 meq/l the saturated conductivity of this soil fell to 1875 cm/day. Thus, there was only a small change in saturated conductivity of the krasnozem soil as the cation concentration was reduced to low values in this solution of SAR 20. In the studies on changes of the moisture release curves in salt solutions (Figure 10), this soil showed very little swelling in low salt solutions of SAR 20. Therefore, the small changes in

saturated conductivity of this soil apparently indicate a low tendency for clay dispersion in the presence of low salt solutions of SAR 20, in addition to the low swelling of the soil aggregates.

The saturated hydraulic conductivities of alluvial and red brown soil packed to bulk densities of 1.13 and 1.22 g/cm³ respectively, for salt solutions of SAR 40, 20 and 10, as the salt concentration was progressively reduced from 640 or 160 to 2.5 meq/l in solutions of a specific SAR value, are given in Table 10. These saturated conductivity values are average values obtained from three separate soil columns. The average saturated conductivity for a given soil in the solution with the highest cation concentration at each SAR showed some differences, largely due to variations in the pore geometry of the separately packed soil columns used in these experiments. Hence, in order to make comparisons between the saturated conductivity values of a soil for low salt solutions of different SAR values obtained from these separately packed soil columns, the values of the saturated conductivity in solutions of the highest cation concentration at each SAR were averaged and all measured values of conductivity for the lower salt solutions were adjusted to correspond to this average value in the high salt solution, on a simple proportionate basis (McNeal and Coleman, 1966).

Figures 30 and 31 show the average adjusted saturated conductivity values of alluvial and red brown soils respectively in different salt solutions. In solutions of the highest cation concentration at all SARs the alluvial and red brown soils have average saturated conductivities of 235.7 and 37.3 cm/day respectively. With solutions of SAR 10, as the salt concentration was reduced the saturated conductivity of alluvial and red brown soils slowly decreased to values of 47.9 and 4.2 cm/day respectively for solutions

Table 10. The average saturated conductivity (cm/day) of krasnozem, alluvial and red brown soils packed to bulk densities of 1.00, 1.13 and 1.22 g /cm³ respectively, for salt solutions of SAR 40, 20 and 10 and different cation concentrations.

CATION CONCENTRATION (meq/l)	KRASNOZEM	ALLUVIAL			RED BROWN		
	SAR 20	40	SAR 20	10	40	SAR 20	10
640		251.2			36.3		
160	2237	180.5	230.4	225.5	29.6	35.4	40.2
80	..	97.9	9.44
40	2113	26.1	92.7	169.8	0.40	18.7	35.5
20	..	0.5	49.1	..	0.04	5.98	..
10	21.2	108.7	..	0.66	19.9
5	6.7
2.5	1875	..	1.7	45.8	4.4

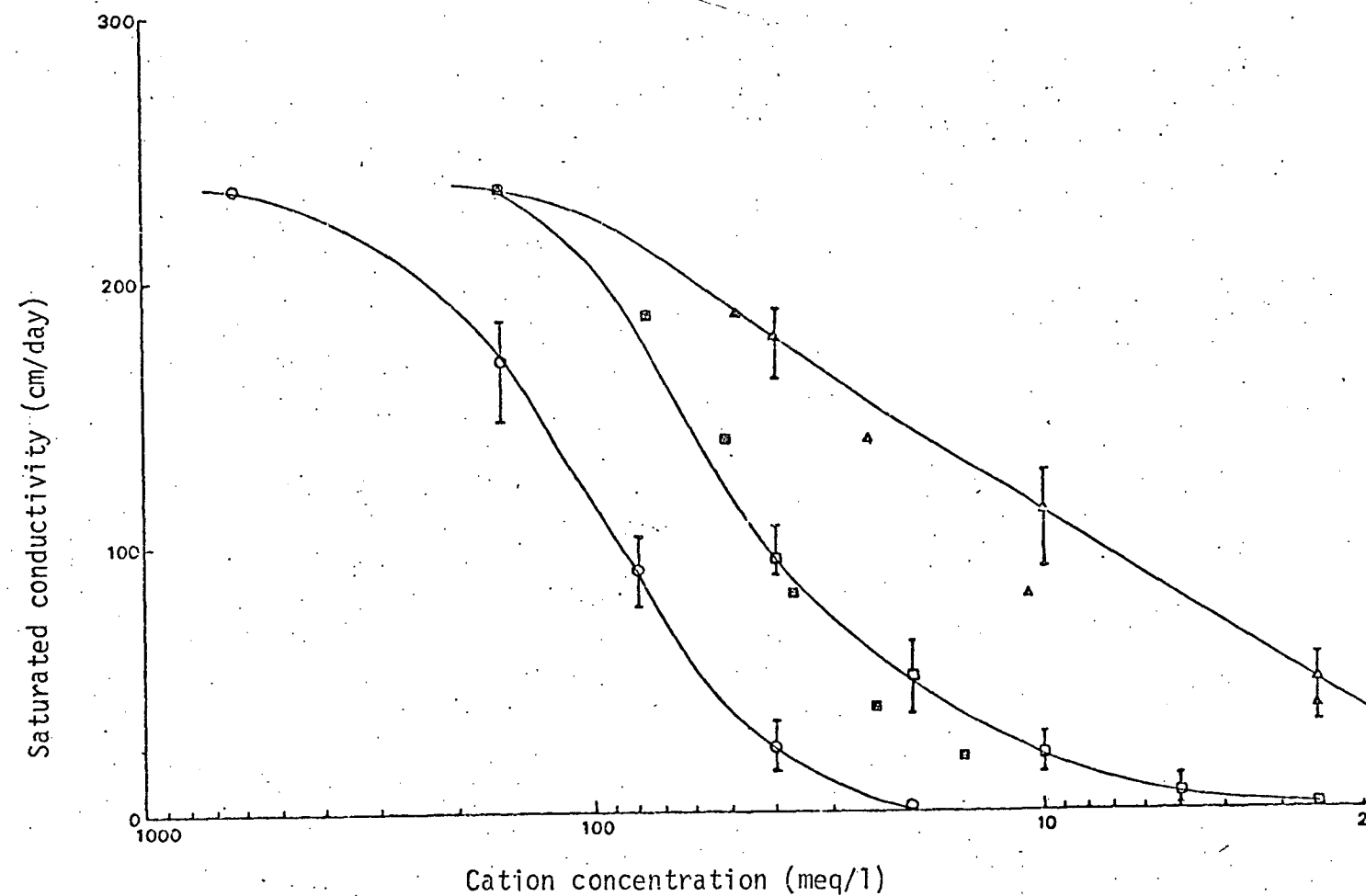


Fig. 30. Adjusted average experimental and predicted saturated conductivities of alluvial soil for different salt solutions. Experimental saturated conductivity for solutions of SAR \circ 40, \square 20, \triangle 10 and different cation concentrations. Predicted saturated conductivity for solutions of SAR \blacksquare 20, \blacktriangle 10 and different cation concentrations. Bulk density of soil, 1.13 g/cm^3 .

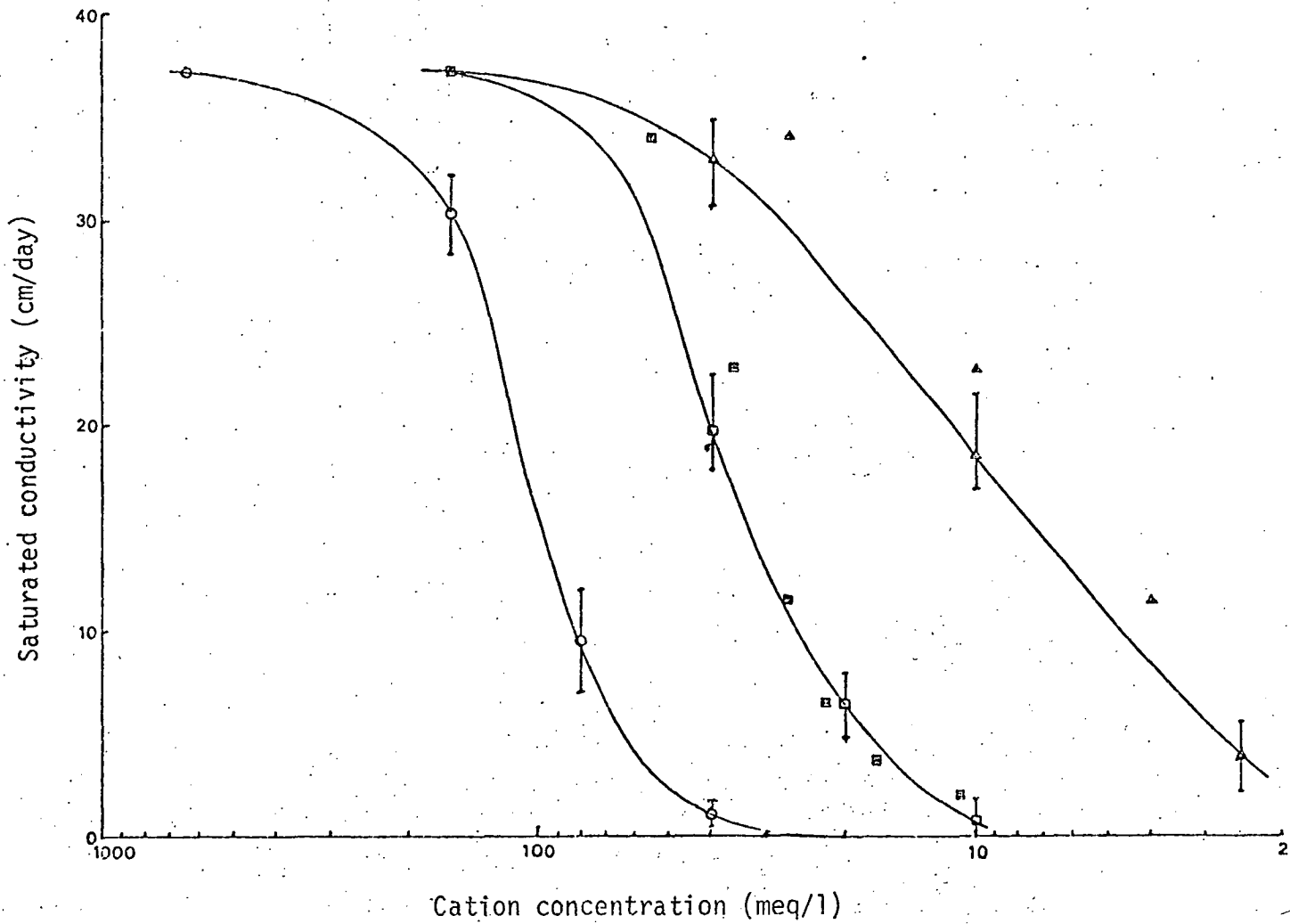


Fig. 31. Adjusted average experimental and predicted saturated conductivity of red brown soil for different salt solutions. Experimental saturated conductivity for solutions of SAR \circ 40, \square 20, \triangle 10 and different cation concentrations. Predicted saturated conductivity for solutions of SAR \blacksquare 20, \blacktriangle 10 and different cation concentrations. Bulk density of soil, 1.22 g/cm³.

whose cation concentration was 2.5 meq/l. In the case of the alluvial soil, as the cation concentration was reduced in solutions of SAR 40 and 20, the saturated conductivity decreased markedly to reach values of nearly zero at cation concentrations of 20 and 2.5 meq/l respectively. Very marked decreases in saturated conductivity were also observed in the red brown soil, as the cation concentration was reduced in solutions of SAR 40 and 20, to reach near zero values when salt concentration was 40 and 10 meq/l respectively. Hence, in both alluvial and red brown soils the rate of decrease in saturated conductivity with reduction in cation concentration was more rapid for solutions with the higher value of SAR. This is also indicated on a quantitative basis by the values of the experimentally determined average saturated conductivities of alluvial and red brown soils in low salt solutions, expressed as a fraction of the value of saturated conductivity for salt solutions of the highest cation concentration at the same SAR (Tables 11 and 12).

Tables 11 and 12 also show that the rate of decrease in saturated conductivity, with reduction in salt concentration in solutions at all SARs (40, 20 and 10) was more rapid for the red brown soil than for the alluvial soil. This result is in accord with the values of the pore size index of the soils in salt solutions, which indicate that the red brown soil was less stable than the alluvial soil in the presence of low salt, high sodium solutions. In a previous detailed study of a number of soils with a wide range of clay mineralogy, McNeal and Coleman (1966) found that, except for stable soils, the saturated conductivity decreased as the salt concentration was reduced in salt solutions of fixed SAR and this decrease was given by a curve of sigmoid shape. The alluvial and red brown soils used

in the present study, although showing differences in their absolute saturated conductivity values in low salt solutions, also gave curves of a sigmoid shape. There is qualitative agreement between the results from the present study and the results obtained by McNeal and Coleman (1966).

The relationship between the changes in experimentally determined saturated conductivity of the alluvial and red brown soils and the changes in pore geometry of the same soil, as the cation concentration was reduced at a given SAR, was examined on a quantitative basis. Several methods (Childs and Collis-George, 1950; Marshall, 1958; Millington and Quirk, 1959) have been suggested for computing the unsaturated conductivity-pressure head curves of soils from their moisture release curves. Recent studies (Jackson *et al.*, 1965; Kunze *et al.*, 1968) have shown that these methods could yield satisfactory predictions provided that a matching factor was used to match a computed value of unsaturated conductivity to an experimentally determined value at any single pressure head. The values of matching factors for different soils was found to vary considerably (Kunze *et al.*, 1968). However, if the value of the matching factor for a given soil at saturation remains constant in the presence of different salt solutions, the relative decrease in experimental saturated conductivity with reduction in cation concentration to a specific value at a given SAR would equal the relative decrease in computed saturated conductivity in the same low salt solutions. This suggestion was evaluated using the following procedure.

The saturated conductivities of alluvial and red brown soils for different salt solutions were computed from the moisture release curves in the specific salt solutions (Tables 6 and 7), using the equation proposed by Kunze *et al.* (1968). These computed saturated

conductivity values were then expressed as a fraction of the value of computed saturated conductivity for solutions of the highest cation concentration at the same SAR (Tables 11 and 12). These tables also show the experimentally measured saturated conductivity values for the low salt solutions expressed as a fraction of their value for the solution of highest cation concentration at the same SAR. A comparison of these sets of values showed that for both soils the fractional reduction in experimental saturated conductivities, as the cation concentration was reduced to a specific value at a given SAR, was generally greater than the corresponding fractional decrease in computed saturated conductivity values in the same low salt solution. This difference could be attributed to several possible causes.

Firstly, it is suggested that if the individual soil aggregates shrink on drying the curve indicating the actual pore size distribution of the soil at saturation will differ from the experimentally determined moisture release curves from which the computed saturated conductivity is obtained. A theoretical model to show the qualitative relationship between the moisture release curves and the curves indicating the pore size distribution at saturation and other pressure heads is described in Appendix B. The relative shift of the curve indicating the pore size distribution at saturation to the right of the moisture release curves becomes larger with increase in the extent of soil aggregate shrinkage on drying. Since the extent of shrinkage of aggregates is likely to increase with a reduction in cation concentration at a given SAR, the relative shift of the curve indicating the pore size distribution at saturation to the right of the moisture release curve, is likely to be greater for low salt solutions. Hence, we could expect a

Table 11. Relative* experimental and computed saturated conductivities of alluvial soil for different salt solutions. Bulk density of soil, 1.13 g/cm³.

SAR	Cation Concentration (meq/l)	Relative experimental saturated conductivity	Relative computed saturated conductivity
40	640	1.000	1.000
	160	0.719	0.654
	80	0.390	0.472
	40	0.104	0.206
	20	0.002	0.051
	10
	5
	2.5
20	160	1.000	1.000
	40	0.401	0.777
	20	0.213	0.589
	10	0.092	0.287
	5	0.029	0.087
	2.5	0.007	..
10	160	1.000	1.000
	40	0.753	0.790
	10	0.482	0.581
	2.5	0.203	0.318

*Relative saturated conductivity is the conductivity in a solution of lower cation concentration, expressed as a fraction of the conductivity value in the solution of highest cation concentration at the same SAR.

Table 12. Relative* experimental and computed saturated conductivities of red brown soil for different salt solutions. Bulk density of soil, 1.22 g/cm^3 .

SAR	Cation Concentration (meq/l)	Relative experimental saturated conductivity	Relative computed saturated conductivity
40	640	1.000	1.000
	160	0.815	0.904
	80	0.260	0.317
	40	0.011	0.105
	20	0.001	..
	10
	5
	2.5
20	160	1.000	1.000
	40	0.528	0.763
	20	0.169	0.320
	10	0.019	0.033
	5
	2.5
10	160	1.000	1.000
	40	0.882	0.838
	10	0.496	0.483
	2.5	0.110	0.207

*Relative saturated conductivity is the conductivity in a solution of lower cation concentration, expressed as a fraction of the conductivity value in the solution of highest cation concentration at the same SAR.

proportionately greater fractional decrease in experimental saturated conductivity, with reduction in cation concentration at any given SAR, than indicated by the reduction in computed saturated conductivity values calculated from the moisture release curve.

Another possible cause is a greater blocking of conducting pores due to enhanced clay dispersion or particle displacement in the soil columns used in the saturated conductivity determination than in the soil columns used in the moisture release curve determination due to faster flow rate of salt solutions (because of the absence of a porous plate to reduce the flow rate). The use of two separately packed soil columns for the saturated conductivity and moisture release curve determinations of a soil for a specific salt solution could also have contributed to this difference.

(ii) Equivalent salt solutions

The saturated conductivity values in Figures 30 and 31 can be used to determine "equivalent salt solutions for saturated conductivity", which can be defined as salt solutions of different combinations of SAR and cation concentrations for which the soil has the same adjusted saturated conductivity. The "equivalent salt solutions for saturated conductivity" for alluvial and red brown soils packed to bulk densities of 1.13 and 1.22 g/cm³ respectively, are given in Figures 28 and 29.

It was suggested in an earlier section (p. 87), that the "equivalent salt solutions for swelling" could be used to predict the saturated conductivities of soils for the same salt solutions. The accuracy of these predictions would depend on the degree of agreement between the "equivalent salt solutions for swelling and for saturated conductivity". In red brown soil the equivalent salt solutions producing a specific percentage decrease in saturated conductivity

(Figure 29) have higher cation concentrations at all SARs than the equivalent salt solutions for swelling producing the same percentage decrease in pore size index. However, the curves indicating the "equivalent salt solutions for swelling and for saturated conductivity" of this soil coincide or are generally parallel to each other. For instance, in the red brown soil the equivalent salt solutions which reduced the saturated conductivity to 30 per cent of its value for the solution of highest cation concentration, coincided with the equivalent salt solutions for which the pore size index had a value of 0.7. This indicated a similarity in composition of the "equivalent salt solutions for swelling and for saturated conductivity". In the alluvial soils (Figure 28) too, the "equivalent salt solutions for swelling and for saturated conductivity" show similar composition.

The method for predicting the saturated conductivity of soils for different salt solutions using the values of "equivalent salt solutions for swelling" was also evaluated using the following procedure. The decreases in experimentally determined saturated conductivity values in a given soil as the cation concentration was reduced in solutions of a given SAR (e.g. SAR 40) were plotted, as in Figures 30 and 31. The values of "equivalent salt solutions for swelling" of the soil were then used to determine the cation concentration of salt solutions of SARs 20 and 10 at which the saturated conductivity could be expected to have the same value as for solutions of SAR 40 and any specific cation concentration. For example, the alluvial soil, according to the concept of equivalent salt solutions, will have the same saturated conductivity for salt solutions of SAR 20 and cation concentration of 77 meq/l, as for solutions of SAR 40 and cation concentration of 195 meq/l, since these two solutions are equivalent salt solutions. The predicted

saturated conductivities for salt solutions of SAR 20 and 10 and different cation concentrations in alluvial and red brown soil derived in this manner are given in Figures 30 and 31 respectively.

Figure 30 shows that there is similarity between the predicted and calculated values of saturated conductivity in alluvial soil, although the agreement is much closer for the solutions of SAR 20 than for the solutions of SAR 10. In the red brown soil (Figure 31) there is also similarity of the predicted and experimentally determined values of saturated conductivity, especially for the solutions of SAR 20. Figures 30 and 31 also show the range of values of experimentally measured saturated conductivities for specific salt solutions obtained in separate columns of a given soil. The predicted values of saturated conductivity generally fell within these limits or close to this range of experimental values. The relatively greater deviation of the predicted values of conductivity in both soils for solutions of SAR 10 than for solutions of SAR 20, could be at least partly a result of inaccuracies in the determination of cation concentrations at which pore size indices give specific percentage decreases from their values in a high salt solution. This is due to the more gradual changes in these values as the cation concentration is reduced in this relatively lower sodium (SAR 10) solution.

The accuracy of the predictions of saturated conductivity for salt solutions (Figures 30 and 31) depends partly on the accuracy with which the pore size indices used in these methods were determined. Instead of using the pore size indices for determining the equivalent salt solutions, it may be possible to use the values for macroscopic swelling of clay pads, provided that the swelling of these clay pads adequately represents the swelling

of the clays in the soils.

Other methods available for predicting saturated conductivity include a theoretical method (Lagerwerff et al., 1969) based on the diffuse double layer theory, which appears to have some limitations on its use (McNeal, 1974). A semi-empirical method developed by McNeal (1968) using the swelling values of clays gave satisfactory predictions for a limited number of soils with similar clay mineralogy. The complex interaction of the many soil factors which influence the effect of salt solutions on saturated conductivity of soils (Figure 2) could present difficulties in developing a method for predicting saturated conductivities based on the swelling of clays extracted from the soils, which could be generally applied to all soils. The method based on equivalent salt solutions is more direct, as the equivalent salt solutions can be determined on the soil cores for which the saturated conductivity has to be predicted and theoretically this method can be applied to any soil. However, a fair amount of experimental work is required to determine the equivalent salt solutions of a soil and also saturated conductivities as the cation concentration is reduced in solutions of a given SAR.

The saturated conductivities for salt solution of alluvial soil packed to a higher bulk density (1.22 g/cm^3), after adjusting to correspond to an average saturated conductivity in high salt solutions at all SARs as described previously, are given in Figure 32. These values represent an average of three measurements of conductivity in replicate soil columns. The range of variation in saturated conductivities of the three soil columns in the presence of specific salt solutions is indicated in these figures. As the cation concentration was reduced at any given SAR, the saturated conductivity showed decreases which were relatively more marked for higher values of SAR (Table 13).

Table 13. The average experimental saturated conductivities of alluvial soil for salt solutions of SAR 40, 20 and 10 and different cation concentrations. Bulk density of soil, 1.22 g/cm³.

Cation Concentration (meq/l)	SAR 40		SAR 20		SAR 10	
	Saturated conductivity (cm/day)	Relative* saturated conductivity	Saturated conductivity (cm/day)	Relative* saturated conductivity	Saturated conductivity (cm/day)	Relative* saturated conductivity
640	100.1	1.000
160	83.4	0.833	87.6	1.000	90.1	1.000
80	36.1	0.361
40	13.0	0.130	48.9	0.558	71.4	0.792
20	0.40	0.004	19.5	0.223
10	9.4	0.107	42.1	0.467
5	4.2	0.048
2.5	0.18	0.002	23.5	0.261

*Relative saturated conductivity is the conductivity in a solution of lower cation concentration, expressed as a fraction of the conductivity value in the solution of highest cation concentration at the same SAR.

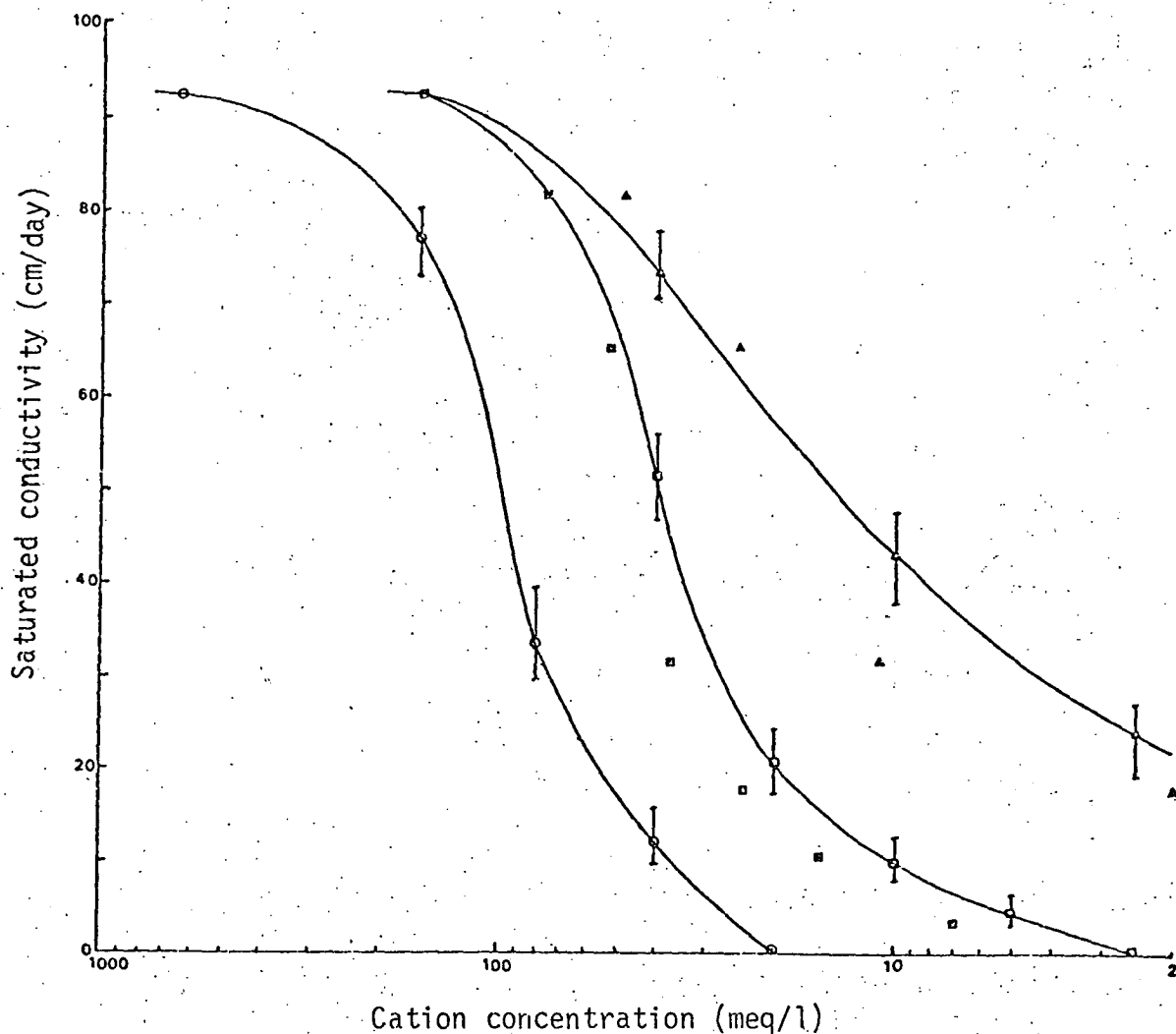


Fig. 32. Adjusted average experimental and predicted saturated conductivity of alluvial soil for different salt solutions. Experimental saturated conductivity for solutions of SAR \circ 40, \square 20, \triangle 10 and different cation concentrations. Predicted saturated conductivity for solutions of SAR \blacksquare 20, \blacktriangle 10 and different cation concentrations. Bulk density of soil, 1.22 g/cm^3 .

The effect of changes in bulk density of this soil on the saturated conductivity for different salt solutions can be assessed by comparing the fractional decrease in saturated conductivity of the soil (Tables 11 and 13), packed to different bulk densities, as the cation concentration was reduced to specific values at a given SAR. The fractional decrease in saturated conductivity in any specific low salt solution at any SAR was generally slightly greater for the soil packed to the lower bulk density. This may be due to a greater fractional decrease in the relatively large sized pores in the alluvial soil packed to lower bulk density (Table 8) as the cation concentration was reduced at a given SAR. The faster flow rate of salt solutions in the soil packed to a lower bulk density during the conductivity measurements, leading to enhanced clay dispersion and blocking of some conducting pores, may also have been a contributing factor.

Figure 32 shows the values of saturated conductivities for salt solutions predicted from the equivalent salt solution method using the procedure previously described. In making these predictions the values of equivalent salt solutions given in Figure 28 were used. Although the predicted values are generally smaller than the experimentally determined values, they appear to be of a similar magnitude. The agreement between the measured and predicted values was closer for the SAR 20 solutions than for the SAR 10 solutions, possibly for the reasons given earlier.

On the basis of the limited data presented in this study, it appears that the equivalent salt solutions method could be used to predict saturated conductivity for salt solutions, although further experimental testing of the method is required.

D. Effect of Salt Solution Composition on Unsaturated Conductivity of Soils

(i) Changes in unsaturated conductivities

The unsaturated conductivities of krasnozem, alluvial and red brown soils, at pressure heads between 0 and -500 cm of water, in the presence of salt solutions of SAR 20 and different cation concentrations are given in Figures 33, 34 and 35 respectively. The nature and magnitude of the changes in values of unsaturated conductivity as the cation concentration was reduced at a given SAR, varied according to the soil as well as the pressure head. Saturated conductivity values for each soil shown in these figures were measured in a soil column packed in a permeameter. The unsaturated conductivity values were measured for each of the solutions used in this experiment on separate, short columns of each soil. Therefore, it may be expected that some of the differences between the unsaturated conductivity values at specific pressure heads in different salt solutions were due to variations in the initial pore geometry of the soils prior to introduction of salt solutions arising from differences in packing of these separate soil columns. However, the studies on moisture release curves in an earlier section showed that such variation could be relatively small when compared to changes caused by salt solutions (p. 71), with proper precautions during packing and experimentation. Hence the observed differences in the unsaturated conductivity-pressure head curves of a soil in salt solutions largely reflect the effects of different salt solution composition.

In salt solutions of SAR 20 and cation concentration 160 meq/l, the krasnozem soil packed to a bulk density of 1.00 g/cm^3 showed high saturated conductivity (Figure 33) and a marked decrease in unsaturated conductivity as the pressure head was gradually reduced. The

unsaturated conductivity values for this soil in solutions of SAR 20 and cation concentrations 2.5 meq/l were slightly smaller at pressure heads greater than -190 cm of water and slightly larger at pressure heads less than -190 cm of water. However, the difference between the values of saturated and unsaturated conductivities for this soil in the two solutions was relatively small, in accord with the very small changes in moisture release curves and the pore geometry as the cation concentration was reduced (Appendix E).

The saturated and unsaturated conductivity at specific high pressure heads for alluvial soil packed to a bulk density of 1.22 g/cm^3 showed very marked decreases with reduction in cation concentration from 160 to 40 meq/l in solutions of SAR 20 (Figure 34). But as the pressure head was reduced to -40 cm of water the difference in unsaturated conductivity for the two solutions became very small. Further, the values of unsaturated conductivity for the salt solution of lower cation concentration (40 meq/l) at specific pressure heads between -40 and -110 cm of water were higher, and at specific pressure heads less than -110 cm of water were lower, than the corresponding values of unsaturated conductivity at specific pressure heads in salt solutions of cation concentration 160 meq/l. However, too much significance cannot be attached to this crossover of the two curves as the differences in values of unsaturated conductivity at specific pressure heads less than -40 cm of water in the presence of the two salt solutions were small. For a salt solution of SAR 20 and a cation concentration 10 meq/l the conductivity values at saturation and at specific high pressure heads was markedly less than the values of unsaturated conductivity for solutions of SAR 20 and cation concentrations 40 meq/l and 160 meq/l. This difference in the values of unsaturated conductivity in different salt solutions became progressively smaller

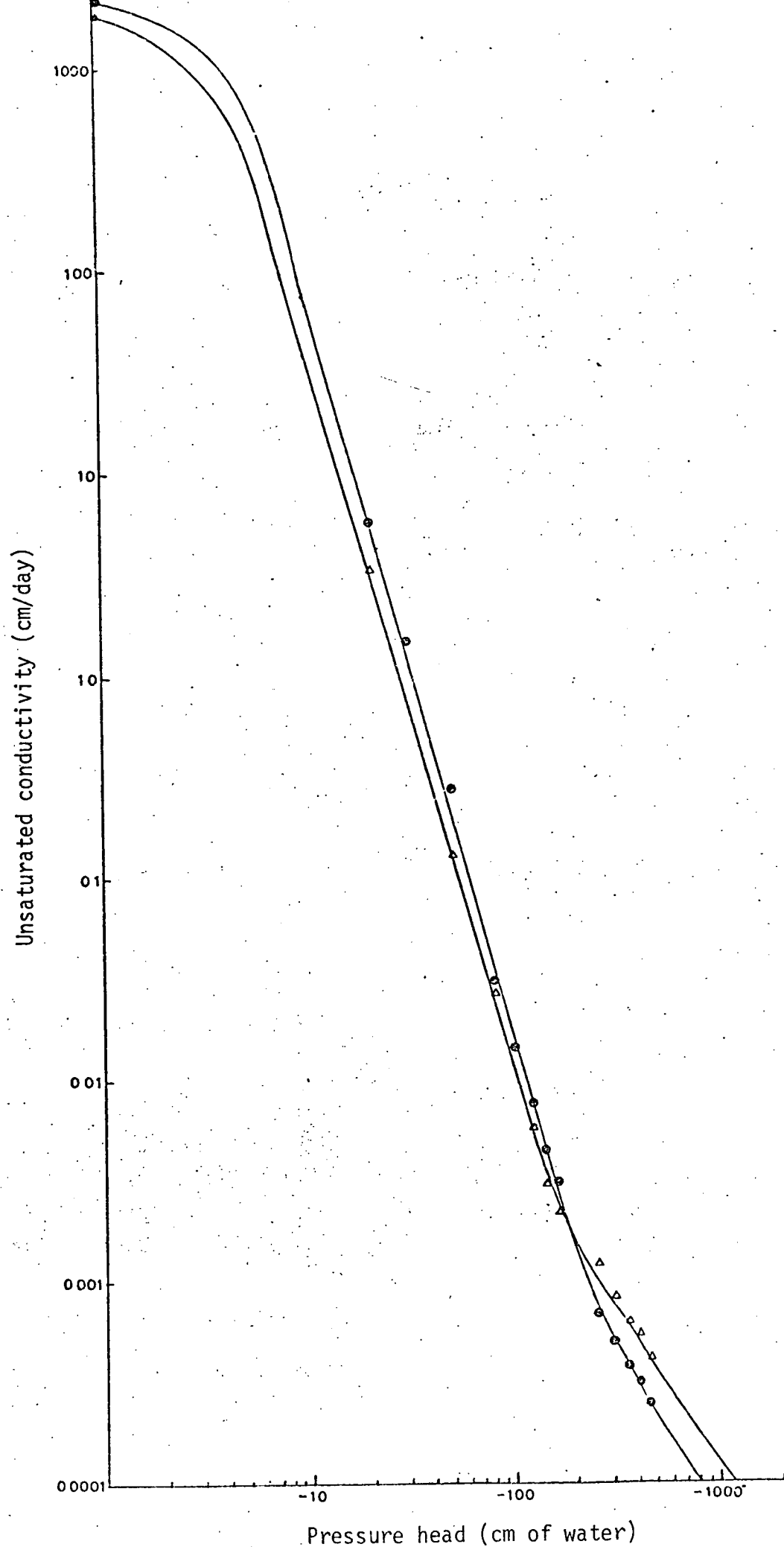


Fig. 33. Unsaturated conductivity-pressure head curves of krasnozem soil for salt solutions of SAR 20 and cation concentrations (meq/l): ● 160, △ 2.5. Bulk density of soil, 1.00 g/cm³.

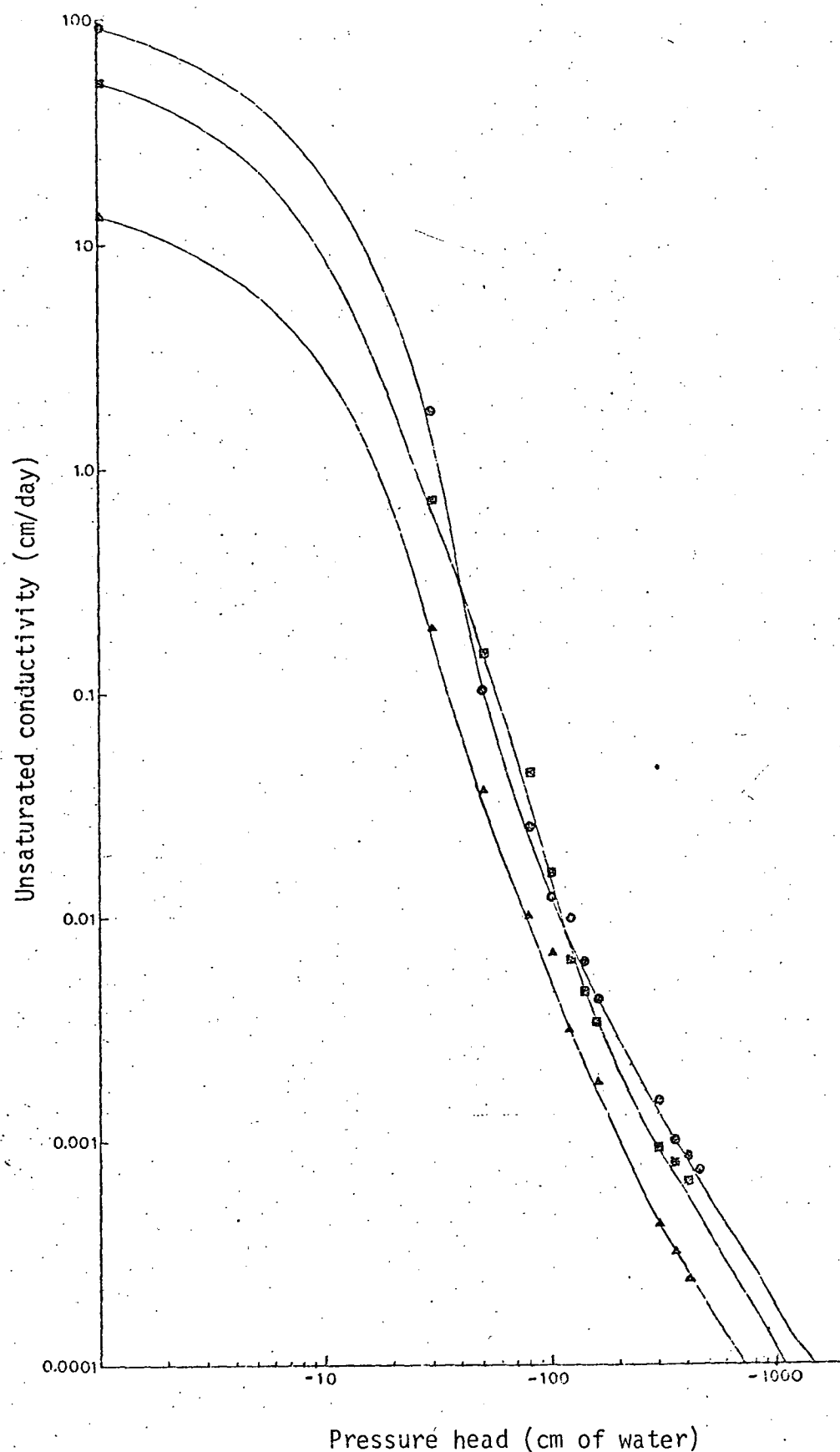


Fig. 34. Unsatuated conductivity - pressure head curves of alluvial soil for salt solutions of SAR 20 and cation concentrations (meq/l): ○ 160, ■ 40, ▲ 10. Bulk density of soil, 1.22 g/cm^3 .

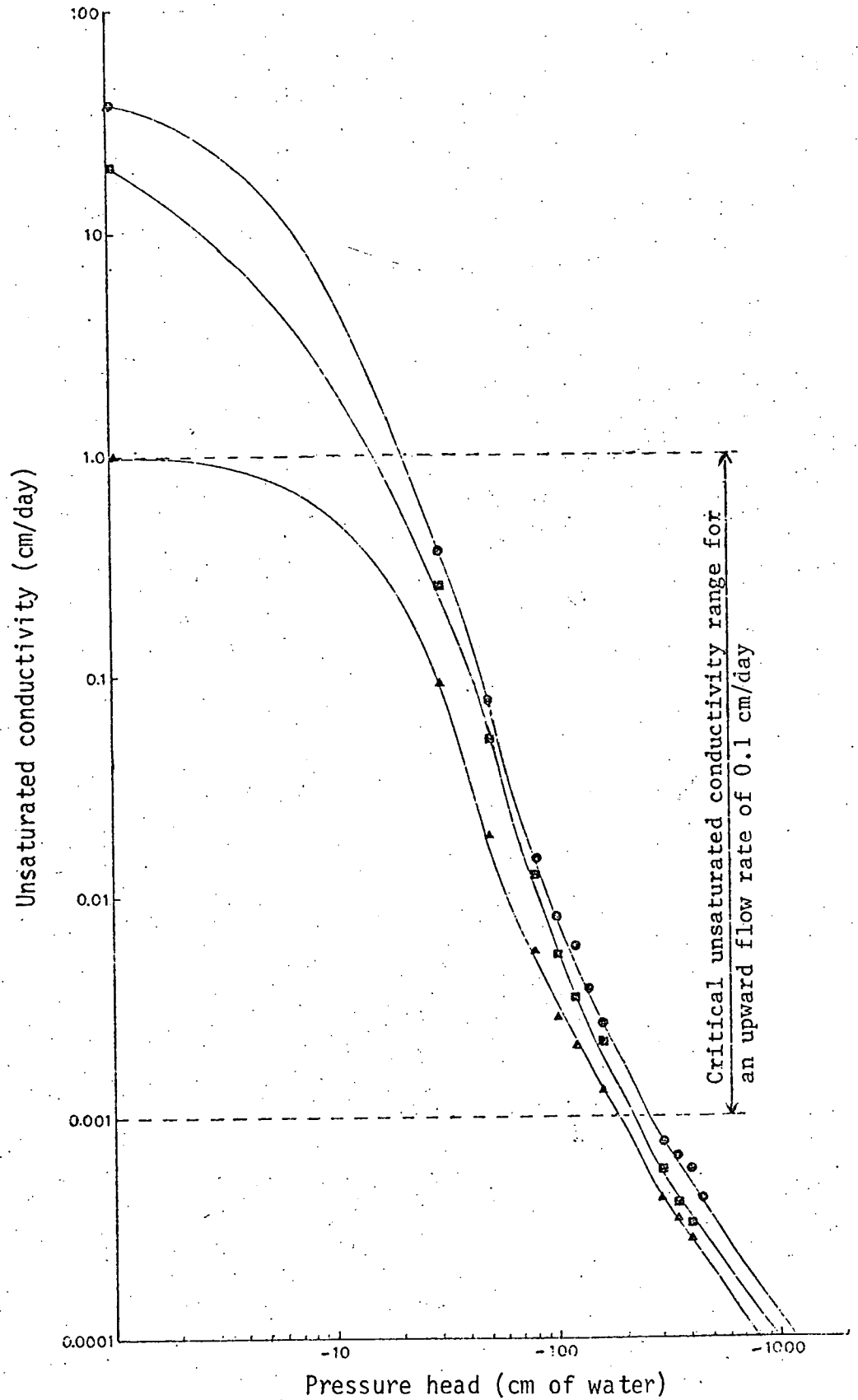


Fig. 35. Unsaturated conductivity-pressure head curves of red brown soil for salt solutions of SAR 20 and cation concentrations (meq/l); ● 160, ■ 40, ▲ 10. Bulk density of soil, 1.22 g/cm^3 .

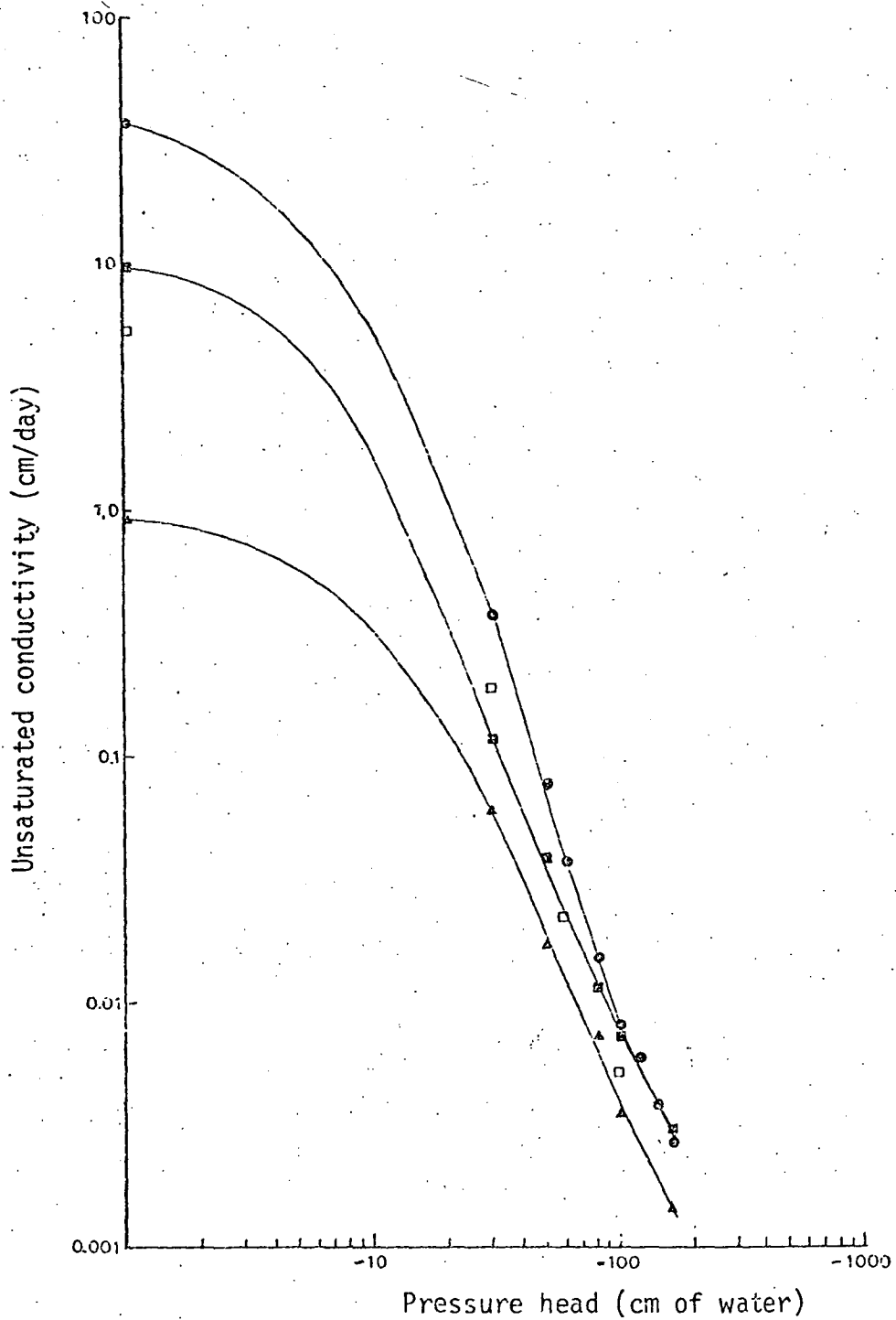


Fig. 36. Unsaturated conductivity-pressure head curves of red brown soil for salt solutions of SAR 40 and cation concentrations (meq/l): \square 80, \triangle 40 and SAR 20 and cation concentrations (meq/l): \circ 160. Predicted unsaturated conductivity values at different pressure heads for solutions of SAR 40 and cation concentration (meq/l): \square 80. Bulk density of soil, 1.22 g/cm^3 .

at lower pressure heads. However, the values of unsaturated conductivity at all specific pressure heads were lower for this solution of cation concentration 10 meq/l than the corresponding values of unsaturated conductivity for the other two solutions of higher cation concentration (160 and 40 meq/l).

The red brown soil packed to a bulk density of 1.22 g/cm^3 showed very marked decreases in conductivity at saturation and at high pressure heads (Figure 35) as the salt concentration was reduced from 160 meq/l to 40 meq/l and then to 10 meq/l in solutions of SAR 20. In this soil too the difference in values of unsaturated conductivity at specific pressure heads became relatively smaller as the pressure head was decreased. However, the values of unsaturated conductivity at all specific pressure heads for solutions of SAR 20 and cation concentrations of 10 meq/l were always lower than the corresponding values of unsaturated conductivity for solutions of cation concentration 40 meq/l and SAR 20, which in turn were lower than the corresponding values of unsaturated conductivity for salt solutions of cation concentration 160 meq/l and SAR 20.

The unsaturated conductivity at pressure heads greater than -160 cm of water in red brown soil packed to a bulk density of 1.22 g/cm^3 for salt solutions of SAR 40 and cation concentrations 80 and 40 meq/l are given in Figure 36. The unsaturated conductivity-pressure head curve of this soil for salt solutions of SAR 20 and cation concentration 160 meq/l is also included. This latter curve could be expected to represent the unsaturated conductivities of this soil for solutions of high cation concentration at all SARs, including SAR 40. These conductivities are likely to be similar on account of the low clay swelling in such high salt solutions. As the cation concentration of this SAR 40 solution was

changed to 80 meq/l and then to 40 meq/l there were marked decreases in conductivities at saturation and at high pressure heads. However, although there were decreases in conductivities at lower pressure heads in solutions of lower cation concentration, these decreases were relatively smaller.

The results obtained in the present study on the unsaturated conductivities of the alluvial and red brown soil for salt solutions show that as the cation concentration decreased in a solution of a given SAR, there were marked decreases in conductivity at saturation and at high pressure heads, but the unsaturated conductivities at low pressure heads generally show much smaller decreases. In a previous study, Sharma (1972) found that three soils with exchangeable sodium percentage of 5 to 20, gave a greater unsaturated conductivity at high moisture contents for water treated with gypsum than for untreated water. But the difference in unsaturated conductivity at low water contents for gypsum treated and untreated water was small. This result is in qualitative agreement with the results of the present study. Gardner *et al.* (1959) showed that the mean-diffusivity of Pachappa soil decreased as the electrolyte concentration was reduced in solutions of specific SARs, and this was largely due to decreases in diffusivity at high water contents. But they also found significant decreases in diffusivity at low water contents for high sodium, low salt solutions and this observation differs from the results of Sharma and those of the present study. These differences could be due to relatively greater swelling or clay dispersion of the Pachappa soil in the presence of salt solutions. It is also possible that the use of a fraction of soil which excluded finer size particles and dusty material in the present study could have resulted in a smaller extent of clay dispersion in low salt, high sodium solution than if the entire fraction below 2mm was used.

(ii) Equivalent salt solutions

The method proposed for predicting changes in unsaturated conductivities for salt solutions using "equivalent salt solutions for swelling" was evaluated using a very similar procedure to that used in predicting saturated conductivities. The values of unsaturated conductivities of red brown soil at a specific pressure head for solutions of SAR 20 and different cation concentrations were plotted against the cation concentration and a curve drawn through these values. The values of the "equivalent salt solutions for swelling" (Figure 29) were used to find and plot the cation concentrations in SAR 40 solutions which would have the same values of unsaturated conductivity, at this specific pressure head as the values of unsaturated conductivity for SAR 20 solutions. A curve was drawn through these values of unsaturated conductivity in solutions of SAR 40 and different cation concentration. The value of unsaturated conductivity at this specific pressure head for a solution of SAR 40 and cation concentration 80 meq/l was read from this curve. A similar procedure was used to determine the unsaturated conductivity at other specific pressure heads for this solution (Figure 36).

Figure 36 shows that the predicted and experimentally determined values of unsaturated conductivity of red brown soil are fairly close, after allowing for experimental error in determining the unsaturated conductivities and the values of equivalent salt solutions. However, it is possible that the closeness of agreement at low pressure heads is due to the small differences in absolute values of conductivity at low pressure heads in different salt solutions, which would necessarily make any deviations of the predicted values from the experimental values very small. Further tests are thus necessary to evaluate the use of this method for salt solutions which cause larger changes in conductivity

at low pressure heads or for other soils where such wide changes occur. This method could not be tested for solutions of SAR 40 and cation concentration of 40 meq/l since its equivalent salt solution with SAR 20 had a cation concentration less than 10 meq/l.

(iii) Computed unsaturated conductivities

The moisture release data for krasnozem, alluvial and red brown soils obtained from the soil columns from which the unsaturated conductivity in the presence of salt solutions were determined, are given in Appendix E. These data were used to compute the unsaturated conductivities of the soils in the presence of the same salt solution (Appendix F) using the equation proposed by Kunze *et al.* (1968). It was shown by these workers that in computing unsaturated conductivities a matching factor had to be used to match a computed value to an experimentally determined value at any single pressure head. It is proposed that the value of this matching factor at different pressure heads could be used as an index of the accuracy of the computation method. A matching factor which remained unchanged at all pressure heads would indicate a close agreement between the computed and experimental values of unsaturated conductivity, while a matching factor which varied widely with pressure head would indicate a poor correlation between these values.

The values of the matching factor at different pressure heads for krasnozem, alluvial and red brown soils for solutions of SAR 20 and different cation concentrations are given in Table 14. The data in Table 14 indicate that for all soils and all salt solutions the value of the matching factor changes as the pressure head is reduced from 0 to -200 cm of water. At pressure heads between -200 and -500 cm of water the matching factor is fairly constant. Hence the values of unsaturated conductivity computed from

the moisture release curves of these soils for specific salt solutions at pressure heads greater than -200 cm of water would not be accurate. However, it appears that it is possible to use this method for computing the values of unsaturated conductivity at lower pressure heads between -200 and -500 cm of water, provided a matching factor in this pressure head range is used.

Previous studies have suggested some of the reasons for the need of a matching factor and the cause of the variation of this matching factor at different pressure heads resulting in a poor agreement between computed and experimental curves. The reasons given include the inherent limitations in the use and accuracy of these methods which have been stressed by the authors of these computation methods (Gardner, 1974), changes in viscosity near soil particle surfaces in heavy textured soils (Low, 1960) and experimental inaccuracies in determining unsaturated conductivity and moisture release curves. While some of these factors could affect the values in Table 14, since their effects are likely to be common to all the values, it could be expected that the differences in these values for different salt solutions in a given soil largely reflect the relative effects of the salt solutions on the value of the matching factor.

Comparison of the matching factors at specific pressure heads in krasnozem soil for the salt solutions with cation concentrations of 160 and 2.5 meq/l and SAR 20 shows that there were only small differences in these sets of values at all pressure heads. On the other hand, in alluvial and red brown soils, as the cation concentration was reduced from 160 to 10 meq/l in this solution of SAR 20, the matching factor at all specific pressure heads showed consistent and fairly marked decreases. These decreases in matching factor in alluvial and red brown soils as the cation concentration was reduced,

Table 14. Values of the matching factor at different pressure heads in krasnozem, alluvial and red brown soils for salt solutions of SAR 20 and different cation concentrations.

Pressure head (cm of water)	KRASNOZEM		ALLUVIAL			RED BROWN		
	Cation concentration (meq/l) 160 2.5		Cation concentration (meq/l) 160 40 10			Cation concentration (meq/l) 160 40 10		
-30	0.051	0.031	0.029	0.011	0.002	0.037	0.023	0.015
-40	0.063	0.039	0.027	0.021	0.002	0.033	0.023	0.007
-50	0.061	0.038	0.022	0.020	0.003	0.026	0.015	0.004
-60	0.048	0.031	0.018	0.021	0.003	0.016	0.010	0.003
-70	0.038	0.025	0.017	0.019	0.003	0.012	0.008	0.002
-80	0.032	0.021	0.015	0.017	0.004	0.009	0.006	0.002
-90	0.026	0.018	0.014	0.014	0.004	0.008	0.005	0.002
-100	0.022	0.015	0.013	0.012	0.004	0.007	0.004	0.002
-150	0.011	0.008	0.009	0.005	0.002	0.004	0.002	0.001
-200	0.006	0.007	0.007	0.004	0.002	0.003	0.002	0.001
-250	0.005	0.007	0.006	0.003	0.001	0.003	0.002	0.001
-300	0.005	0.007	0.005	0.003	0.001	0.002	0.001	0.001
-500	0.004	0.007	0.004	0.002	0.001	0.002	0.001	0.001

can be explained on a qualitative basis using the model (Appendix B) proposed to describe the changes in pore size distribution in soils with a dynamic internal pore geometry, as the pressure head is changed.

As the shrinkage of soil aggregates on drying is likely to be greater in solutions of lower cation concentration at a given SAR, the relative shift of the curve indicating the pore size distribution at any specific pressure head from the moisture release curve, is likely to be greater for the lower salt solution. Since the computed values of unsaturated conductivity are obtained from the moisture release curves while the actual pores available for solution flow are represented by the curve indicating the pore size distribution, the difference between the computed and measured values of unsaturated conductivity at specific pressure heads is likely to be greater for soils in the presence of the solution with the lower cation concentration. Hence according to this theoretical model the value of the matching factor at any specific pressure head could be expected to be lower for these soils in the presence of solutions of lower cation concentration at a given SAR. However, it is possible that other causes related to the presence of low salt solutions, such as increased clay dispersion, could also have contributed to the observed decreases in the matching factor at specific pressure heads, as the cation concentration was reduced at a given SAR.

E. Effect of Salt Solution Composition on Upward Flow of Solutions from a Water Table

(i) Computed pressure head profiles

The computed pressure head profiles above a water table during steady upward flow of solutions of SAR 20 and different cation

concentration in krasnozem, alluvial and red brown soils, packed to bulk densities of 1.00, 1.22 and 1.22 g/cm³ respectively, are given in Figures 37 to 44. The pressure head profile for a particular soil and groundwater salt composition was calculated by solving equation (1) (p.11) using the unsaturated conductivity-pressure head curves of the same soils in the specific salt solution (Figures 33 to 35). The solution of equation (1) for computing upward flow was carried out using a numerical method (Philip, 1957) and a flux/unsaturated conductivity ratio method which was developed during the present study (Appendix A).

The pressure head profiles in krasnozem soil (Figures 37 and 38) were very similar for salt solutions of SAR 20 and cation concentrations of 160 and 2.5 meq/l. In the alluvial soils the differences in the pressure head profiles were small for solutions of SAR 20 and cation concentrations of 160 and 40 meq/l (Figures 39 and 40). However, the pressure head profile in salt solution of SAR 20 and cation concentration 10 meq/l (Figure 41) was markedly different from the pressure head profiles in salt solutions of SAR 20 and higher cation concentrations of 160 and 40 meq/l. In the red brown soil the pressure head profiles in salt solutions of SAR 20 and cation concentrations of 160 and 40 meq/l (Figures 42 and 43) differed little, but as the cation concentration was reduced to 10 meq/l in this SAR 20 solution, marked changes occurred in the shapes of the pressure head profiles (Figure 44).

Thus, although the hydraulic conductivity at saturation and at high pressure heads was markedly reduced in alluvial and red brown soils, as the cation concentration was reduced from 160 to 40 meq/l, in solutions of SAR 20, there were only small changes in the pressure head profiles. But with a further reduction in cation concentration

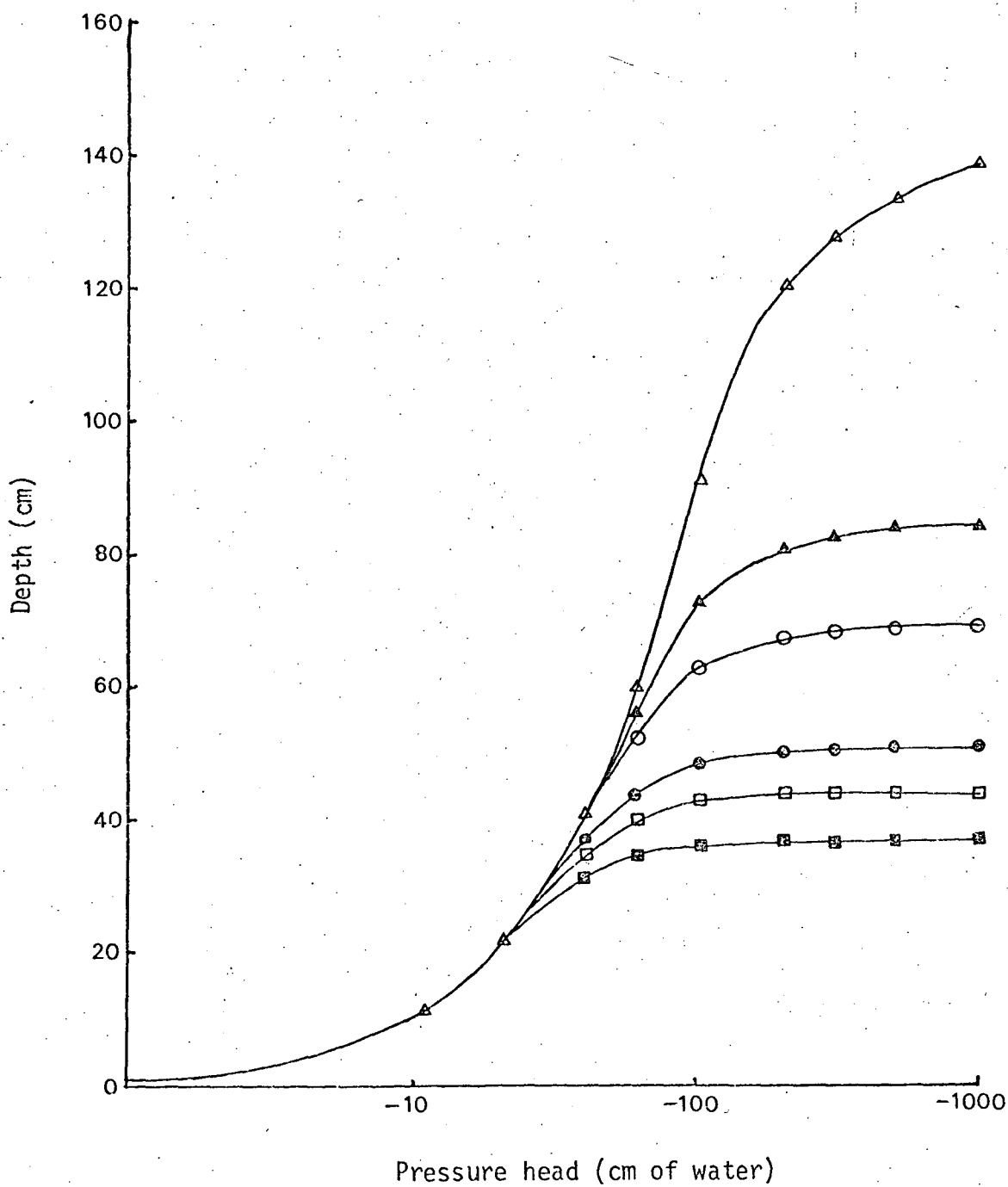


Fig. 37. Computed pressure head-depth curves above a water table in krasnozem soil with groundwater salt composition of SAR 20 and cation concentration 160 meq/l, for upward flow rates (cm/day): Δ 0.01, \triangle 0.05, \circ 0.1, \odot 0.3, \square 0.5, \boxtimes 1.0. Bulk density of soil, 1.00 g/cm³.

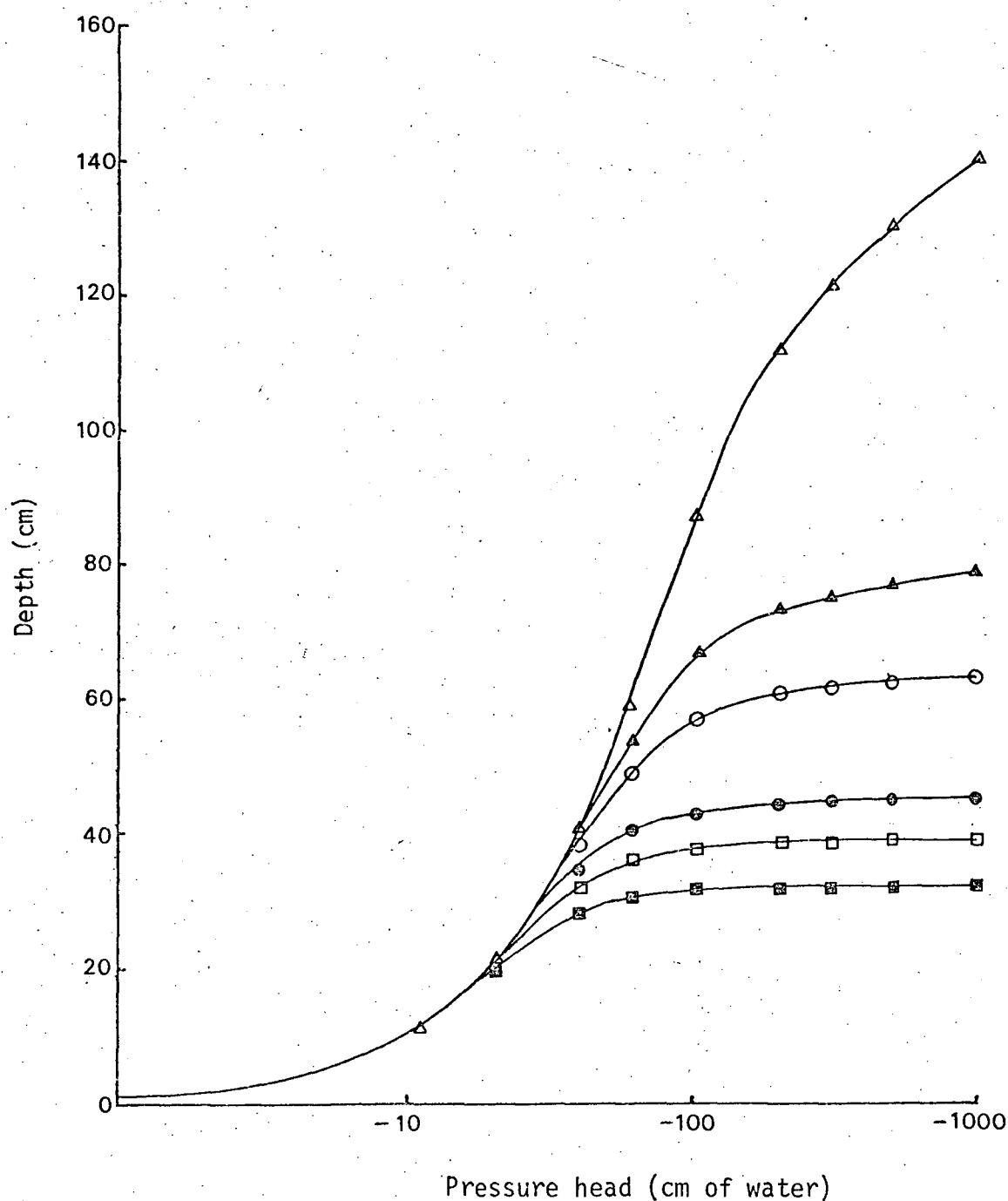


Fig. 38. Computed pressure head-depth curves above a water table in kraznosem soil with groundwater salt composition of SAR 20 and cation concentration 2.5 meq/l, for upward flow rates (cm/day): \triangle 0.01, \blacktriangle 0.05, \circ 0.1, \bullet 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.00 g/cm³.

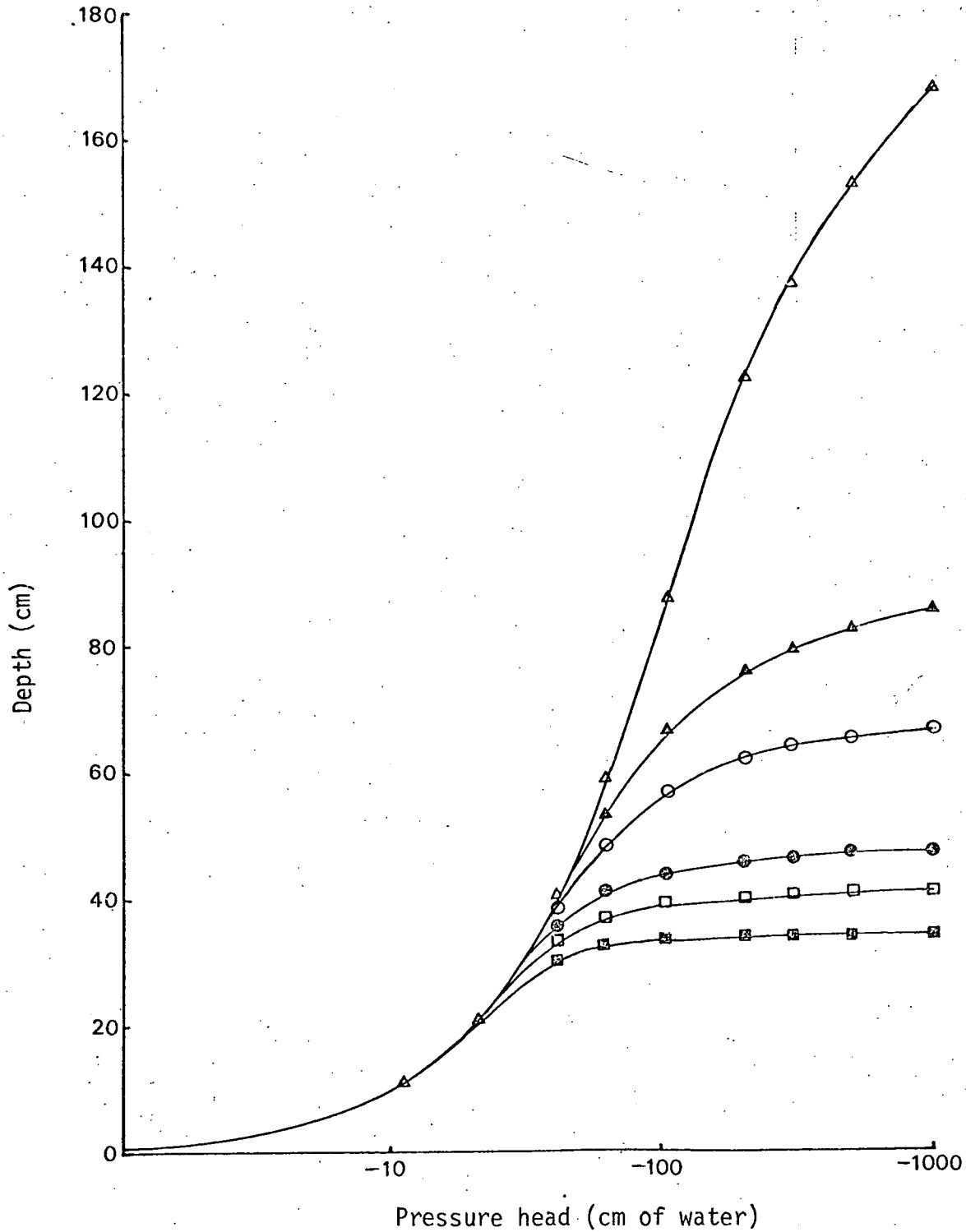


Fig. 39. Computed pressure head-depth curves above a water table in alluvial soil with groundwater salt composition of SAR 20 and cation concentration 160 meq/l, for upward flow rates (cm/day): Δ 0.01, \triangle 0.05, \circ 0.1, \odot 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.22 g/cm³.

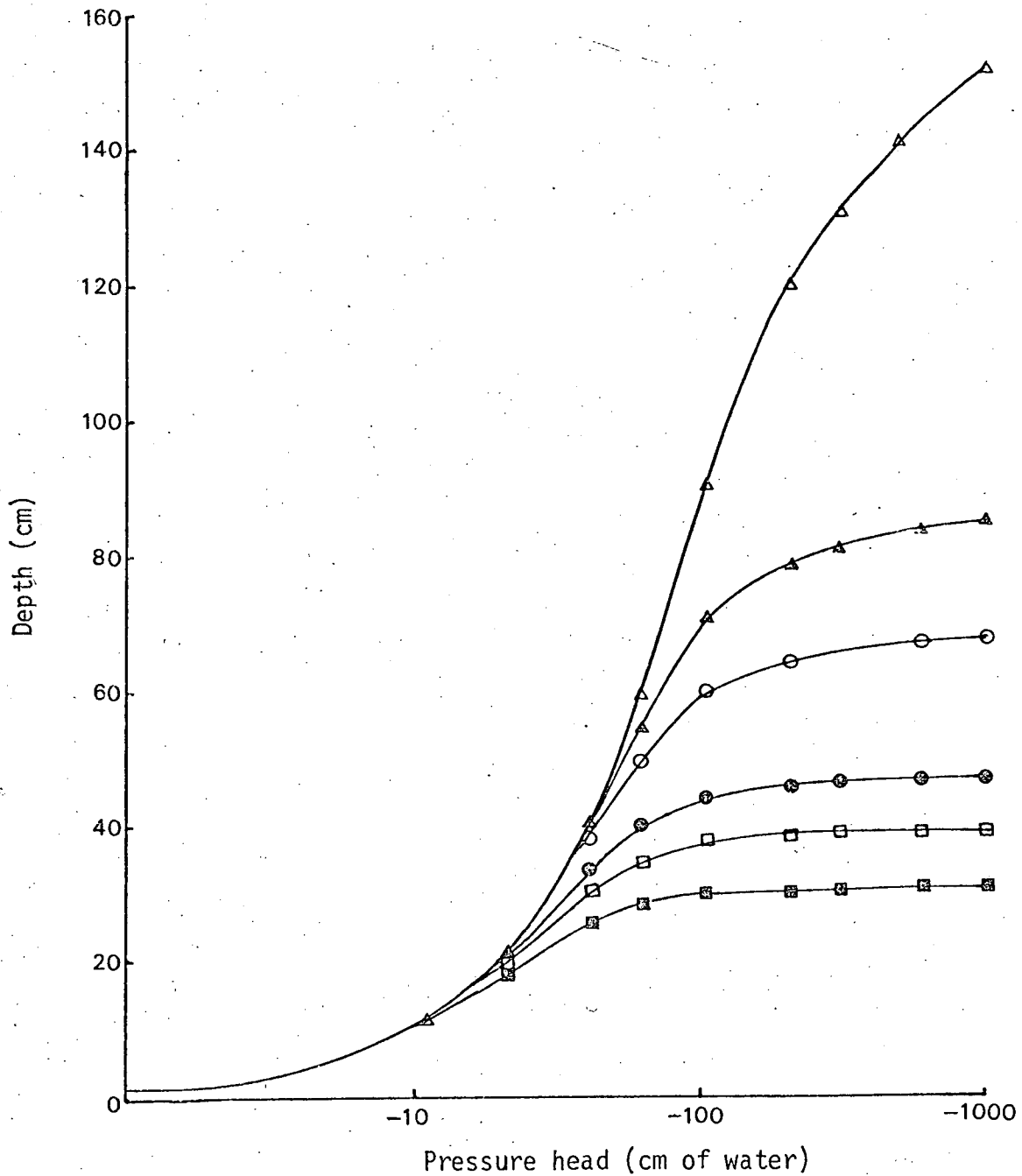


Fig. 40. Computed pressure head-depth curves above a water table in alluvial soil with groundwater salt composition of SAR 20 and cation concentration 40 meq/l, for upward flow rates (cm/day): \triangle 0.01, \blacktriangle 0.05, \circ 0.1, \odot 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.22 g/cm^3 .

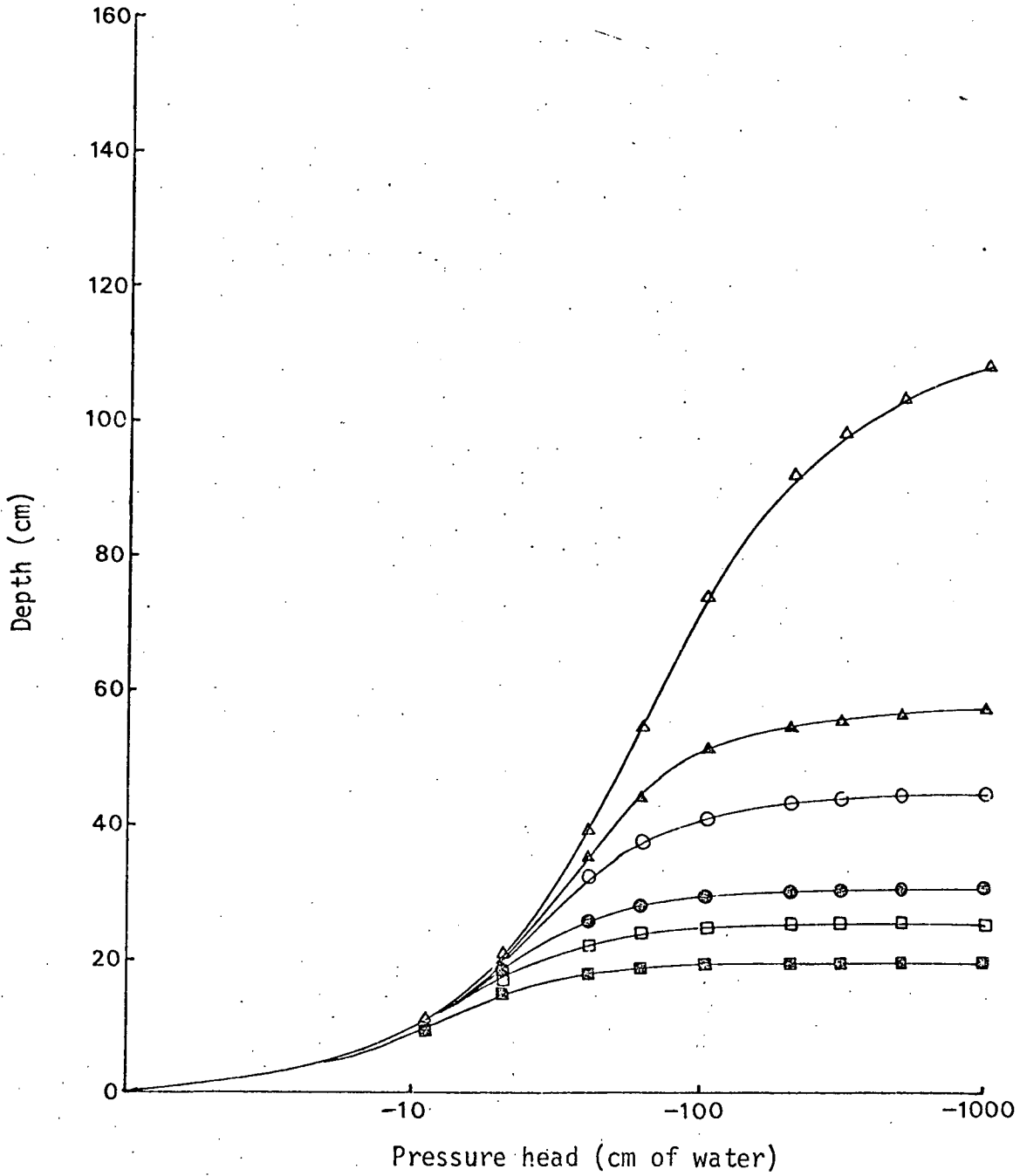


Fig. 41. Computed pressure head-depth curves above a water table in alluvial soil with groundwater salt composition of SAR 20 and cation concentration 10 meq/l, for upward flow rates (cm/day): Δ 0.01, \blacktriangle 0.05, \circ 0.1, \odot 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.22 g/cm^3 .

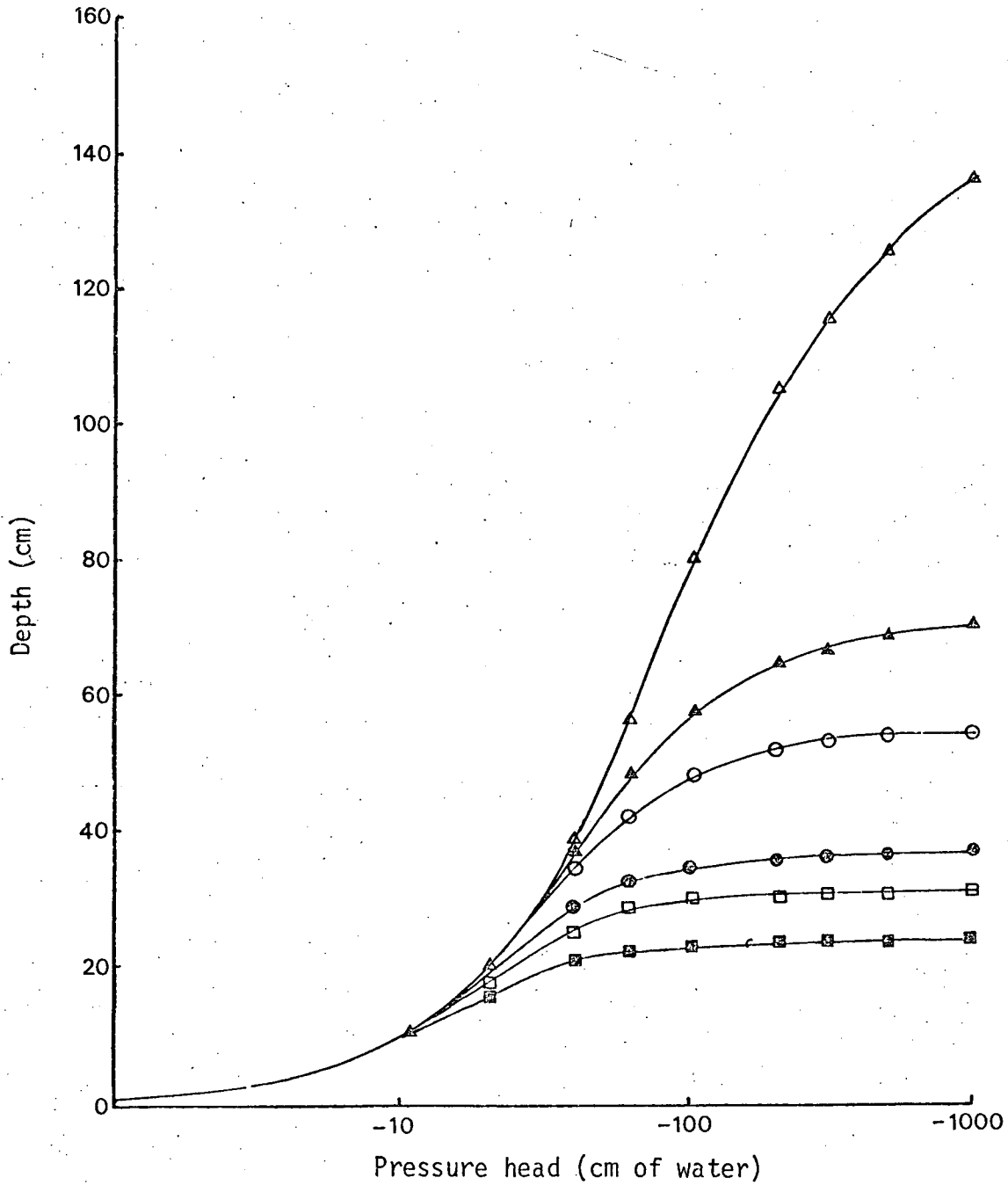


Fig. 42. Computed pressure head-depth curves above a water table in red brown soil with groundwater salt composition of SAR 20 and cation concentration 160 meq/l, for upward flow rates (cm/day): Δ 0.01, \triangle 0.05, \circ 0.1, \odot 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.22 g/cm^3 .

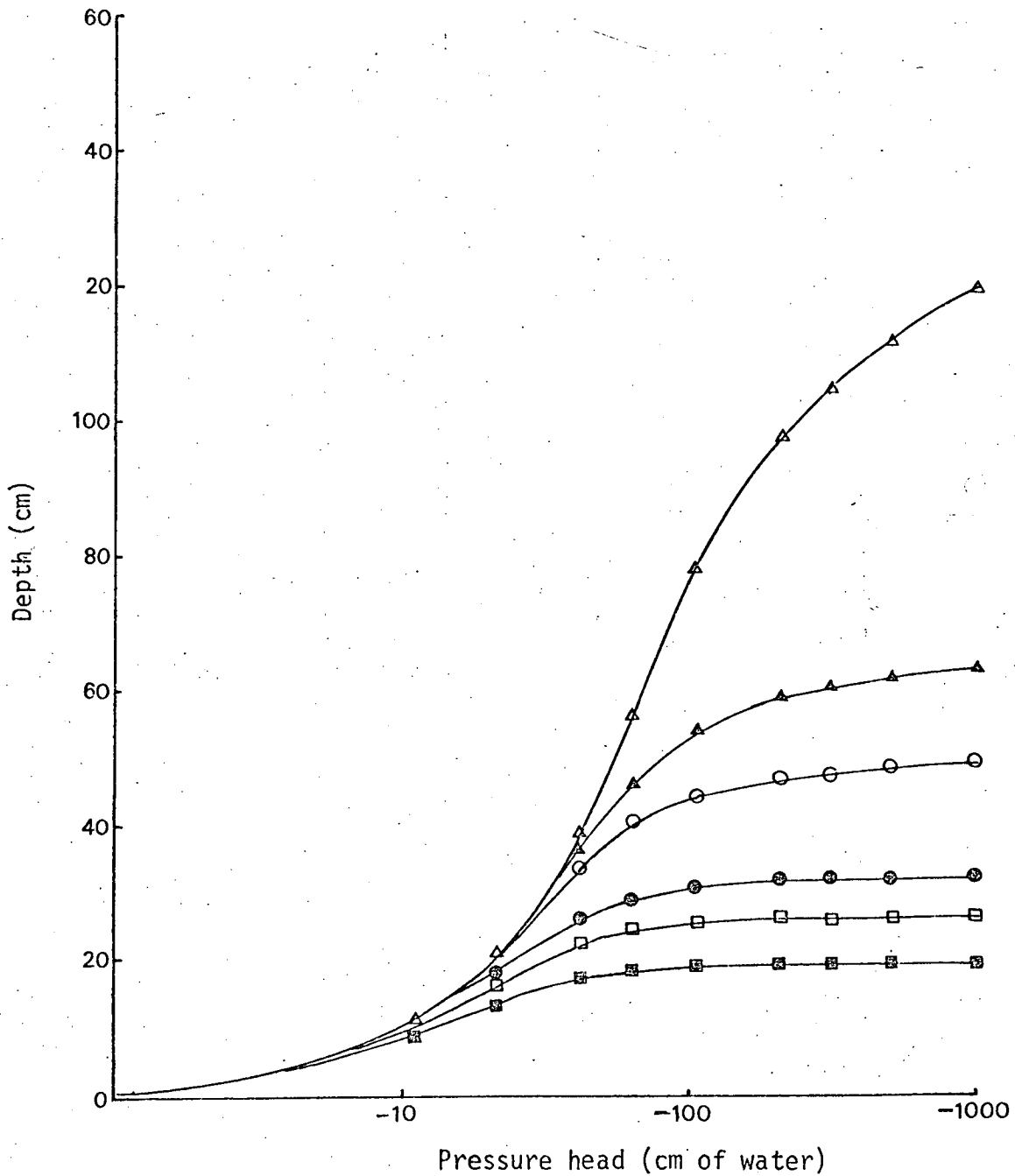


Fig. 43. Computed pressure head-depth curves above a water table in red brown soil with groundwater salt composition of SAR 20 and cation concentration 40 meq/l; for upward flow rates (cm/day): Δ 0.01, \blacktriangle 0.05, \circ 0.1, \otimes 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.22 g/cm^3 .

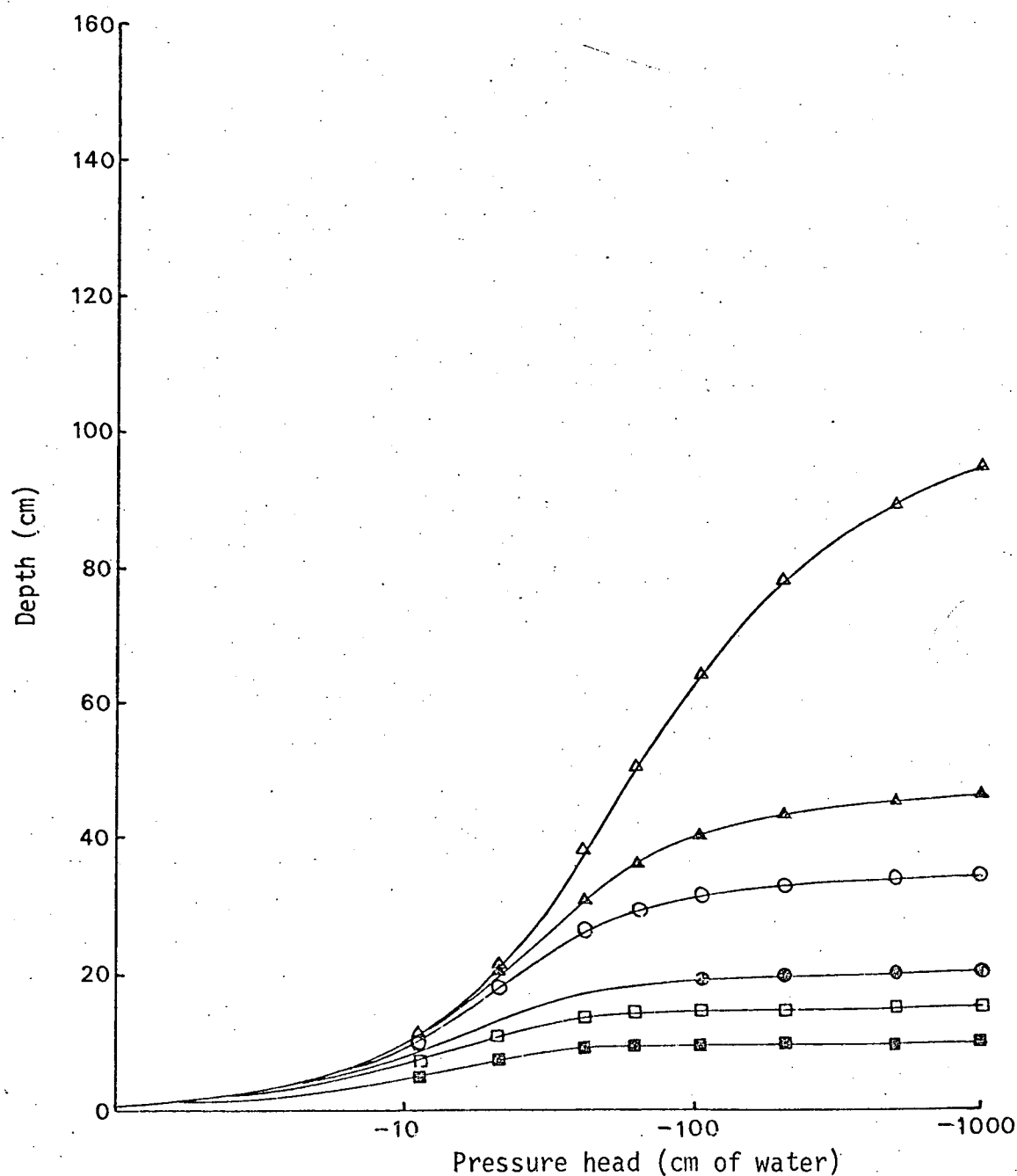


Fig. 44. Computed pressure head-depth curves above a water table in red brown soil with groundwater salt composition of SAR 20 and cation concentration 10 meq/l, for upward flow rates (cm/day): Δ 0.01, \triangle 0.05, \circ 0.1, \bullet 0.3, \square 0.5, \blacksquare 1.0. Bulk density of soil, 1.22 g/cm³.

to 10 meq/l there were marked changes in both the hydraulic conductivities at all pressure heads, and in the pressure head profiles. These differences in the effect of specific salt solutions on conductivity properties and on the pressure head profiles can be explained by demarcating the critical unsaturated conductivity range and the corresponding critical pressure head range (Appendix A) on the pressure head-unsaturated conductivity curves of these soils.

(ii) Critical ranges

The critical pressure head range was defined (Appendix A) as the pressure head range within which the absolute values of unsaturated conductivity significantly affect the shape of the pressure head profile. Thus any marked changes in the absolute values of conductivity within this critical range will lead to marked changes in shape of pressure head profile and hence in rates of upward flow, while any changes in conductivity outside the critical range will not cause significant changes in the pressure head profiles.

The limits of a critical pressure head range appear to have been recognized in earlier studies (Wind, 1955; Philip, 1957; Gardner, 1958). However, it was also shown that the lower limit of this pressure head range could vary in different soils from -300 to -2000 cm of water (Talsma, 1963). In developing the flux/unsaturated conductivity ratio method of computing upward flow and examining the values of the constants used in this method (Appendix A), a more precise definition of critical range applicable to most unsaturated conductivity-pressure head curves was found. This was achieved by demarcating the critical unsaturated conductivity range and then determining the corresponding critical pressure head range for unsaturated conductivity-pressure head curves of different soils.

For unsaturated conductivity-pressure head curves with slopes greater than 2, the critical unsaturated conductivity range and the corresponding critical pressure head range may be defined as k_{u2} to k_{12} and h_{u2} to h_{12} respectively (Appendix A). The rates of upward flow which are important in the field are in the range 0.1 to 2.0 cm/day, since the critical depth corresponds to a rate of flow of 0.1 cm/day (Talsma, 1963) and field evaporation rates seldom exceed 2 cm/day. The critical unsaturated conductivity range corresponding to the range of rates of upward flow of 2.0 to 0.1 cm/day is given by values of unsaturated conductivity between 20 and 0.001 cm/day.

The use of the critical range for comparing upward flow in a soil in the presence of different salt solutions may be illustrated in the following manner. The critical pressure head range for upward flow rates of 0.1 cm/day were marked on the pressure head-unsaturated conductivity curves of red brown soil in solutions of SAR 20 and cation concentrations of 160, 40 and 10 meq/l (Figure 35). The differences in values of unsaturated conductivity in solutions of cation concentration 160 and 40 meq/l within the critical range were small and hence the differences in pressure head-depth curves were small. Hence the depth to water table for this specific rate of upward flow showed only a small decrease from 58 cm to 51 cm. But the decreases in values of unsaturated conductivity within the critical range, as the cation concentration of this SAR 20 solution was reduced from 40 to 10 meq/l, were much larger and this resulted in significant changes in shape of pressure head-depth curves for this specific rate of upward flow. Hence the depth to water table for this rate of upward flow of 0.1 cm/day decreased to 36 cm. Further, since the saturated conductivity in solutions of cation concentration 10 meq/l was close to the upper limit of the critical range, any

further reductions in cation concentration of this SAR 20 solution could lead to decreases in conductivities within the critical pressure head range. Hence such reductions in cation concentration of this solution could be expected to lead to further marked decreases in the depth to the water table for this specific rate of upward flow. If the cation concentration was reduced to such an extent that the saturated hydraulic conductivity fell below the lower limit of the critical range for a specific rate of upward flow, the maximum depth to water table for that rate of flow would be extremely small.

Thus, in the alluvial and red brown soils which are unstable in the presence of salt solution, the decrease in cation concentration below the threshold concentration in solutions of a given SAR led to marked decreases in hydraulic conductivity at saturation and at high pressure heads. But this would not necessarily lead to a decrease in rate of upward flow, unless there were simultaneous decreases in values of unsaturated conductivity within the critical range. The magnitude of the changes in rate of upward flow due to salt solutions will depend on the nature and magnitude of the changes in values of conductivities within the critical range.

The definition of the critical range has a number of other applications to field studies relating to upward flow of water and salts from a water table. Accurate computation of upward flow requires that the unsaturated conductivity-pressure head curves should be determined accurately within the critical range on samples representative of the field soil or *in situ*. However, where the saturated conductivity is very much greater than the upper limit of the critical range of unsaturated conductivity, the data from a field core should be adequate for soils which do not show bulk swelling because, while the disturbance of the core in sampling is

likely to affect the conductivity at high pressure heads above the critical range, it is unlikely to affect significantly the unsaturated conductivity in the critical range. On the other hand, if the saturated conductivity falls within or close to the critical range, it is necessary to take all precautions to prevent any disturbance in taking the field soil core, or alternately, it may be necessary to make the determinations of unsaturated conductivity at high pressure heads in the field, if accurate computations of upward flow are required. Thus the definition of the critical range could help in determining the type of sample to use for measuring unsaturated conductivities. Further, in view of the experimental difficulties involved in determining the unsaturated conductivity-pressure head curves of soils, often requiring a combination of several methods, the definition of the critical range will help in selecting the best method or methods for accurate measurement of the unsaturated conductivities within the critical range, while the simpler, but possibly less accurate methods, such as the methods for computing the conductivities from the moisture release curves, will suffice for determining the values of conductivities outside the critical range. This could also result in a saving in time, cost and effort in making computations of upward flow, especially in field studies.

(iii) Comparison of computed and measured rates of upward flow.

The maximum depth to water table for specific rates of steady upward flow of salt solutions of SAR 20 and different cation concentrations in krasnozem, alluvial and red brown soils were obtained from the computed pressure head profiles for the same salt solutions and soils (Figures 45, 46 and 48). The rates of upward flow

from groundwaters of the same salt solution compositions were also measured in long columns of the soils. These long columns were packed to the same bulk density as the short columns of these soils used for measuring the unsaturated conductivities from which the upward flow rates were computed. These measured rates of upward flow were used firstly for comparison with the computed rates of upward flow and secondly for evaluation of the relative changes of computed and measured rates of upward flow in a given soil from specific depths to water table, as the cation concentration was reduced at a given SAR.

In the krasnozem soil there was close agreement between the computed and experimentally determined rates of upward flow from specific depths to water tables with salt composition of SAR 20 and cation concentration of 160 meq/l and 2.5 meq/l (Figure 45). Both the computed and the experimentally measured rates of upward flow indicated that the reduction in cation concentration from 160 to 2.5 meq/l in solutions of SAR 20 had only a very small effect.

In the alluvial and red brown soils the experimentally determined rates of upward flow from specific depths to water table with salt composition of SAR 20 and cation concentration 160 meq/l, were generally less than the corresponding computed rates of upward flow (Figures 47 and 49). These differences between the measured and the experimentally determined rates of upward flow from a water table may be attributed largely to the use of separate columns, a short soil column for determining the unsaturated conductivity values utilized in computing upward flow and a long soil column for measuring the rates of upward flow. It is also possible that small variations in bulk density and internal pore geometry at different depths within the long soil column could have contributed to these differences. However,

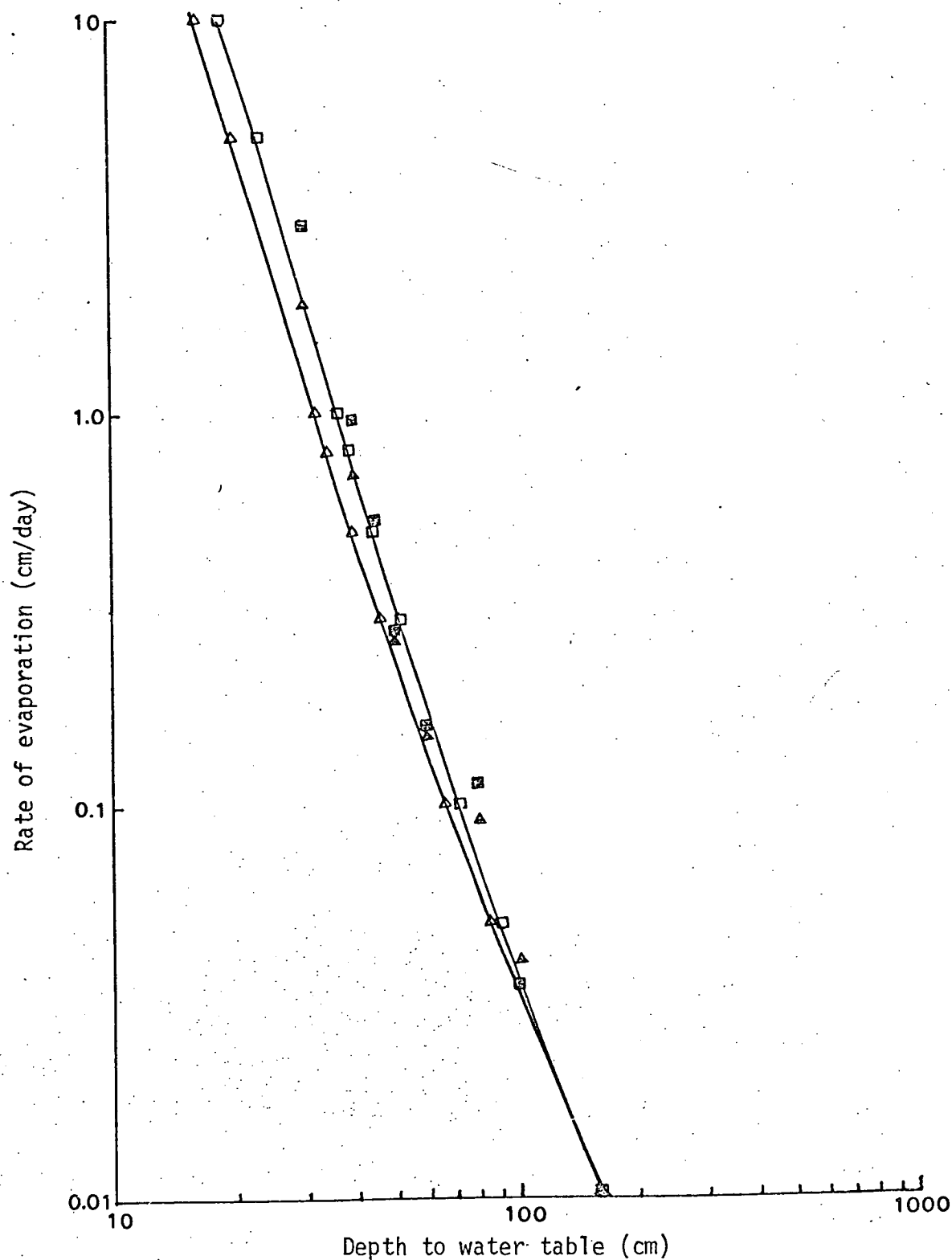


Fig. 45. The computed and experimental rates of upward flow in krasnozem soil from different depths to water tables with groundwater salt composition of SAR 20 and cation concentrations of 160 and 2.5 meq/l. Computed rates of flow for salt solutions of SAR 20 and cation concentration (meq/l): □ 160, △ 2.5. Experimental rates of flow for salt solutions of SAR 20 and cation concentration (meq/l): ■ 160, ▲ 2.5. Bulk density of soil, 1.00 g/cm³.

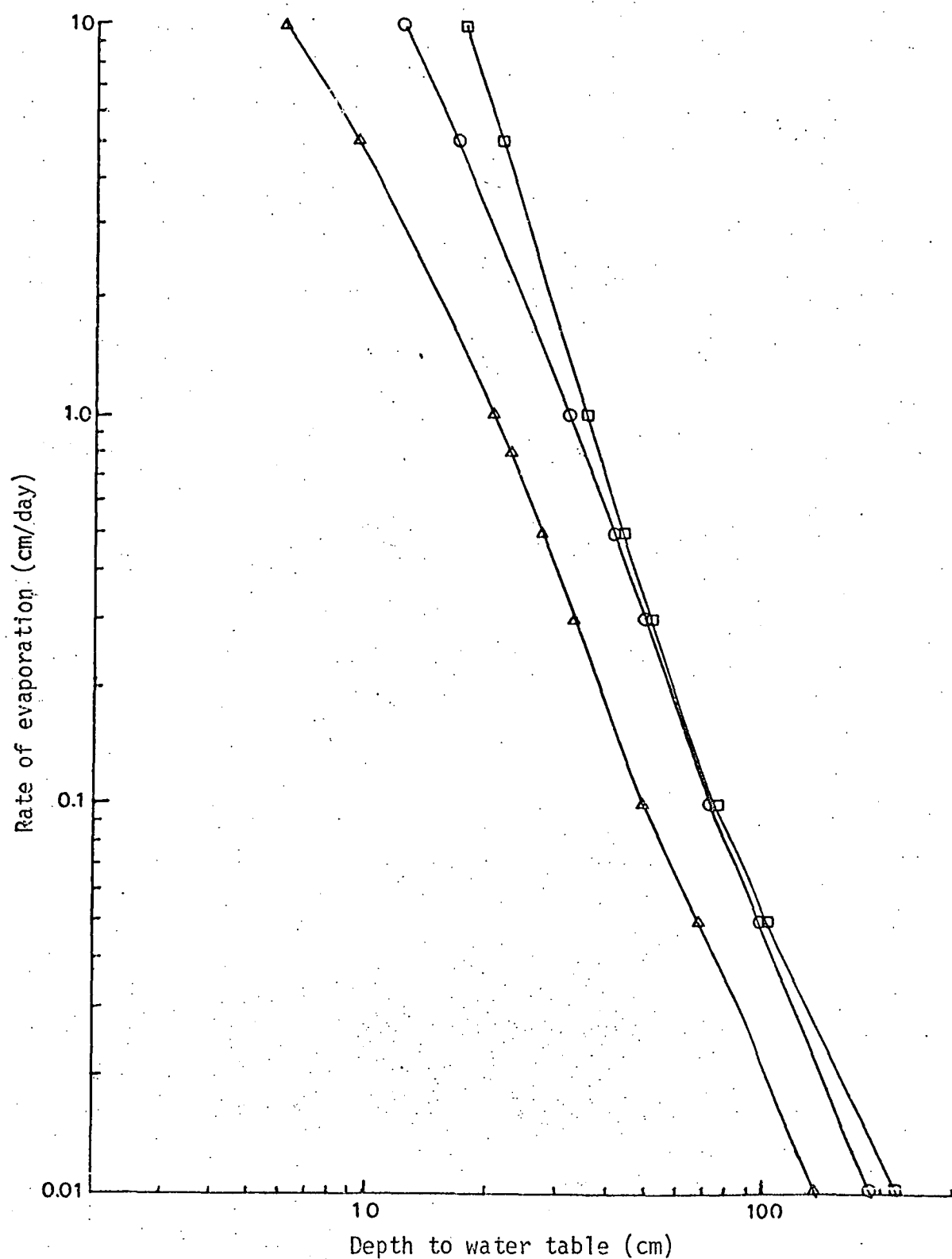


Fig. 46. The computed rates of upward flow in alluvial soil from different depths to water tables with groundwater salt composition of SAR 20 and cation concentrations (meq/l): \square 160, \circ 40, \triangle 10. Bulk density of soil, 1.22 g/cm^3 .

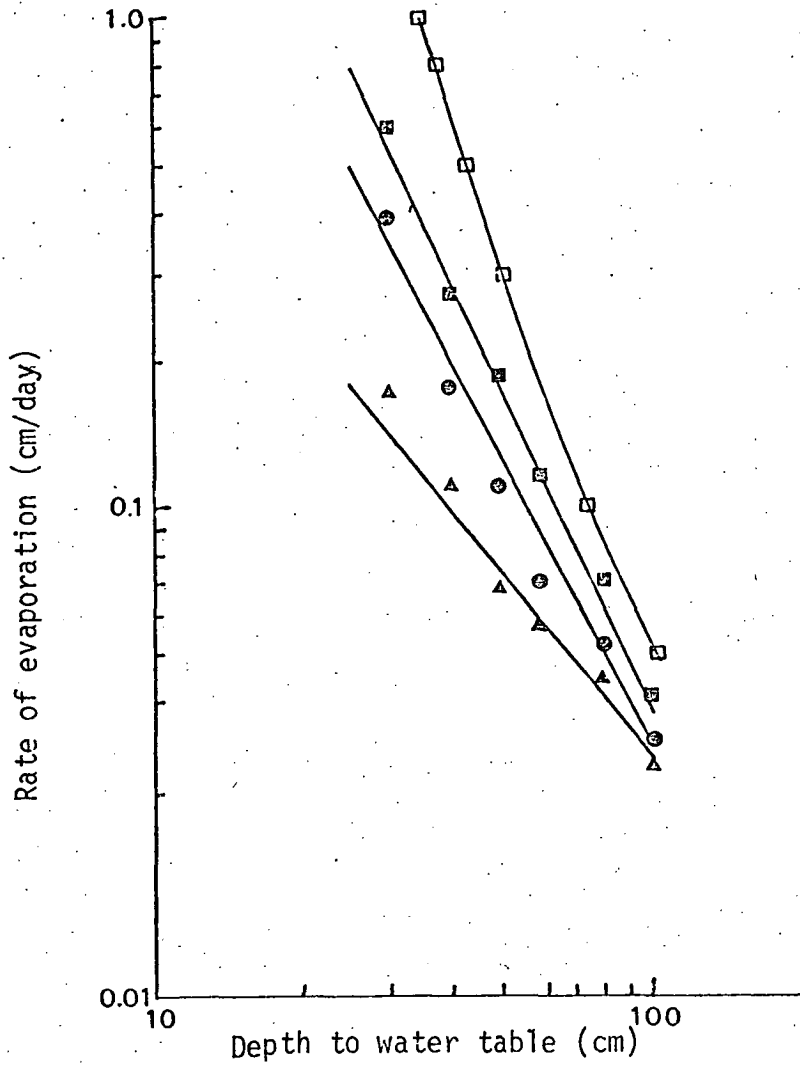


Fig. 47. The experimental rates of upward flow in alluvial soil from different depths to water tables with groundwater salt composition of SAR 20 and cation concentrations (meq/l): \blacksquare 160, \odot 40, \blacktriangle 10. Computed rates of upward flow in alluvial soil with groundwater salt composition of SAR 20 and cation concentration (meq/l): \square 160. Bulk density of soil, 1.22 g/cm^3 .

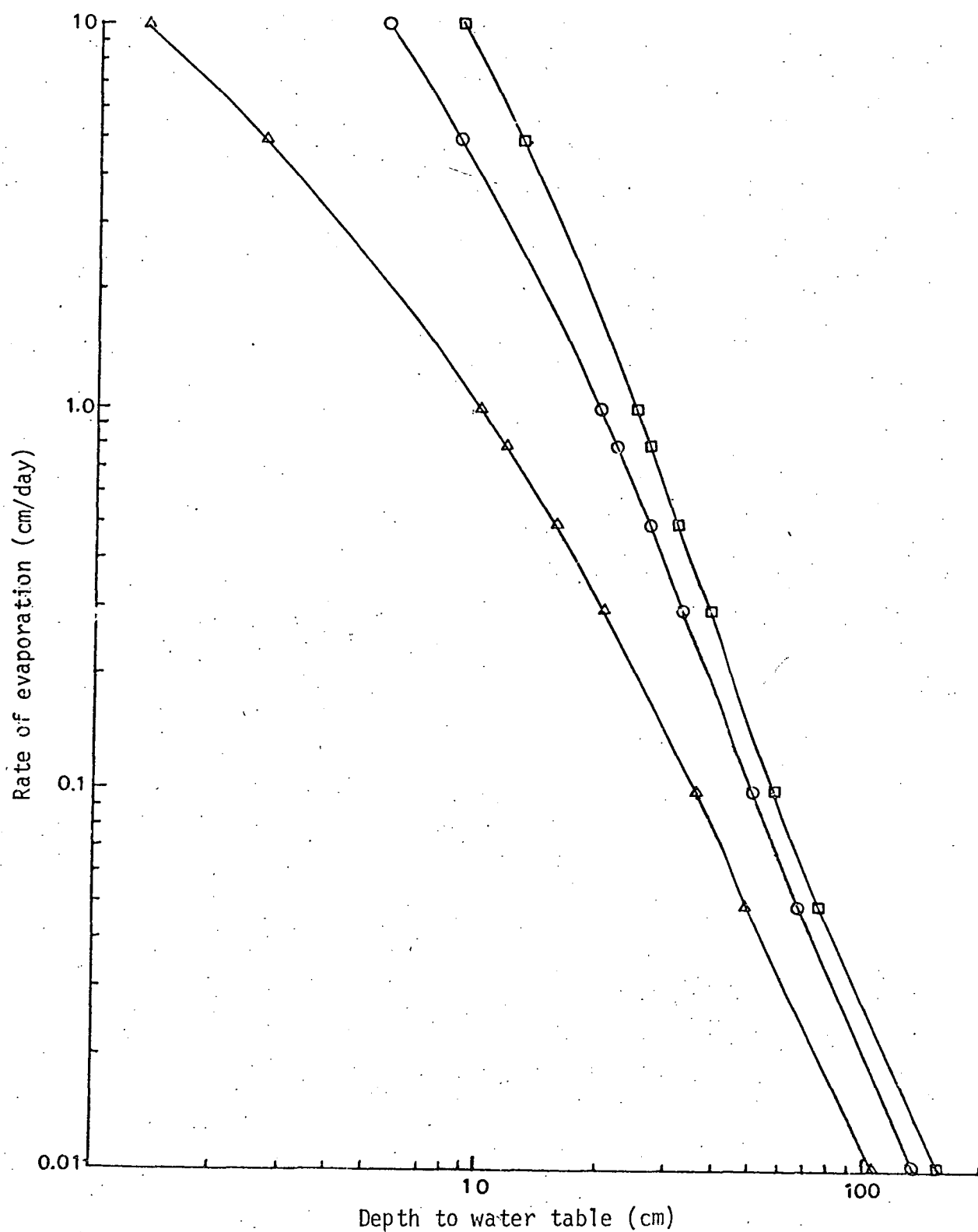


Fig. 48. The computed rates of upward flow in red brown soil from different depths to water tables with groundwater salt composition of SAR 20 and cation concentrations (meq/l): ◻ 160, ○ 40, Δ 10. Bulk density of soil, 1.22 g/cm³.

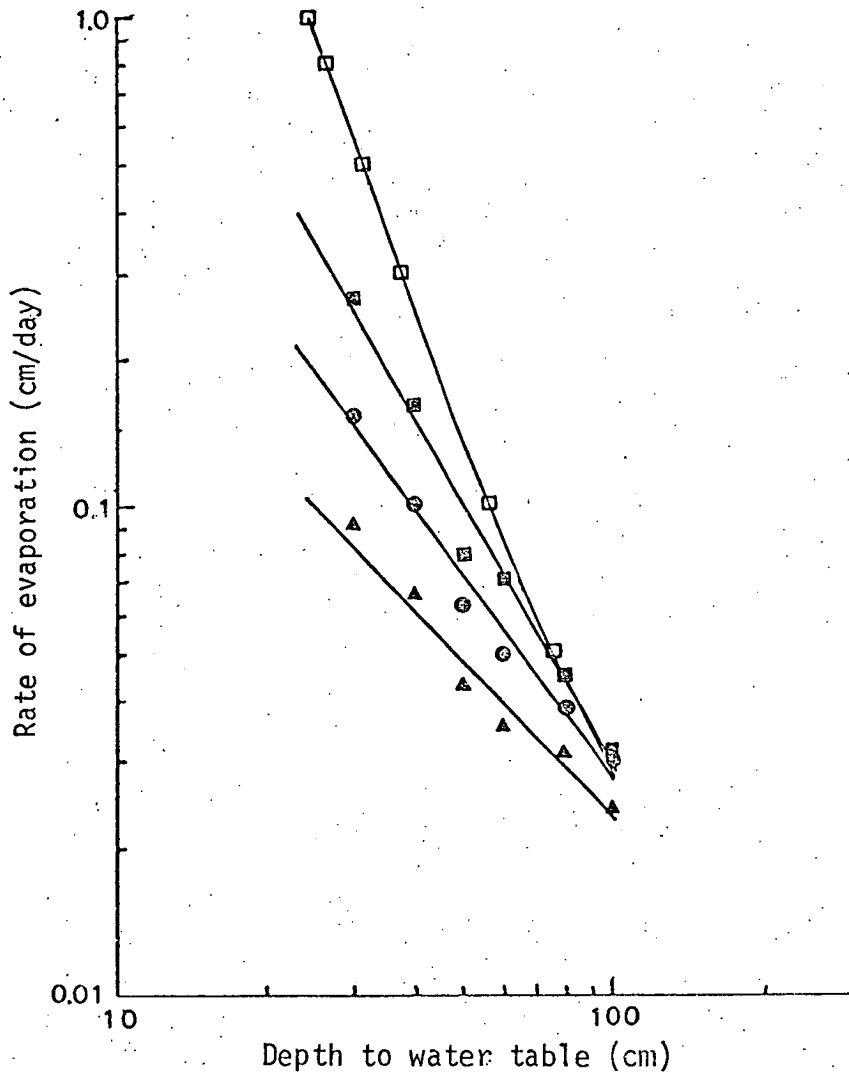


Fig. 49. The experimental rates of upward flow in red brown soil from different depths to water tables with groundwater salt composition of SAR 20 and cation concentrations (meq/l): \square 160, \circ 40, \triangle 10. Computed rates of upward flow in red brown soil with groundwater salt composition of SAR 20 and cation concentration (meq/l): \square 160. Bulk density of soil, 1.22 g/cm^3 .

allowing for such variations due to packing, the measured rates of upward flow appear to be of similar magnitudes to the rates of upward flow computed from the isothermal, steady flow model, especially for greater depths to water table.

In the alluvial soil the computed and measured rates of upward flow from specific depths to the water table for salt solution of SAR 20 and cation concentration 40 meq/l were lower than the rates of upward flow from the same depths to water table for salt solution of SAR 20 and cation concentration 160 meq/l (Figures 46 and 47). With a reduction in cation concentration to 10 meq/l in this SAR 20 solution there was a further decrease in rate of upward flow from all specific depths to water table. In the red brown soils too, the computed and measured rates of upward flow from specific depths to water table decreased for groundwater solutions of SAR 20 in order of decreasing cation concentration from 160 to 40 to 10 meq/l. Thus, in alluvial and red brown soils there is qualitative agreement between the changes in computed and measured rates of upward flow from specific depths to water table with changing salt composition of the groundwater.

A comparison of the changes in the computed and experimental rates of upward flow of different salt solutions could be made on a quantitative basis in the following manner. The computed rates of upward flow in alluvial and red brown soil with reduction in cation concentration to specific concentrations in solutions of SAR 20, expressed as a fraction of the computed rate of upward flow for the solution of highest salt concentration from the same water table depth, are given in Tables 15 and 16, respectively. The decreases in experimentally determined rates of upward flow with reduction in cation concentration in SAR 20 solutions, expressed in a similar manner, are also given in Tables 15 and 16. Comparison of these values shows that as the cation

Table 15. Relative* computed and experimental rates of upward flow from specific depths to watertable in alluvial soil with reduction in cation concentration of ground water of SAR 20. Bulk density of the soil, 1.22 g/cm^3 .

Depth to watertable (cm)	RELATIVE COMPUTED RATES*			RELATIVE EXPERIMENTAL RATES*		
	Cation concentration (meq/l)			Cation concentration (meq/l)		
	160	40	10	160	40	10
20	1.00	0.49	0.16	1.00
30	1.00	0.69	0.24	1.00	0.65	0.27
40	1.00	0.80	0.26	1.00	0.71	0.35
50	1.00	0.88	0.30	1.00	0.72	0.41
60	1.00	0.94	0.36	1.00	0.73	0.49
70	1.00	0.92	0.39	1.00	0.77	0.58
80	1.00	0.92	0.41	1.00	0.79	0.65
90	1.00	0.89	0.41	1.00	0.83	0.73
100	1.00	0.89	0.40	1.00	0.89	0.82
120	1.00	0.86	0.39
140	1.00	0.81	0.37

*The relative rate is the rate of upward flow from a groundwater of lower cation concentration expressed as a fraction of the upward flow rate from groundwater of the highest cation concentration at the same SAR. The values for computing these relative rates were read from the curves in Figs. 46 and 47.

Table 16. Relative* computed and experimental rates of upward flow from specific depths to watertable in red brown soil with reduction in cation concentration of groundwater of SAR 20. Bulk density of soil, 1.22 g/cm³.

Depth to watertable (cm)	COMPUTED RATES*			RELATIVE EXPERIMENTAL RATES*		
	Cation concentration (meq/l)			Cation concentration (meq/l)		
	160	40	10	160	40	10
10	1.00	0.49	0.13
20	1.00	0.62	0.19
30	1.00	0.65	0.25	1.00	0.60	0.33
40	1.00	0.69	0.31	1.00	0.67	0.40
50	1.00	0.67	0.34	1.00	0.72	0.48
60	1.00	0.73	0.38	1.00	0.77	0.53
70	1.00	0.76	0.40	1.00	0.82	0.60
80	1.00	0.74	0.42	1.00	0.84	0.66
90	1.00	0.74	0.43	1.00	0.89	0.72
100	1.00	0.74	0.44	1.00	0.90	0.77

*The relative rate is the rate of upward flow from a groundwater of lower cation concentration expressed as a fraction of the upward flow rate from groundwater of the highest cation concentration at the same SAR. The values for computing these relative rates were read from the curves in Figs. 48 and 49.

concentration was reduced in this SAR 20 solution, the percentage decrease in measured rate of flow was similar to the percentage decrease in computed rate of upward flow, especially for shallow depths to water table. The relatively larger differences in the two sets of values for greater depths to water table could be due to inaccuracies in experimental measurement of low rates of upward flow from deep water tables. For each soil the percentage decreases in measured and computed rates of upward flow with reduction in cation concentration of this SAR 20 solution, were also greater for smaller depths to water table than for greater depths to water table.

The present study does not provide any experimental data for changes in the rate of upward flow of salt solutions from a water table with reducing cation concentration in solutions of SAR values other than SAR 20. However, these changes in rates of upward flow for solutions of other SARs could be expected to be qualitatively similar to those observed with reducing cation concentration in SAR 20 solution, according to the concept of equivalent salt solutions. The limited data available indicate that this concept may be applied to red brown soil.

(iv) Applications to field studies and areas for future studies

While the present study was carried out on fragmented soils packed into soil columns, the results obtained have several practical applications to field studies. One of the main problems in computing upward flow from water tables in the field would be to determine the unsaturated conductivity-pressure head curves of the soils for salt solutions, especially in locations where the chemical composition of groundwaters changes over relatively short distances. The

definition of critical range would narrow the range of pressure heads within which the unsaturated conductivity must be determined accurately on samples representative of field soils. As pointed out in an earlier discussion (p.135), the definition of the critical range could aid the field research worker in choosing the samples and methods for determining unsaturated conductivities which would result in the most accurate computations of upward flow with possibly the minimum time, effort and cost.

Since moisture release curves can be easily determined, unlike unsaturated conductivities which are more difficult and time consuming, the use of moisture release curves to compute unsaturated conductivities would appear to be a practical alternative. In this respect, the presence of wide variations in chemical composition of groundwaters need not be a problem, since as shown in the present study, moisture release curves for a wide range of salt solutions may be calculated from moisture release data for a few selected salt solutions. However, in the alluvial and red brown soils used in the present study the existing methods for computing unsaturated conductivities from moisture release curves gave accurate predictions of conductivity values only at pressure heads between -200 and -500 cm of water. Therefore a combination of methods involving experimental determination of unsaturated conductivity at pressure heads greater than -200 cm of water and computation of unsaturated conductivities at lower pressure heads could be used for such soils.

Another approach to this problem in field studies would be to use the concept of equivalent salt solutions. The very limited experimental data available indicate the possibility that the equivalent salt solutions concept could be used to predict saturated conductivity and possibly unsaturated conductivities for different salt solutions, at

least for some soils. Although further testing would be necessary, this method, if proved, could provide a useful means of relating and predicting upward flow from groundwaters of different salt composition. For instance, if the values for the equivalent salt solutions are known and the changes in rate of upward flow of salt solutions from a water table with reduction in cation concentration at a given SAR are determined, the flow of salt solutions of any other SAR could be predicted. This method could also have a similar application in predicting the flow of salt solutions under other boundary conditions.

The present study was conducted on confined soil columns. Therefore the results obtained show the effect of salt solutions on the pore geometry, hydraulic conductivities and upward flow in soils whose bulk density remains constant. Most mineral soils do not show bulk swelling in the presence of salt solutions, apparently due to the presence of non-swelling material such as sand, silt and non-expanding lattice clays being held or cemented together by various cementing agents. But if the clay matrix lining the pores of these soils contains some clays of expanding lattice type, a swelling of this clay matrix into the pores in the presence of salt solutions could change the pore geometry, hydraulic conductivity properties and rates of upward flow from a water table in these soils. Presence of salt solution could also change the pore geometry and conductivity properties of highly colloidal soils which show significant changes in their bulk volumes during wetting and drying. However, in considering the effects of salt solutions on upward flow in such soils, the effect of the overburden potential needs to be considered. Experimental studies to measure the effects of salt solutions on upward flow rates in such soils will need to be carried out in the field or in columns of very large diameter (Collis-George, 1961; Lal et al., 1970) or in longitudinally self-expanding laboratory columns (Smiles, 1972; Talsma, 1977).

The present study deals with only one aspect of the effect of salt solution composition on upward flow of soil solution from the water table to the surface. The salt solutions used in this investigation contained sodium, calcium and chloride ions. The presence of other cations, such as magnesium instead of calcium or other anions in the groundwaters, could alter the rate of upward flow. The other possible effects which need to be considered include the precipitation of less soluble salts near the evaporating surface and the effects of salt accumulation near the soil surface on the rate of upward flow.

Several areas for future studies may be briefly identified. This study on the effects of salt solution composition on rates of upward flow and salinisation could be extended to other soils, particularly soils in which a greater degree of clay dispersion occurs. The accuracy of the equivalent salt solutions method for predicting saturated and unsaturated conductivities in salt solutions and for relating rates of upward flow of different salt solutions under specific boundary conditions needs to be evaluated. The development of a quantitative method for determining curves indicating pore size distribution at different pressure heads in a soil with a dynamic internal pore geometry could be of theoretical and practical importance. The applicability of the results obtained and methods developed in the present study to similar studies in the field, needs to be evaluated in future studies.

CONCLUSIONS

1. The pore geometry, saturated and unsaturated conductivity and the capacity of the soil to transmit soil solution from a water table to the surface changed very little in krasnozem soil exposed to different salt solutions, but marked and varied changes occurred in the case of alluvial and red brown soils.
2. With reduction in cation concentration in solutions of a given SAR, the moisture release curves of alluvial and red brown soils were displaced to the right indicating a decrease in size of inter-aggregate pores. This was attributed to the swelling of soil aggregates into this pore space. The relative shift of the moisture release curves in solutions of specific cation concentration was more marked for solutions of higher SAR.
3. A pore size index, given by the ratio of radii of pore necks of common pores in low salt solution to the radii in solutions of the highest salt concentration at the same SAR, has been proposed as a quantitative measure of the changes in moisture release curves due to swelling. The values of this pore size index indicated that the krasnozem soil was the most stable and the red brown soil the least stable to high sodium, low salt solutions.
4. As the cation concentration of the solution at a given SAR was reduced, the volumes of the large sized pore groups, in alluvial and red brown soils, continued to decrease, while the volumes of the medium sized pore groups increased to a maximum and then progressively decreased.

5. As the cation concentration was reduced in solutions of a given SAR, the saturated conductivity and unsaturated conductivity at high pressure heads of alluvial and red brown soils decreased markedly while the unsaturated conductivity at lower pressure heads showed only small decreases.
6. It is proposed that in the presence of equivalent salt solutions, which are defined as solutions with combinations of SAR and cation concentration producing the same extent of swelling, a given soil will have the same pore geometry, the same saturated and unsaturated conductivities and hence the same capacity to transmit salt solutions under specific boundary conditions. The equivalent salt solutions method gave fairly close predictions of saturated conductivity of alluvial and red brown soils for different salt solutions and of the unsaturated conductivity-pressure head curve for a single salt solution in red brown soils. Further testing of this method is necessary.
7. The relative decrease in measured saturated conductivity in alluvial and red brown soil as the cation concentration was reduced at a given SAR, was greater than the relative decrease in saturated conductivity computed from the moisture release curves of the soils in the respective salt solutions, using existing computation methods. Changes in internal pore geometry during desaturation as a result of aggregate shrinkage on drying, could partly contribute to these differences. These changes have been described qualitatively in terms of a theoretical model of the relationship between the moisture release curves and the curves indicating the pore size distribution at different pressure heads.

8. The existing methods for computing unsaturated conductivity from moisture release curves of alluvial and red brown soils could be used at pressure heads between -200 and -500 cm of water, provided that a matching factor from within this pressure head range was employed.

9. The computed maximum depths to water table for specific rates of upward flow in alluvial and red brown soils decreased as the cation concentration was reduced at a given SAR. However, marked decreases in maximum depths to water table for specific rates of upward flow occurred only in solutions of low cation concentration, in the presence of which the values of unsaturated conductivity within the critical range (defined in Appendix A) showed marked decreases.

10. The measured rates of upward flow from specific water table depths in alluvial and red brown soil, showed decreases with decreasing cation concentration in solutions of a given SAR. They differed somewhat from the computed rates, using shorter columns. However, the relative decreases of measured rates using longer columns, were of a similar magnitude to the relative decreases in the computed rates.

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APPENDIX A

THE FLUX/UNSATURATED CONDUCTIVITY RATIO METHOD FOR COMPUTING UPWARD FLOW IN SOILS FROM A WATER TABLE*

Theory

The method proposed in this study is based on an analysis of the k - h curves in relation to their use in the flow equation. This method involves the division of the right-hand side of equation (1) into a number of components. The limits of pressure head range of each of these components correspond to specific values of unsaturated conductivity (Table 17) which, in turn, are determined by the value of the flux/unsaturated conductivity ratios ($q/k[h]$ ratio) for any given rate of upward flow. The absolute values of the limits of the pressure head ranges of each of these components for any given k - h curve also varies according to the rate of flow (Fig. 50). But the general method of analysis for a specific rate of flow described below applies to all rates of flow.

Hence we can write equation (1) as follows,

$$\begin{aligned}
 -z_{\max} = & \int_0^{h_{u1}} \frac{1}{1+q/k[h]} dh + \int_{h_{u1}}^{h_{u2}} \frac{1}{1+q/k[h]} dh + \int_{h_{u2}}^{h_m} \frac{1}{1+q/k[h]} dh \\
 & + \int_{h_m}^{h_{11}} \frac{1}{1+q/k[h]} dh + \int_{h_{11}}^{h_{12}} \frac{1}{1+q/k[h]} dh + \int_{h_{12}}^{h_{13}} \frac{1}{1+q/k[h]} dh \\
 & + \int_{h_{13}}^{h_{14}} \frac{1}{1+q/k[h]} dh + \int_{h_{14}}^{h_{15}} \frac{1}{1+q/k[h]} dh + \dots \dots \dots (2)
 \end{aligned}$$

The value of the first and second components of equation (2) can be very closely approximated by the terms $0.995h_{u1}$ and $0.96(h_{u2}-h_{u1})$ respectively. Further, where the saturated hydraulic conductivity of a soil is greater than k_{u1} and the pressure head/conductivity

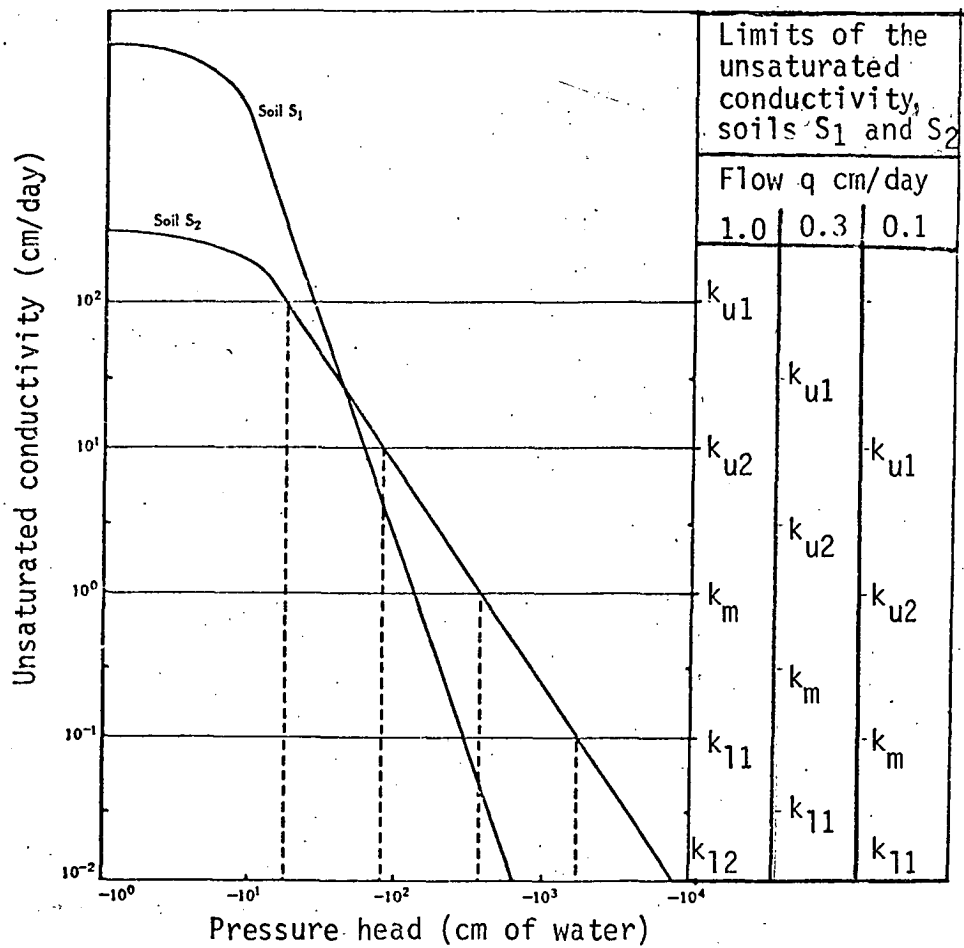
*Sections of this appendix were published as a paper in the Australian Journal of Soil Research, Vol. 15, 17-25, 1977.

Table 17. The limits of the pressure heads and saturated conductivity of the components in equations (2), (3) and (4).

Component	Lower limits of unsaturated conductivity range $k_x[q]^*$	Flux/unsaturated conductivity ratio $q/k[h]$	$\frac{1}{1+q/k[h]}$	Lower limits of range of pressure head $h_x[q]^*$
1	k_{u1}	0.01	0.99	h_{u1}
2	k_{u2}^{**}	0.10	0.9091	h_{u2}
3	k_m	1.00	0.5000	h_m
4	k_{11}	10	0.0909	h_{11}
5	k_{12}^{**}	100	0.0099	h_{12}
6	k_{13}	1000	0.0010	h_{13}
7	k_{14}	10000	0.0001	h_{14}
8	k_{15}	100000	0.00001	h_{15}
n	$k_{1(n-3)}$	10^{n-3}		$h_{1(n-3)}$

*The terms within square brackets indicating the functional dependence will be omitted in the text where a fixed rate of upward flow is considered.

**The upper and lower limits of the critical unsaturated conductivity range are usually given by k_{u2} and k_{12} respectively.



Limits to pressure head ranges of soil S ₂	q 1.0	h _{u1}	h _{u2}	h _m	h _{l1}	h _{l2}
	q 0.1		h _{u1}	h _{u2}	h _m	h _{l1}

Fig. 50. The limits of unsaturated conductivity and pressure head ranges of the components of equation (2) for soils S₁ and S₂, for rates of upward flow of 1.0, 0.3 and 0.1 cm/day.

relationship at pressure heads less than h_{u2} plotted on a log-log scale, is represented by a line of a given uniform slope n , the mean value of $1/(1 + q/k[h])$ is a constant for a given component of equation (2) irrespective of the pressure head range over which it applies. The mean value is defined as the total value of the component divided by the pressure head range.

Hence we can write,

$$-z \text{ max.} = 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + a(h_m - h_{u2}) + b(h_{11} - h_m) + c(h_{12} - h_{11}) \\ + d(h_{13} - h_{12}) + e(h_{14} - h_{13}) + f(h_{15} - h_{14}) + 10^{-1}f(h_{16} - h_{15}) + \dots \dots (3)$$

where a, b, c, d, e, f are the mean values of $1/(1 + q/k[h])$ for components 3, 4, 5, 6, 7, 8 respectively of equation (3) and their empirically determined values can be obtained from Table 18 for $k-h$ curves with slopes n equal to 1.5, 2, 3, 4 and 6 and by graphical interpolation for non-integer values of n . The value of $1/(1 + q/k[h])$ of any particular component in equation (3) increases with the increase in slope of the $k-h$ curves. It is also seen from Table 18 that the mean value of $1/(1 + q/k[h])$ of successive components after the third component for a given $k-h$ curve show an approximately ten-fold decrease. But this decrease is partly balanced by a much smaller increase in the value of the pressure head range of the successive components of equation (3). The inverse ratio (x) of the pressure head range of any two successive components, except for the first component, is constant for a $k-h$ curve of a given uniform slope (Table 18).

Therefore, we can express the value of each component of equation (3) in terms of the pressure head range ($H = h_m - h_{u2}$) of component 3 in equation (3) as shown below,

$$\begin{aligned}
 -z \text{ max.} &= 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + aH + bxH + cx^2H + dx^3H + ex^4H \\
 &+ fx^5H + 10^{-1}fx^6H + \dots \dots \dots (4) \\
 &= 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + (a+bx+cx^2+dx^3+ex^4+fx^5+10^{-1}fx^6+\dots)H \\
 &= 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + C_1H \dots \dots \dots (5)
 \end{aligned}$$

where C_1 is a constant for any given value of slope n . The pressure head range (H) of component 3 in equations (4) and (5) is related to h_{u2} as follows,

$$H = h_{u2}(x - 1) = C_2h_{u2} \dots \dots \dots (6)$$

Substituting equation (6) in equation (4) and solving,

$$\begin{aligned}
 -z \text{ max.} &= 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + aC_2h_{u2} + bxC_2h_{u2} + cx^2C_2h_{u2} \\
 &+ dx^3C_2h_{u2} + ex^4C_2h_{u2} + fx^5C_2h_{u2} + 10^{-1}fx^6C_2h_{u2} + \dots \dots \dots (7) \\
 &= 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + C_1C_2h_{u2}
 \end{aligned}$$

$$-z \text{ max.} = 0.995h_{u1} + 0.96(h_{u2} - h_{u1}) + C_nh_{u2} \dots \dots \dots (8)$$

where $C_2 = (x - 1)$ and $C_n = C_1C_2$ are constants for any given value of slope n . The values of each component of equations (7) and (8) in terms of h_{u2} , for $k-h$ curves with slope n equal to 1.5, 2, 3, 4, and 6 calculated by using the values in Table 18, are given in Table 19. For non-integer values of n , solutions to equations (7) and (8) can be obtained using the appropriate values for the products of the constants taken from Figure 51.

Table 18. The mean value of $1/(1 + q/k[h])$ of the components in equation (3) and the inverse ratio (x) of the pressure head ranges of successive components for k-h curves with slope n.

Slope n	Mean value of $1/(1+q/k[h])$						x
	a	b	c	d*	e*	f*	
1.5	68.7×10^{-2}	21.1×10^{-2}	28.2×10^{-3}	28.6×10^{-4}	28.9×10^{-5}	29×10^{-6}	4.64152
2	70.1×10^{-2}	22.1×10^{-2}	30.2×10^{-3}	31.2×10^{-4}	31.7×10^{-5}	32×10^{-6}	3.16228
3	71.3×10^{-2}	23.3×10^{-2}	32.8×10^{-3}	33.6×10^{-4}	33.9×10^{-5}	34×10^{-6}	2.15444
4	72.0×10^{-2}	24.0×10^{-2}	33.4×10^{-3}	34.4×10^{-4}	34.8×10^{-5}	35×10^{-6}	1.77828
6	73.0×10^{-2}	25.1×10^{-2}	34.1×10^{-3}	34.8×10^{-4}	35.5×10^{-5}	35.5×10^{-6}	1.46780

*The approximations $f \approx 10^{-1}$ $e \approx 10^{-2}$ d may be used with a very small error.

Table 19. The values of the components in equation (7) in terms of h_{u2} for k-h curves with uniform slope n.

Slope n	Total value of component/ h_{u2}									C_n
	C_2a	C_2bx	C_2cx^2	C_2dx^3	C_2ex^4	C_2fx^5	$C_210^{-1}fx^6$	$C_210^{-2}fx^7$	$C_210^{-3}fx^8$	
1.5	2.502	3.565	2.212	1.041	0.488	0.226	0.106	0.047	0.023	10.230
2	1.516	1.511	0.653	0.214	0.069	0.022	0.0069	0.0022	0.00069	3.996
3	0.823	0.580	0.175	0.039	0.0084	0.0018	0.00039	0.000085	0.000018	1.628
4	0.560	0.332	0.082	0.015	0.0027	0.00048	0.000081	0.000016	0.0000028	0.993
6	0.341	0.173	0.034	0.0052	0.00077	0.00011	0.000017	0.0000024	0.00000036	0.554

Table 20. The values of k_z for a rate of upward flow of 0.1 cm/day ($k_z[q_{0.1}]$) for k-h curves with uniform slope n.

Slope n	1.5	2	3	4	6
C_n	10.230	3.996	1.628	0.993	0.554
$k_z[q_{0.1}]$	0.026	0.040	0.055	0.063	0.071

This method can thus be used to compute pressure head profiles using equations (3) or (7) and maximum depth to water table for any rate of upward flow (equation (8)) from the values of pressure heads h_{u1} , h_{u2} , etc., and the slope n of the k - h curve, all of which can be obtained directly from the k - h curves. It is possible to extend this method to analyse k - h curves with any integer or non-integer values of the slopes within or outside the range of values discussed.

Further simplification can be achieved by combining components 1 and 2 in equations (3), (7) and (8). The mean values of $1/(1+q/k[h])$ for combined components 1 and 2 for different k - h curves, ranges from slightly less than 1.0, where the saturated conductivity is very much greater than k_{u2} , to 0.91 where the saturated conductivity is equal to k_{u2} and air entry value is h_{u2} . Although the appropriate value can be easily estimated, an average value of 0.95 may be used which will generally result only in a small error.

Hence,

$$-z_{\max} \approx (0.95 + C_n) h_{u2} \quad (9)$$

Thus, if the depth to the water table is known, we can use equation (9) to find the value of h_{u2} and from the k - h curves we can determine k_{u2} and hence the corresponding rate of upward flow. Since in this analysis the upward flow is chiefly characterized by only two parameters, h_{u2} and n , whose direct inter-relationship is given by equation (9), it provides a basis for comparing rates of upward flow in different soils.

In spite of the use of simplifying assumptions, this method gives computed rates of upward flow which were almost as accurate as those obtained from numerical analysis, and yields quick solutions using hand calculations.

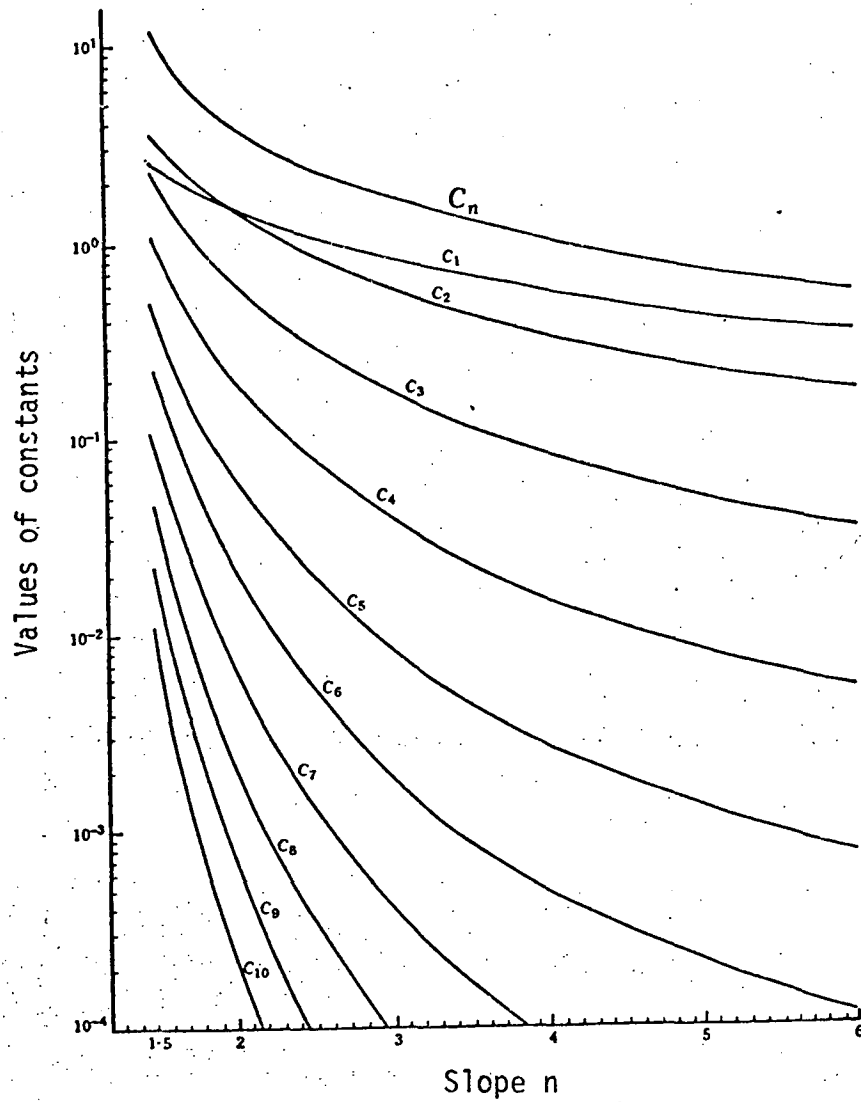


Fig. 51. The values of constants in components of equation (4) for k-h curves with slopes n between 1.5 and 6.

Furthermore, in soils where the saturated conductivity is very much greater than k_{u2} , the maximum depth to water table (z_{\max}) is given by the value $(1+C_n)h_{u2}$. It can be shown that

$$k_z[q_x]/k_{u2}[q_x] = (h_{u2}[q_x])^n (h_z[q_x])^{-n} \quad (10)$$

where $k_z[q_x]$ is the value of unsaturated conductivity at a pressure head $h_z[q_x]$ which numerically equals the value of the maximum depth to the water table ($z_{\max}[q_x]$) for a given rate of upward flow q_x . Substituting the value of z_{\max} for h_z in equation (10),

$$k_z[q_x] = 10q_x(1+C_n)^{-n} \quad (11)$$

The value of k_z for a rate of upward flow 0.1 cm/day ($k_z[q_{0.1}]$) for k-h curves of a given slope (Table 20) can be calculated using

$$k_z[q_{0.1}] = (1+C_n)^{-n} \quad (12)$$

Therefore the maximum depth to the water table for any rate of upward flow can be read directly from the k-h curve. In soils where the mean value of $1/(1+q/k[h])$ of the first two components is less than one, to get the value of maximum depth to the water table using equation(12), a small correction (which can be calculated) has to be applied to account for the lower mean value of $1/(1+q/k[h])$.

Equation (7) and other equations derived from it can be applied directly only to soils with saturated conductivity exceeding k_{u2} , and at conductivities below k_{u2} the pressure head/conductivity relationship plotted on a log-log scale has a uniform slope. These conditions are usually satisfied in many soils used in laboratory and field studies, for the range of rates of upward flow which are important (0.1 - 2.0 cm/day). However, for soils with variable slopes, but slopes which can be represented by a straight line within

the limits of the range of each component, we can apply equation (7), in which case a hypothetical value for h_{u2} for each component has to be obtained by extrapolating the k-h curve of this component to intercept k_{u2} and also using the constants which are applicable for that particular value of slope n. For soils with saturated conductivity less than k_{u2} or k-h curves with variable, non-uniform slopes, the values of components which cannot be represented by a straight line can be calculated by numerical analysis, and these values can be added to those of other components using equation (3) or (7).

Generalized Equations for Different Rates of Upward Flow

A feature of the method of analysis discussed is that the value of parameter h_{u2} is determined by the rate of upward flow that is considered. However, generalized equations for different rates of upward flow q_x may be obtained from equations (7), (8) and (9) using the relationship,

$$h_{u2}[q_x] = h_{u2}[q_{1.0}]/(q_x)^{1/n} \quad (13)$$

where $h_{u2}[q_x]$ and $h_{u2}[q_{1.0}]$ are the values of h_{u2} for upward flow rates of q_x and 1.0 cm/day respectively, provided that both these values lie on the portion of the k-h curves with uniform slope.

Thus, substituting equation (13) in equation (9),

$$-z_{\max}[q_x] \approx (C_n + 0.95)h_{u2}[q_{1.0}]/(q_x)^{1/n} \quad (14)$$

where $z_{\max}[q_x]$ is the maximum depth to water table for any rate of upward flow q_x .

Critical Range

In computing the rate of upward flow of water from a water table using k-h curves, due to the experimental difficulties in determining the values of unsaturated conductivity in soils, it is useful to define the critical range of unsaturated conductivity and the

corresponding critical pressure head range over which the value of $k[h]$ significantly affects the shape of pressure head profiles and the computed rates of upward flow.

Gardner (1958) and Philip (1957) have shown that the k - h relationship at very low pressure heads is not important in determining upward flow. Talsma (1963) showed that the lower limit of pressure head which significantly affected the computed values of maximum depth to the water table for given rates of upward flow, varied in different soils from -300 to -2000 cm of water. It is seen from Figure 51 that for k - h curves with slope greater than 2 the relative value of each successive component decreases and the rate of decrease is greater with increase in value of slope of curve. For the curves with slopes greater than 2, we can define the lower limit of the critical pressure head range as h_{12} , since the relative values of the succeeding components are small. For soils with slope of k - h curves less than 2, the lower limit of the critical range can be taken as h_{13} . Wind (1955) found that the exact values of $k[h]$ of soils at and near saturation is not very important in determining upward flow, but from equation (3) we can see that this is not applicable for soils with saturated conductivity less than k_{u2} . The upper limit of the critical pressure head range can be defined as h_{u2} , since at pressure heads greater than h_{u2} the absolute values of unsaturated conductivity have little effect on the shape of the computed pressure head profile. Hence for k - h curves with slope n greater than 2, the critical unsaturated conductivity range and the corresponding critical pressure head range can be defined as k_{u2} to k_{12} and h_{u2} to h_{12} respectively.

The critical pressure head range as defined above applies to a

specific rate of upward flow, but it can be defined for a range of flows, in which case h_{u2} is defined for the highest rate of flow and h_{12} is defined for the lowest rate of flow.

APPENDIX B

A THEORETICAL MODEL OF THE CHANGES IN THE PORE SIZE DISTRIBUTION AT DIFFERENT PRESSURE HEADS IN A CONFINED SOIL, SHOWING ALTERATIONS IN ITS INTERNAL PORE GEOMETRY DUE TO SWELLING AND SHRINKAGE OF SOIL AGGREGATES.

In the present study it was observed that a swelling of aggregates of alluvial and red brown soils into the interaggregate pore space occurred in the presence of high sodium, low salt solutions. On application of a suction to these soils initially saturated with such salt solutions, it is likely that shrinkage of individual soil aggregates could occur, in addition to drainage of pores. Therefore, as such soils dry in response to increasing applied suction during the moisture release curve determination, there could be continuous changes in the internal pore geometry of the soils.

Childs (1940) showed that in a rigid soil of fixed internal pore geometry, the experimentally determined moisture release curve indicated the volumes of pores of specific pore neck size ranges and that these curves could be used to derive the pore size distribution of the soil. The pores which drain on application of a specific increment of suction, during moisture release curve determination, will have an average pore neck radius corresponding to the average suction applied. However, the same pore will have different radii when the soil is subject to other suctions, if the individual soil aggregates shrink with drying on application of a suction. Hence in such soils the "hypothetical moisture release curves" indicating the volumes of pores of specific pore neck size ranges when the soil is at different pressure heads (which will be referred to as pore size distribution indicating (PSDI) curves) will vary from each other and from the experimentally determined moisture release curve, unlike in a soil of fixed internal pore geometry.

A simplified, theoretical model for the pore size distribution indicating (PSDI) curves (Figure 1) at different pressure heads in such internally swelling soils may be developed in the following manner. At saturation all pores will be smaller than indicated by the moisture release curve and hence the PSDI curve at saturation will lie to the right of the moisture release curve. The extent of this shift to the right will be determined by the magnitude of the shrinkage of the soil aggregates on drying. At a slightly lower pressure head h_1 , the moisture release curve will give the experimentally determined water-filled porosity at pressure head h_1 , and hence the PSDI curve at the same pressure head should pass through it. The pores undrained at this pressure head will, on account of aggregate swelling, have radii smaller than indicated by the moisture release curve but larger than shown in the PSDI curve for the saturated soil and will probably be given by a curve of intermediate slope. But the drained pores will have radii larger than shown by the moisture release and PSDI curves at saturation and will be given by a curve which lies to the left of these curves.

At a lower pressure head h_2 , the PSDI curve at h_2 passes through the moisture release curve at h_2 . The pore size distribution of undrained pores at this pressure head will be given by a curve lying between the moisture release curve and the PSDI curves at pressure head h_1 and will probably have an intermediate slope. The drained pores at this pressure head h_2 , will have radii larger than the moisture release curve and the PSDI curve at pressure head h_1 and will be given by a curve which lies to the left of these curves. Similarly, a series of pore size distribution indicating (PSDI) curves at progressively lower pressure heads may be constructed. The difference

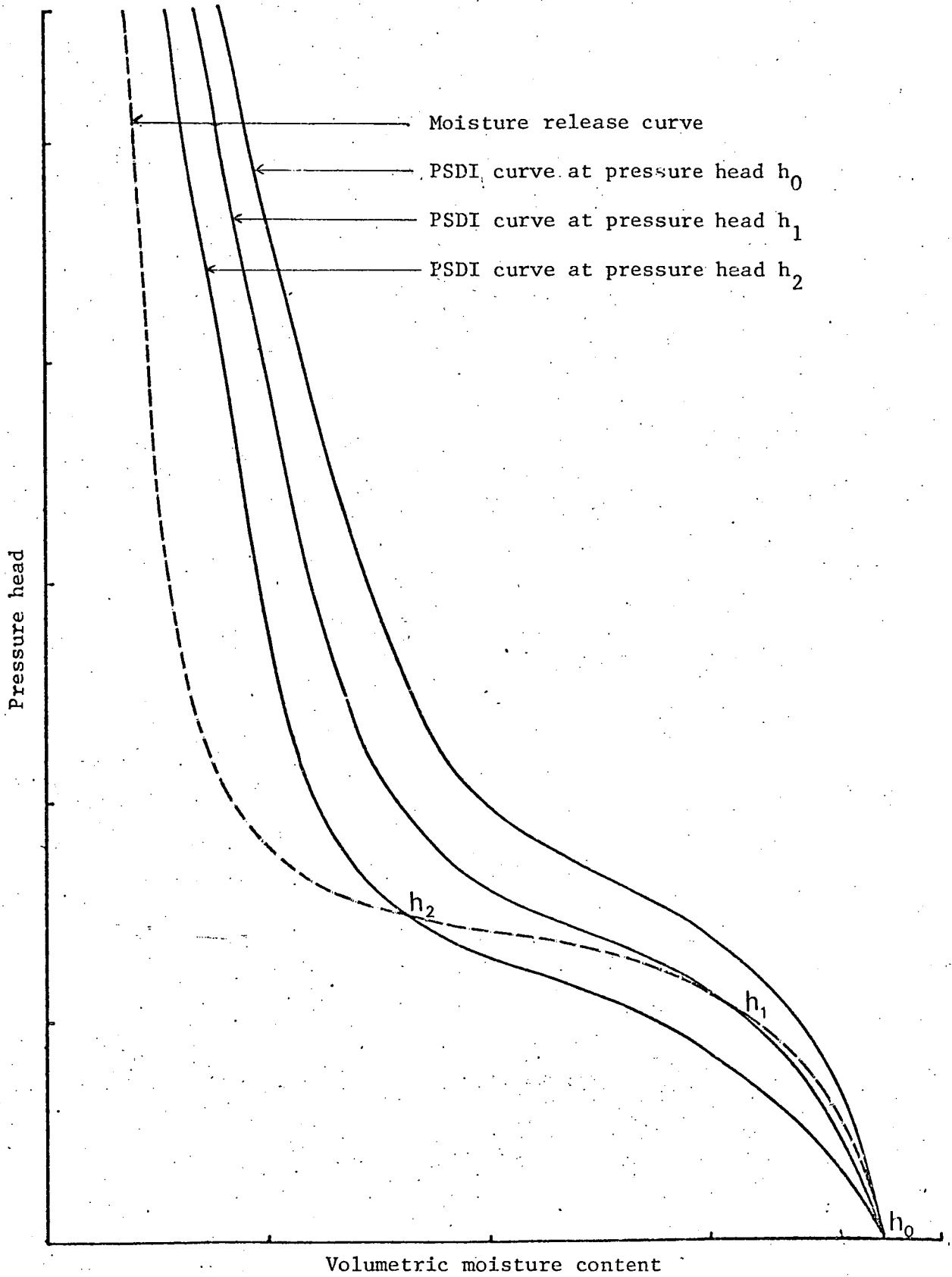


Fig. 52. Theoretical relationship between the moisture release curve and the curves indicating the pore size distribution at different pressure heads in a soil showing swelling which is internally accommodated. The moisture release curve is determined experimentally by applying increments of suction and measuring the outflow. The curves indicating the pore size distribution (PSDI curves) at different pressure heads h_0 (saturation) $> h_1 > h_2$ are hypothetically derived curves.

between the moisture release curve and the PSDI curves of undrained pores at specific pressure heads will become progressively smaller as the pressure head is decreased. Figure 51 thus illustrates the gradual changes in the pore size distribution as this internally swelling soil dries. If the unsaturated conductivity of this soil at any specific pressure head needs to be computed, instead of the moisture release curve of the soil, the PSDI curve at this specific pressure head should be used.

In solutions of high salt concentration at a given SAR, since the shrinkage of soil aggregates as the soil dries is small, the shift of the PSDI curves at all pressure heads is likely to be small. But as the salt concentration is decreased at the same SAR, the shrinkage of the soil aggregates on drying is likely to be relatively greater and the relative shift of the PSDI curve at all specific pressure heads will be larger than in the high salt solution.

A method for computing the pore size distribution curves at different pressure heads from the experimentally determined moisture release curves needs to be devised, possibly by using the shrinkage characteristics of soil aggregates on drying. However, this theoretical model itself could prove to be useful in predicting, comparing and explaining, on a qualitative basis, the differences in experimentally determined conductivity properties of these soils which show changes in their internal pore geometry and the conductivities computed from the moisture release curves of the same soils. However, it must be stressed that this model is purely hypothetical and it also does not provide a quantitative evaluation of the importance of shrinkage of soil aggregates on conductivity characteristics of soils.

APPENDIX C

THE CONCENTRATIONS OF CALCIUM AND SODIUM (meq/l) PRESENT IN THE
DIFFERENT SALT SOLUTIONS.

SAR	CATION CONCENTRATION (meq/l)	CALCIUM (meq/l)	SODIUM (meq/l)
40	640	220.2440	419.7560
	160	23.3440	136.6560
	80	6.7136	73.2864
	40	1.8220	38.1780
	20	0.4765	19.5235
	10	0.1220	9.8780
	2.5	0.0078	2.4922
20	640	368.5165	271.4835
	160	55.0610	104.9390
	80	18.7549	61.2452
	40	5.8360	34.1640
	20	1.6784	18.3216
	10	0.4555	9.5445
	5	0.1191	4.8809
5	2.5	0.0305	2.4695
	640	556.5891	83.4109
	160	121.0940	38.9060
	40	23.0322	16.9678
	10	3.4413	6.5587
0	2.5	0.3647	2.1353
	160	160.0000	-
	40	40.0000	-
	10	10.0000	-
	2.5	2.5000	-

THE VOLUMES OF THE COMMON INTER-AGGREGATE PORES PER UNIT VOLUME OF KRASNOZEM, ALLUVIAL AND RED BROWN SOILS IN DIFFERENT SALT SOLUTIONS.

SOIL	KRASNOZEM			ALLUVIAL				RED BROWN			
CATION CONCENTRATION (meq/l)	SAR			SAR				SAR			
	0	20	40	0	10	20	40	0	10	20	40
640			0.222				0.130				0.150
160	0.274	0.286	..	0.189	0.164	0.188	0.106	0.182	0.132	0.148	0.104
40	0.150	0.194	0.054	..	0.110	0.138	0.056
10	0.176	0.120	0.138	<0.010	0.158	0.102	0.028	<0.010
2.5	0.264	0.268	0.205	0.160	0.070	0.016	<0.010	0.144	0.038	<0.010	<0.010

APPENDIX D

THE MOISTURE RELEASE DATA* OF KRASNOZEM, ALLUVIAL AND RED BROWN SOILS, PACKED TO BULK DENSITIES OF 1.00, 1.22 and 1.22 g/cm³ RESPECTIVELY, FOR SALT SOLUTIONS OF SAR 20 AND DIFFERENT CATION CONCENTRATIONS.

PRESSURE HEAD (cm of water)	VOLUMETRIC MOISTURE CONTENT (cm ³ /cm ³)							
	KRASNOZEM		ALLUVIAL			RED BROWN		
	Cation concentration (meq/l)		Cation concentration (meq/l)			Cation concentration (meq/l)		
	160	2.5	160	40	10	160	40	10
0	0.656	0.654	0.543	0.542	0.542	0.578	0.577	0.578
-10	0.619	0.611	0.532	0.536	0.538	0.558	0.561	0.576
-20	0.488	0.500	0.511	0.521	0.531	0.531	0.538	0.570
-30	0.450	0.460	0.467	0.478	0.507	0.515	0.522	0.565
-50	0.426	0.436	0.433	0.445	0.465	0.505	0.513	0.558
-80	0.416	0.427	0.420	0.430	0.437	0.498	0.504	0.550
-100	0.412	0.422	0.415	0.424	0.429	0.495	0.500	0.546
-120	0.410	0.420	0.412	0.420	0.424	0.492	0.497	0.543
-140	0.407	0.417	0.410	0.418	0.421	0.490	0.495	0.540
-160	0.405	0.413	0.408	0.415	0.418	0.489	0.494	0.538
-200	0.401	0.408	0.404	0.410	0.414	0.485	0.490	0.534
-500	0.378	0.383	0.383	0.384	0.392	0.460	0.467	0.497

*Moisture release data obtained from soil columns used in the experiments to determine the unsaturated conductivity-pressure head curves.

COMPUTED VALUES OF UNSATURATED CONDUCTIVITY OF KRASNOZEM, ALLUVIAL AND RED BROWN SOILS, PACKED TO BULK DENSITIES OF 1.00, 1.22 AND 1.22 g/cm³ RESPECTIVELY, FOR SALT SOLUTION OF SAR 20 AND DIFFERENT CATION CONCENTRATIONS.

PRESSURE HEAD (cm of water)	KRASNOZEM		ALLUVIAL			RED BROWN		
	CATION CONCENTRATION (meq/l)		CATION CONCENTRATION (meq/l)			CATION CONCENTRATION (meq/l)		
	160	2.5	160	40	10	160	40	10
0	6053	5993	733	533	302	848	552	27
30	25	25	49	62	86	10	10.4	6.3
60	2.1	2.1	2.9	4.4	6.8	2.25	2.70	3.4
100	0.73	0.73	1.0	1.34	1.52	1.12	1.48	2.2
150	0.34	0.34	0.57	0.75	0.90	0.73	0.95	1.55
200	0.21	0.21	0.41	0.53	0.64	0.56	0.70	1.22
300	0.11	0.11	0.26	0.33	0.41	0.39	0.44	0.83
500	0.05	0.05	0.16	0.18	0.22	0.23	0.24	0.45

APPENDIX F

Values of computed unsaturated conductivity given above were read from unsaturated conductivity-pressure head curves of the soils for the specific salt solutions. The data for plotting this curve was calculated from the moisture release curves of the soils for the specific salt solutions (Appendix E) using the equation proposed by Kunze et al. (1968).