THE CONTROL OF SALINE GROUNDWATER

T. TALSMA

Dit proefschrift met stellingen van

TJEERD TALSMA

landbouwkundig ingenieur, geboren te Wanswerd, 14 januari 1928, is goedgekeurd door de promotor, Dr. W. R. VAN WIJK, hoogleraar in de natuurkunde, meteorologie en klimatologie.

De Rector Magnificus der Landbouwhogeschool
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Wageningen, 25 september 1963

THE CONTROL OF SALINE GROUNDWATER

(MET EEN SAMENVATTING IN HET NEDERLANDS)

PROEFSCHRIFT

TER VERKRIJGING VAN DE GRAAD
VAN DOCTOR IN DE LANDBOUWKUNDE
OP GEZAG VAN DE RECTOR MAGNIFICUS IR. W. F. EIJSVOOGEL,
HOOGLERAAR IN DE HYDRAULICA, DE BEVLOEIING,
DE WEG- EN WATERBOUWKUNDE EN DE BOSBOUWARCHITECTUUR,
TE VERDEDIGEN TEGEN DE BEDENKINGEN
VAN EEN COMMISSIE UIT DE SENAAT
DER LANDBOUWHOGESCHOOL TE WAGENINGEN
OP VRIJDAG, 25 OKTOBER 1963 TE 16 UUR

DOOR

T. TALSMA

H. VEENMAN EN ZONEN N.V. – WAGENINGEN – 1963

STELLINGEN

I

De geographische verspreiding van oplosbaar zout langs de middenloop van de Murrumbidgee duidt op een overwegende fluviatiele afzetting van de bovenste bodemlagen; de mogelijkheid van omvangrijke aeolische afzettingen moet daarom in twijfel worden getrokken.

B. E. BUTLER, Parna – an aeolian clay. Aust. J. of Sci. 18: 145–151, 1956.
T. LANGFORD-SMITH, The dead river systems of the Murrumbidgee. The Geograph. Rev., Vol. L: 368–389, 1960.
Dit proefschrift.

II

De verbouw van katoen als alternatief voor de rijstteelt op de zware kleigronden in de irrigatiegebieden langs de Murrumbidgee is, in verband met de vrij geringe hoeveelheid beschikbaar bodemvocht slechts mogelijk indien frequent bevloeid wordt. De hieruit voortvloeiende consequenties voor ziekte en onkruidbestrijding zullen tot gevolg hebben dat de rentabiliteit van deze teelt minder gunstig zal zijn dan in de V.S. vaak het geval is.

III

De waarde van de verzonken verdampingspan voor het bepalen van de potentiële verdamping wordt door dit onderzoek bevestigd.

Dit proefschrift.

IV

Voor vele geïrrigeerde gewassen wordt de meest economische opbrengst bereikt bij een watervoorziening die belangrijk minder kan zijn dan de voor een maximum opbrengst benodigde – gelijk of iets minder dan de potentiële evapotranspiratie – watervoorziening. Het is gewenst bij de bepaling van de benodigde hoeveelheid irrigatiewater voor nieuw te irrigeren gebieden hiermee rekening te houden.

J. WESSELING en W. R. VAN WIJK, Soil physical conditions in relation to drain depth. Am. Soc. Agron., Monograph. VII (J. N. Luthin, Ed.) 461–504, 1957.
O. W. ISRAELSEN, Irrigation principles and practices, 2nd Ed. John Wiley and Sons, 1950.

V

Oppervlakkige grondbewerking en aanpassende teeltmaatregelen zijn meestal van ondergeschikt belang voor de bestrijding van het verzoutingsgevaar in irrigatiegebieden. Deze kunnen pas dan een gunstige invloed hebben wanneer zij gecombineerd worden met geschikte drainagemaatregelen.

D. B. KRIMGOLD, KOSTIAKOV on prevention of waterlogging and salinity of irrigated land. Agr. Eng. 26: 327–328, 1945.
Dit proefschrift.

Proefschrift van T. TALSMA
Wageningen, oktober 1963

VI

De mogelijkheid om de doorlatendheid van gronden met een hoog colloïdgehalte tot op grotere diepte in stand te houden of te verbeteren, door in de gewassenrotatie een of meer sterk bodemuitdrogende gewassen op te nemen, verdient nadere studie.

W. W. EMERSON, The rate of water uptake of soil crumbs at low suctions. J. Soil Sci. 6: 147–159, 1955.

VOORWOORD

Bij het afsluiten van dit proefschrift neem ik gaarne de gelegenheid te baat om allen, die tot de voltooiing hiervan hebben bijgedragen, hartelijk te danken.

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Officers of the N.S.W. Water Conservation and Irrigation Commission readily supplied the background information necessary for this project and it is a pleasure to acknowledge their comments on some of the early results of this investigation.

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THE CONTROL OF SALINE GROUNDWATER

(met een samenvatting in het Nederlands)

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CONTENTS

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LIST OF PRINCIPAL SYMBOLS

(Generally only those symbols occurring in more than one section of this paper are listed. There is some duplication in order to preserve the notation of the original references).

a	soil constant in equation (17)	cm^{n+1} day^{-1}
b	soil constant in equation (17)	cm^{n+1}
B	chloride concentration per 100 gr. dry soil	%
c	pan coefficient	
C_{Cl^-}	chloride concentration in soil moisture	g l^{-1}
e	saturation vapour pressure, e_a: at mean air temp., e_o: at dew point, e_s: determined from temp. means	mm Hg
E	evaporation rate from soil, also in the steady state case: vertical flow velocity of soil water	cm day^{-1}
E_a	actual evapotranspiration	cm day^{-1}
E_{Aust}	evaporation from an Australian standard pan	cm day^{-1} or cm $month^{-1}$
E_0	evaporation from an open water surface, neglecting heat storage; here termed potential evaporation	cm day^{-1}
E_s	evaporation rate from nearly saturated soil	cm day^{-1}
E_t	potential evapotranspiration	cm day^{-1}
$E(z_w)$	maximum capillary flow rate (in the steady state also evaporation rate) possible through soil with watertable at depth z_w	cm day^{-1}
f	ratio E_t/E_0	
k, k_{sat}	hydraulic conductivity	cm day^{-1}
k, k_{unsat}	capillary conductivity	cm day^{-1}
n	soil constant in equation (17)	
n/N	ratio of actual to possible hours of sunshine	
p	ratio E_a/E_t	
q	capillary flow rate, in steady state equal to E	cm day^{-1}
R	radiation, R_s = shortwave, R_a = outside atmosphere, R_0 = clear day radiation on a horizontal surface, R_l = net longwave radiation	cal cm^{-2} day^{-1} or mm day^{-1}
S	soil moisture suction	cm
W, θ_v	volumetric moisture content	g cm^{-3} or %

Z	chloride content in a soil column with unit surface area	g cm^{-2}
z, z_w	vertical coordinate, watertable depth	cm
θ_w	moisture content per 100 gr. dry soil	%

INTRODUCTION

In almost every major irrigation scheme it has been found that large scale drainage works are essential if intensified agriculture in semi-arid and arid areas is to be of a permanent nature. Drainage problems result from the fact that it is hardly ever feasible to achieve full efficiency in the process of irrigation. For example losses by seepage from the channel system, use of excess irrigation water and the incidence of rainfall during the irrigation season contribute to the drainage problem. Even if full efficiency were possible there is often the need to use a certain amount of irrigation water for leaching out residual salts from the irrigation water. The soils of semi-arid and arid areas often contain soluble salts, which are redistributed by a developing watertable and appear at the surface where the watertable approaches a dangerous level.

It is therefore clear that disposal of a certain quantity of excess water to a depth specified by aeration requirements of the root zone is not the only factor in designing drainage systems in irrigation areas. In addition, excess salts accumulating in the root zone either from irrigation water containing salt or from saline groundwater at shallow depth, should be leached out to maintain a favourable salt balance in the soil profile.

The field investigations reported in this study were carried out in the Murrumbidgee Irrigation Areas of south-east Australia, where the salinity problem is caused by the occurrence of saline groundwater in certain districts, including large areas with rather fine textured soils. This aspect of the drainage problem in semi-arid and arid areas and in particular the physical processes involved in the transfer, in different soil types, of moisture and salt from saline groundwater present at shallow depth, will be treated in considerable detail.

A discussion is presented in Chapter I on the general drainage requirements to control the salinity factor. The required discharge rate is related to the leaching requirements by taking account of the consumptive use of water and salt tolerance of the crop and salinity content of the irrigation water. The desirable drainage depth is discussed in relation to soil physical properties, salinity of groundwater and climatic conditions.

The physiography, geology and original salinity status of various soil groups of the Murrumbidgee Irrigation Areas, together with the properties and effects of the irrigation water, are briefly described in section 1 of Chapter II. Climatic data for this area are described in section 2. The potential evaporation is derived from meteorological records with the aid of the Penman formula and from an examination of pan evaporimeter results. Potential and actual evapotranspiration rates are calculated for pasture and citrus. Section 3 describes the salinization process and the physical properties of soils studied on five experimental sites in the Murrumbidgee Irrigation Areas. The data of sections 2 and 3 are used in section 4 for a comparison with the theory of steady state moisture movement through unsaturated soil, considering the flow of moisture to occur both in the liquid and vapour phase.

The salinization potential of a soil profile is assessed in Chapter III by considering the maximum possible salt transport as dependent on the moisture conducting properties of the soil and on the potential evaporation. Potential salinization of various soils is expressed in terms of the critical depth of watertable. The effect of groundwater salinity is shown in some examples of actual salinization, using the data of Chapter II. These examples also illustrate the principles involved and enable an outline to be given of possible procedures to control salinization from saline groundwater.

I. DRAINAGE REQUIREMENTS TO CONTROL THE SALINITY FACTOR

1. General

The requirements for drainage in general include both the quantity of water to be disposed of and the adequacy of drainage. In humid regions, without salinity problems, it is in principle possible to derive the quantity of drainage water, also termed drainage coefficient or discharge coefficient, from annual, seasonal or periodic rainfall excesses (*e.g.* VISSER, 1953). Where these excesses are absent or unimportant, as is usually the case in semi-arid and arid areas, the drainage coefficient must be derived from unavoidable irrigation excesses (ISRAELSEN, 1950) or from leaching requirements (REEVE, 1957). The latter approach aims directly at control of the salinity factor and will be discussed in section I:2.

The adequacy of drainage may be further defined by specifying in humid regions control of the watertable to a minimum allowable depth during periods of excess precipitation or by calculating an optimal depth which, on the average, will ensure adequate water and air supply of the root zone (WESSELING and VAN WIJK, 1955). In irrigated areas a similar specification may be adopted regarding the minimum depth for the prevention of waterlogging, but determination of the desired depth of drainage usually relates directly to the prevention of salinization of the root zone by saline groundwater. Data from literature relevant to this subject will be discussed in section I:3.

There appears to be no common criterion for the control of the salinity factor in the literature on drainage of irrigated land. The main emphasis in literature from the USA has been on the control of salts brought in by the irrigation water, but little attention has been paid to the depth to which the watertable should be controlled. RICHARDS (1956) makes the general statement that, given adequate water supply and reasonably favourable soil texture, it may well be argued that in arid climates the hazards from a watertable near the root zone considerably outweigh any advantages connected with sub-irrigation. Separation of the watertable and associated capillary fringe from the root zone is therefore desirable. This approach will be of benefit in areas where irrigation water contains an appreciable amount of salt and where watertable control to a great depth is possible.

A somewhat different approach towards drainage requirements for the prevention of salinization is apparent in the literature from the USSR. According to KOVDA (1957, 1961) and AVERIANOV (1957) the task of drainage to improve saline soil consists of:

a. In the first stage or reclamation period: desalinization of the root zone to a pre-determined extent (0.2–0.3% total salt, Cl^- less than 0.01%), followed by gradual desalinization of the groundwater (to 0.2–0.3% total salt). This stops salinization independently of the depth of watertable. Finally the building up of a fresh water layer 5–10 meters thick. Installation of deep, widely spaced drains is recommended for this purpose. The whole process may take from 10–30 years.

b. In the second stage or "exploitation" period, the first task is to maintain a net downward movement of drainage water to prevent resalinization and to wash down excess soluble salts brought in by irrigation water. The second task is to supply fresh groundwater to the plant roots.

A main objective appears to be the introduction of a fresh water layer on top of the saline groundwater. It is stated that the aim should not be to separate the capillary fringe from the root zone, which would mean depriving the crop of 10–30% of its water requirement, or alternatively the use of more irrigation water (AVERIANOV, 1957; LEGOSTAEV, 1958; KRIMGOLD, 1945).

This method of drainage appears quite sound where the soil remains permeable to great depths. Here, in the case of deep, widely spaced drains, drainage and therefore desalinization of the deep subsoil layers is possible, but when drains are shallow and closely spaced this method is not nearly so effective. From results of theoretical work (*e.g.* VAN DEEMTER, 1950) it can be easily shown that for practical purposes nearly all the water flowing to the drains passes through the region delineated by the watertable and a depth below the drains $H \cong 1/6\,s$, where s is the drain spacing. Desalinization is therefore limited to a depth below the drains equal to about $1/6\,s$. For drains at a depth, d, of 0.60 m and spaced at 9 m intervals one obtains therefore a desalinization depth $d + H = d + 1/6\,s = 2.10$ m, but for drains placed 2 m deep and spaced 120 m apart desalinization to 22 m is possible. Soils permeable to a great depth appear to be rather common in the irrigation areas of the USSR (KOVDA, personal communication); this is rare in Australia where slowly permeable layers often occur at shallow depth (TALSMA and FLINT, 1958, TALSMA and HASKEW, 1959).

2. DERIVATION OF THE DRAINAGE COEFFICIENT FROM LEACHING REQUIREMENTS

Considerable effort has been made in the USA (REEVE, 1953 and 1957, RICHARDS, 1954) to derive quantatively an average drainage rate from leaching requirements of crops grown under irrigation. Assumptions underlying this work are that free drainage through the soil is assured (*i.e.* no watertable or restricting layers exist) and that salts concentrating in the root zone originate from evenly distributed irrigation water which, when applied to the land is used by the plants and leaves a salt residue behind. From the law of conservation of mass it follows that:

$$D_{iw}\, C_{iw} - D_{dw}\, C_{dw} - S_s - S_c = 0 \qquad (1)$$

where D_{iw} and D_{dw} are the volumes of irrigation and drainage water respectively, C_{iw} and C_{dw} are the corresponding salt concentrations, S_s is the quantity of salt precipitated in the soil and S_c is the quantity of salt removed by the crop. On a long term basis S_s becomes very small, whilst S_c is always very small. Therefore (1) reduces to:

$$D_{dw}/D_{iw} = C_{iw}/C_{dw} \tag{2}$$

The quantity of irrigation water D_{iw} is related to consumptive use, D_{cw} and the volume of drainage water D_{dw} by:

$$D_{iw} = D_{cw} + D_{dw} \tag{3}$$

Equations (2) and (3) combined give:

$$D_{dw} = D_{cw}\, C_{iw}/(C_{dw} - C_{iw}) \tag{4}$$

HILL (1961) uses the same approach but replaces C_{dw} by C_s, which is the average concentration of salts in the soil solution. He relates the two quantities by $C_s = (C_{iw} + C_{dw})/2$, and thus obtains the relation:

$$D_{dw} = D_{cw}\, C_{iw}/(2C_s - 2C_{iw}) \tag{5}$$

The use of equations (4) or (5), in combination with salt tolerance classes of various agricultural crops, allows calculation of the minimum amount of drainage water to be discharged as a long time average or steady state rate, to maintain a favourable salt balance in the root zone. In the choice of a figure indicating salt tolerance of a crop, C_s, as the average solute concentration, would appear to be the appropriate one. In the steady state case of continuous irrigation, evapotranspiration and drainage this would indeed be so, however irrigation is intermittant and the concentration $C_s = (C_{iw} + C_{dw})/2$ is usually only approached immediately following irrigation, after which the solute concentration increases gradually to values beyond C_{dw}. Equation (4) is therefore somewhat more realistic and will be used in the following examples. Comparing eq. (4) and (5) it is seen that by so doing the drainage requirement is twice that of the ideal steady state case.

As an example may be taken the irrigation of a citrus orchard in the Murrumbidgee Irrigation Areas, N.S.W. The maximum concentration of soluble salts in the irrigation water is 0.01 %, whilst citrus has a consumptive use of approximately 900 mm irrigation water over a period of 200 days annually and can tolerate 0.1 % soluble salt (calculated on dry soil basis) in the lower root zone. This gives a minimum drainage requirement of 100 mm, or a drainage rate of 0.5 mm day^{-1}. On the basis of excess rainfall and irrigation this particular orchard has been drained with a rate of 4 mm day^{-1}, therefore the leaching requirement is entirely satisfied.

Examples of minimum leaching requirements given by REEVE (1957) (see also PRUNSTER, 1956) require discharge rates varying from about 0.5–2.5 mm day^{-1} during the growing season. The lower figure relates to irrigation water of high quality, salt tolerant crops or a low consumptive use. AVERIANOV (1957) gives an average annual discharge rate, in Russian literature termed "coefficient of filtration" or "drainage modulus", between 1.3–4.5 mm day^{-1} for land which is salinized due to high watertables, although it is stated that it is often not feasible to install drainage to cope with the higher discharge rates. KERZUM (1958) found discharge rates of 1.7–2.2 mm day^{-1} satisfactory for reclamation and maintenance of a favourable salt balance in the Vakhsha Valley. The figures quoted from the Russian authors are somewhat higher than those derived from REEVE (1957), but their figures relate to the leaching of soluble salts brought in by irrigation water as well as salts rising from saline groundwater.

The procedure outlined above clearly demonstrates the relationship of drainage requirements with the variable factors of irrigation water quality, salt tolerance and water demand of crops in a fully efficient system of irrigation. Where irrigation is supplementary the additional factor of rainfall is easily taken into account. In practice, inefficiencies in the conveying system can locally increase the drainage rate and inefficiency in application often leads to drainage rates in excess of the leaching requirement.

3. Depth of watertable

a. It has long been recognized that an important factor affecting salinization is the depth at which saline groundwater occurs and nearly all control measures are aimed at lowering the watertable, often to a considerable depth. Early work seems mainly based on field observations, *e.g.* Burgess (1928) states that in Arizona reclamation of saline and alkali soil will not be permanent unless the watertable and associated capillary fringe is separated from the root zone, which in fine grained soils implies watertable control to 8–10 feet (2.40–3.00 m) below the surface. Teakle and Burvill (1945) remark that salt encroachment in Western Australia is generally serious where the groundwater is closer than 5–6 feet (150–180 cm) from the surface. Their observations apply to the wheat belt country, where cropping is under natural rainfall (15–25 inches, 380–630 mm) and the salt problem arose in valley alluvial soils due to excessive clearing of native upland vegetation. Kostiakov (as reviewed by Krimgold, 1945) considers the best watertable depth in saline parts of irrigation areas in the U.S.S.R. to be not less than 1.5–2.5 m; the depth within this range depending on the salt content of the groundwater, soil structure (especially relating to the topsoil, in regard to prevention of excessive evaporation) and cropping practices. For clay soils the watertable depth should not be less than 2–2.5 m, after the leaching period not less than 1.3–1.8 m. It should be allowed to remain at this level for only short periods.

Schoonhover *et al.* (1957) report that salting in the Nile delta is generally bad when the watertable is on the average 100 cm or closer to the surface, although excellent growth of crops is sometimes found on fine textured clay soils with the watertable between 50–60 cm. Groenewegen (1959) reports that salting on fine textured clay soils is not very widespread in the Murrumbidgee Irrigation Areas, even after 30 years of irrigation which has resulted in the occurrence of permanent saline groundwater at 100–150 cm from the surface. For non-irrigated areas, Jackson *et al.* (1956), found salinization to be insignificant in sand profiles if the watertable was below 100 cm. De Mooy (1959) concludes from rather limited data that the problem of salinization is probably overcome in the unirrigated black clay soils of the Murray area, by keeping the watertable from 60–90 cm below the surface all the year round.

b. *Definition and determination of critical depth*

The data presented in the previous section are based on observations of a general nature and are valid only for the conditions obtaining in the particular location investigated. In the U.S.S.R. attempts have been made to generalize such observations by defining and determining in various ways a so called "critical" depth of saline groundwater and relating desirable depth of drainage to this depth. The concept "critical" depth originally put forward by Polynov (1930) is defined as "that maximum height above the watertable, to which the salts contained in the groundwater can rise under natural conditions both by capillary rise and diffusion". It was soon recognized that the critical level depended on a variety of factors, *e.g.* the salt concentration of the groundwater, the velocity of capillary rise, diffusion phenomena, leaching action of rainfall and irrigation and on changes taking place in the capillary fringe during evaporation. FILOSOFOV (1948) argued that, in addition to all the above factors, the groundwater level itself fluctuates during the seasons and that it might be better to replace the term by another concept, *i.e.* the "critical economy" of a soil, a condition in which incoming salt is slightly higher than the removal. However, in recent literature the original concept still dominates (KOVDA, 1961, AVERIANOV, 1957).

The methods employed in determining the critical depth may for convenience be subdivided as follows: (i) those based on field and lysimeter experiments, (ii) those derived from graphs showing the relationship of groundwater depth and salt concentration in the upper soil layers and (iii) from graphs showing the relation between groundwater depth and evapotranspiration.

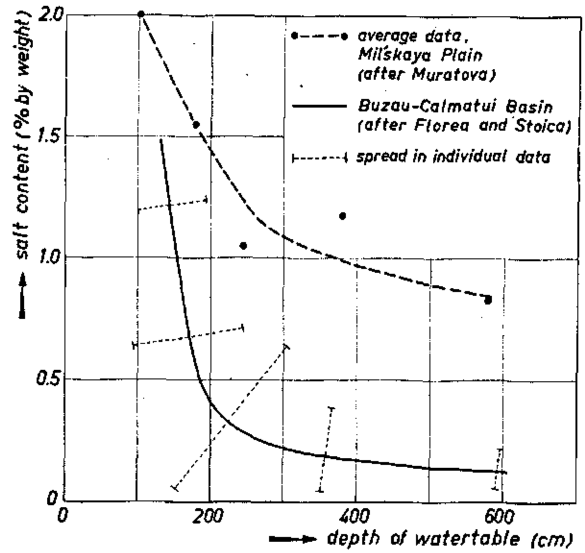
Sub (i) – Field experiments of SUKHACHEV (1958) in the Sokh alluvial fan, Uzbekistan showed that the critical depth, which he investigated in relation to cropping practice, soil profile and salinity of groundwater varied from 110–280 cm, the shallow depth relating to lucerne grown on stratified soils with a slow capillary rise and the greater depth to cotton grown on uniform profiles, where capillary rise was more pronounced. Intermediate values were found for lucerne on uniform soil (200–240 cm), cotton on stratified soil (150–180 cm) and for orchards (120–240 cm). LEGOSTAEV (1958) found in this manner depths in excess of 200 cm for lucerne and cotton on the Golodnaya Steppe, Uzbekistan, where soils are silty clay loams (sierozems and meadow soils) with high capillarity and fair permeability (28 cm day^{-1}). The field experiments of VOLOBUEV (1946) in the Mugan region, Kura–Arax basin, indicated depths of 120–150 cm as being critical.

KABAEV (1958) reports that in lysimeters in a cotton field, the salt concentration in the soil profile increased strongly where the watertable was kept at 1 m, less so at 2 m and none with the watertable at 3 m. He concludes that the salinization process in the Bukhara region will be stopped if the watertable is kept at 2.5–3 m.

Sub (ii) – An analysis of data presented by MURATOVA (1958) and FLOREA and STOICA (1958) yields relationships of watertable depth and average salt content of the upper soil layers of the type presented in figure 1. The data obtained by MURATOVA relate to the Milskaya plain in the lowlands of the Kura and Arax rivers, Uzbekistan; her estimate that the critical depth of groundwater is at 180 cm, however, would appear to be somewhat low, since it is seen that salt accumulation rises rather rapidly when the watertable is closer than 260 cm from the surface. FLOREA and STOICA, on the basis of analyses of 37 profiles from the plain between the Bazau and Calmatui rivers, Rumania, give a more correct interpretation of their data; they estimate the critical level to occur between 250–300 cm from the soil surface. In both cases there is a considerable scatter in the original data, these are explained by differences in soil type, drainage position, irrigation history and salt content of the groundwater. Care should therefore be exercised in utilizing such average data for other than broad scale drainage design.

Sub(iii) – Figure 2 shows the evaporation from groundwater as a function

FIG. 1. Dependence of salt content in soil on depth of watertable.



of watertable depth for various soils. Curves 2, 4 and 5 are data presented by KOVDA (1961). The experiments of KABAEV (1958) for the Bukhara region and LEGOSTAEV (1958) for the Golodnaya steppe (see *sub (i)*), together with curves 4 and 5 show that the critical depth of groundwater corresponds approximately to a possible evaporation rate of 0.1 cm day^{-1}. Applying this criterion to curves 1, 2 and 3 it is seen that the critical depth of groundwater for the Baraba region is about 140 cm, for the basin clay in Holland 100 cm and for Pachappa fine

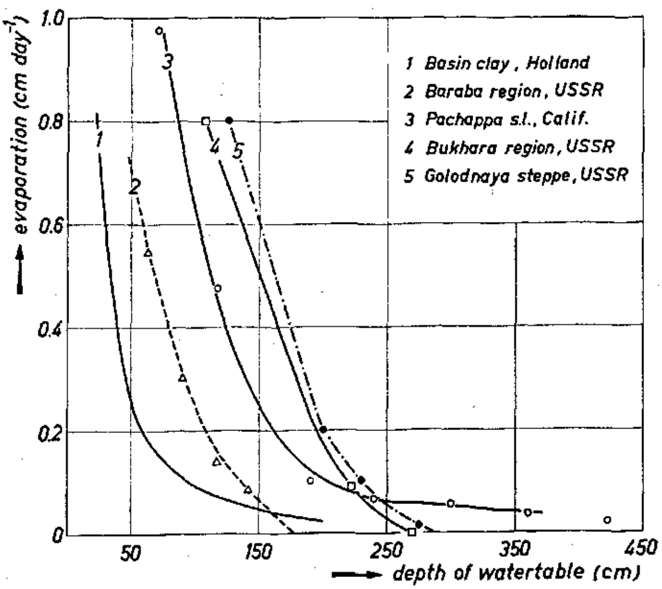


FIG. 2. Dependence of evaporation rate from groundwater on depth of watertable.

sandy loam about 200 cm. The latter figure agrees quite well with that of MARSHALL (1959) and GARDNER and FIREMAN (1958), who considered a depth of 180 cm for Pachappa to be desirable to control salinization. Curve 1 is derived from data presented by WIND (1955) concerning capillary transport in a fine textured river basin clay.

In the previous sections it has become apparent that the critical depth of watertable depends on a number of factors (POLYNOV, 1930, SUKHACHEV, 1958 and KOVDA, 1961), *e.g.* soil physical properties, groundwater salinity, climatic factors (evaporation and rainfall) and vegetation. The latter may influence the critical level in two ways, firstly in relation to moisture withdrawal from the profile by evapotranspiration and secondly because of differences in salt tolerance of various species. Some published data allow an assessment of the influence of these factors, either separately or jointly. These will be discussed in the remaining sections of this chapter.

c. *Soil properties affecting the critical depth of watertable*

JACKSON *et al.* (1956) studied seasonal changes in salinity in the south-east of South Australia, on soil profiles of predominantly sand and clay loam to light clay texture. During the period of maximum rate of salinization, here occuring in spring and early summer (period Sept.–Jan.) their sites 1B and 2 were comparable with regard to vegetation and watertable depth, but groundwater salinity differed. Recalculating their data for equal groundwater salinity shows considerable more salt accumulation in the clay loam profile at site 2 (see table 1). This table also shows a similar result in comparing sites 1C and 3, where the light clay profile (site 3) accumulated about twice as much salt as the sand profile, from a watertable at approximately 100 cm depth.

TABLE 1. Rate of salt accumulation during the period Sept. 1953–Jan. 1954, in different soil types. Calculated from data by JACKSON *et al.* 1956).

Site	Soil Profile	Vegetation	Average watertable (cm)	Depth interval (cm)	Increase salt content (%)
1B	sand	sparse native	70	0–50	100
2	clay loam	sparse native	60	0–50	650
1C	sand	pasture	105	0–100	300
3	light clay	sparse native	103	0–100	590

The greater salt increase in profile 1C, compared to profile 1B (these profiles are nearly identical in mechanical composition) appears anomalous, since the watertable was deeper at site 1C. However, this difference is readily explained by the fact that site 1C supported an improved pasture vegetation including the deep rooting lucerne, which increases both moisture withdrawal by transpiration and capillary transport in the deeper soil horizons. This agrees with data presented by VERHOEVEN (1953) who reported that salt transport through capillary rise of moisture was generally higher on cropped plots than on bare or fallow sites. JACKSON *et al.* (1956) concluded from these data that the critical depth of watertable for the sand was between 100–120 cm, but deeper for the clay loam and light clay profiles.

The work of GROENEWEGEN (1959) in the Murrumbidgee Irrigation Areas has shown that salt accumulation was related to soil permeability, k. His data, together with some of the author's own measurements made at a later date under similar circumstances are shown in figure 3. In order to compare profiles with differing chloride distribution and concentrations use was made of a single value, the "mean depth of chloride" $z = \Sigma\ (x\ d\ C_d)/\Sigma\ (d\ C_d)$, where $x =$ the mean distance from the soil surface of a layer of thickness d and chloride concentration C_d. It is seen that there is a rather significant relation between k and z; in

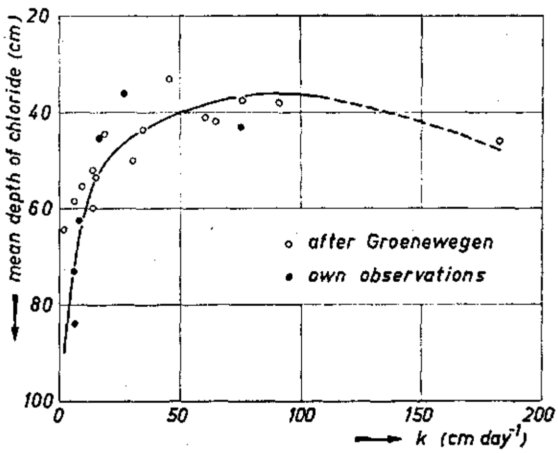


FIG. 3. Dependence of mean depth of chloride (in layer 0–110 cm) on hydraulic conductivity.

slowly permeable soils, which are invariably rather poorly structured clay profiles (TALSMA and FLINT, 1958) there is little accumulation in the surface layers. Passing to the more permeable soils, which vary in texture from light clay to loam, chloride accumulation increases with increasing hydraulic conductivity. There is an indication that chloride accumulation decreases again in soils of high permeability; in the Murrumbidgee Irrigation Areas soils in this permeability class are light clays of a "gravelly" structure (*e.g.* GROENEWEGEN, 1959), characterized by a low percentage of dispersable clay (TALSMA & FLINT, 1958). The relationship suggested by GROENEWEGEN is indicated by the curve shown in figure 3.

The distribution of salts in soils of fine and intermediate texture in the Zevulun Valley (RAVIKOVITCH and BIDNER-BAR-HAVA, 1948) is very similar to that observed in the Murrumbidgee Irrigation Areas. In addition the salt distribution in sand profiles was found by these authors to be similar to the distribution in the light clay profiles of gravelly structure of the Murrumbidgee Irrigation Areas. VOLOBUEV'S (1946) experiments in the Mugan region showed that saline groundwater could be tolerated at shallower depth in clay soils (120–130 cm) than in loamy soils (130–150 cm).

It would appear, from the data discussed in this section, that soils of intermediate texture are most liable to salt accumulation and need watertable control to greater depth than either clay soils of low permeability or sands and other materials of high permeability.

d. *Groundwater salinity*

It is obvious that the higher the concentration of soluble salts in the groundwater, the greater is the danger of salinization. Generally speaking, doubling the concentration under a given set of conditions, will double the danger of salinization. At low concentrations this danger is not very high, *e.g.* KOVDA (1961) states that the groundwater depth is not a very important factor when the concentration of soluble salts of the chloride-sulphate type is around 0.2–0.3%, or 0.07–0.1% for groundwater containing alkali (HCO_3^-); when concentrations of 1.0–1.5% are reached, the critical depth of the watertable, as a general rule will be around 200–250 cm.

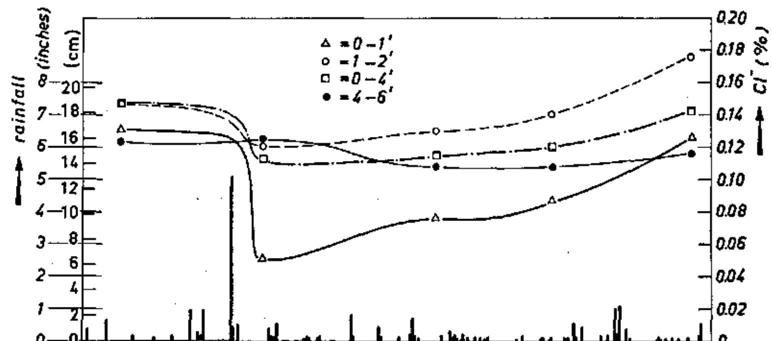
There are often regular patterns in the distribution of both soil and groundwater salinity, associated with the genesis and hydrology of the land surface to be irrigated. The general rule is that slowly permeable clay soils are more saline than soils of intermediate and sand texture (VAN DIJK and TALSMA, in press). This also applies to groundwater, initially occurring in the deeper strata, where coarse textured aquifers often contain water with salt concentrations approaching that of adjacent surface rivers and where brackish water is often found in areas where no aquifers are present (LEGOSTAEV, 1958, PELS, in press). Examples of this condition in non irrigated areas are the Zevulun Valley (RAVIKOVITCH and BIDNER-BAR-HAVA, 1948) where salt concentrations of 3.7–5.1% are found in the area of clay soils and 0.04–0.8% in the area of transition and sandy soils, and the Coonalpyn Downs (JACKSON *et al.*, 1956) where these percentages were 2.6 and 1.9% respectively. In irrigated areas, where redistribution of salt may be expected to take place due to developing watertables, the same regularity apparently persists for long periods. For example, in the Murrumbidgee Irrigation Areas after about 35 years of irrigation it was found, from analyses of some 1500 groundwater samples, that in slowly permeable clay soils (k = 0–9 cm day^{-1}) 44% of the samples had an electrical conductivity exceeding 5 millimhos cm^{-1} ($\pm$ 0.3% total soluble salts); for intermediate soils (k = 9–30 cm day^{-1}) this figure was 34.5% and for permeable soils ($k > 30$ cm day^{-1}), 26%.

e. *Climatic conditions*

The annual amount and distribution of rainfall and evapotranspiration are the main climatic factors causing movement of salt in the soil profile. In periods when rainfall exceeds actual evapotranspiration there is a downward movement of water and soluble salts. When actual evapotranspiration exceeds rainfall there is an upward movement. For irrigated fields within an irrigation area, the amount of water applied has to be added to the rainfall. Diurnal and seasonal temperature variations induce a temperature gradient in the soil profile and may cause some salt transport to the warmer top layers (DURAND, 1956, GURR *et al.*, 1952) although this process is not very likely to be of great importance in the control of salinity in irrigated agriculture, *e.g.* DE VRIES (1958) concludes that temperature gradients usually have little effect, except at low evaporation rates ($<$ 0.01 cm day^{-1}). A few examples, showing the effect of climatic type on salt accumulation, are cited below.

In regions with good winter rain and dry hot summers (Mediterranean climates) the salinity in the upper soil layers is often at its lowest by the end of the winter period. Examples are given by JACKSON *et al.* (1956) and DE MOOY (1959) for the south-east of South Australia and by RAVIKOVITCH and BIDNER-BAR-HAVA (1948). Associated with the high winter rainfall is often a high watertable and wet soil profile in early spring. During spring and the early summer capillary movement of water concentrates the salts in the surface horizons and lowers the watertable. This process is often markedly retarded some time during the summer due to excessive drying of the top soil (*e.g.* JACKSON *et al.*, 1956).

In the Murrumbidgee Irrigation Areas the winter rainfall is often not sufficient to remove soluble salts from the root zone. By way of example, the chloride content of various soil layers, together with rainfall distribution and Australian standard pan evaporation data for the period Dec. 1958–Dec. 1959, are given in figure 4. These data were obtained by the author on a bare site. The soil is a permeable loam ($k = 75$ cm day^{-1}) and the watertable was kept 150–180 cm below the soil surface by means of tile drains at 180 cm. It is seen that the only period of desalinization was after the extremely heavy rainfall in March, 1959. From this period onwards a gradual resalinization has occurred.



		dec'58	jan	feb	march	april	may	june	july	aug	sept	oct	nov	dec59	
rain		1.41	0.34	2.36	7.01	0.75	0.84	1.32	1.03	0.51	0.59	3.06	0.49	0.73	inches
		3.58	0.86	5.99	17.81	1.91	2.13	3.35	2.62	1.30	1.50	7.77	1.24	1.85	cm
evaporation		8.37	10.14	8.03	4.95	4.01	2.69	1.45	1.77	3.09	3.60	5.05	8.62	8.10	inches
		21.26	25.76	20.40	12.57	10.19	6.83	3.68	4.50	7.85	9.14	12.83	21.89	20.57	cm

FIG. 4. Chloride content of soil layers in relation to rainfall and evaporation.

The examples given above show in a general way the influence of climatic factors on the process of salinization. A thorough theoretically founded study of the influence of climatic factors on the degree of salinization has not yet been made. Relations of the type presented in figure 2 represent essentially maximum flux rates through the profile at given watertable depth. These rates can only be maintained if the actual evapotranspiration rate has the same magnitude. At both higher and lower evapotranspiration rates the flux is less (*e.g.* REMSON and FOX, 1955, GARDNER and FIREMAN, 1958). The latter authors studied this problem in an indirect way in laboratory experiments. The conclusion drawn in section I-3-b (Sub iii), that the critical depth of watertable corresponds to a maximum evaporation rate from groundwater of 0.1 cm day^{-1}, must, due to the lack of a reliable analysis of field data, be considered as an approximation only. KOVDA's (1961) generalization, that the relation of evaporation intensity to critical depth of watertable may be approximated by $y = 170 + 8t° \pm 15$, where y = critical depth in cm and t = average annual temperature in °C, is in conflict with the data presented in figure 2.

f. *Conclusions*

Deeper drainage is generally recommended for salinity control than for waterlogging and disposal of excess water only. It has become clear that for the control of the salinity factor, the desirable depth of watertable depends on quite a number of factors, *e.g.* soil physical properties, groundwater salinity, evapotranspiration, rainfall and type of vegetation. General agreement appears to exist that saline groundwater may be tolerated at shallower depth in soils of coarse texture than in soils of intermediate texture, for which depths of 180 cm or more are reported. The groundwater in fine textured soils is generally more saline and it is often recommended to control the watertable in these soils to still greater depths. Analysis of some of the more recent literature on salt movement in fine textured profiles however, does not support this conclusion. This question will be considered in detail later.

There appears to be considerable scope for a more accurate analysis, especially of field data, of the process of salinization; particularly regarding the physical processes involved in the transfer in different soil types and under the influence of actual evapotranspiration rates, of moisture and salt from shallow watertables. An analysis of this kind will be given in the following sections, for conditions existing in the Murrumbidgee Irrigation Areas, Australia.

II. THE MURRUMBIDGEE IRRIGATION AREAS

1. General introduction

a. *Physiography, geology*

The Murrumbidgee Irrigation Areas lie between latitudes 34°–35°S and longitudes 145°–147°E, on the north-eastern fringe of a large inland riverine plain which forms part of the Murray basin of south-east Australia. In physiographic terms they are situated on the northern flank of an extensive alluvial fan of low relief, which has been formed where the present Murrumbidgee river leaves the eastern hill country. To the north the areas are fringed by a series of low ranges. The land surface slopes generally towards the west with an average gradient of 2 ft per mile (38 cm km^{-1}) and natural surface drainage is generally poor.

Geological investigations (Pels, 1960) indicate a maximum thickness of unconsolidated sediments of 465 feet (142 m), at lower depth consisting of nearly horizontal, lacustrine deposits (soft brown coal, quartz sand and kaolinite deposits). The fluviatile deposits of the alluvial fan have a thickness of 160 feet (49 m) in the east to 60 feet (18 m) in the west. Sand and gravelly deposits are generally more extensive and numerous in the east, in the western areas clay deposits are predominant. Since the onset of irrigation large parts of the sand deposits have become aquifers, containing water under sub-artesian pressure, thereby contributing to the internal drainage problem. Although the surface deposits of the Murrumbidgee Irrigation Areas are considered to be partly of aeolean origin the geographical distribution of salt, occurring in the upper deposits conforms generally to the original fluviatile pattern of deposition from the east. In a survey of the salinity status of virgin soils (data partly published by van Dijk and Talsma, in press) the author found the average NaCl content of the 60–75 cm layer in soils of the red brown earth group to be 0.14, 0.18 and 0.24% at distances of 23, 30 and 40 miles (37, 48 and 64 km) from the apex of the alluvial fan.

Water for irrigation is stored on the upper reaches of the Murrumbidgee river. It is of high quality, containing about 0.008 % soluble salts. The average composition in m.e.l^{-1} is Ca^{++}: 0.45, Mg^{++}: 0.45, Na^{+}: 0.30, Cl^{-}: 0.28, SO_4^{--}: 0.24 and HCO_3^{-}: 0.65. Salinity problems that have developed are therefore mainly caused by redistribution of naturally occurring salt, due to rising watertables (GROENEWEGEN, 1957, 1959). It is currently estimated that some 5000 acres, mainly comprising the more permeable soils in the horticultural districts, are to some degree salt affected. Although no serious salinization has as yet occurred in soils of low permeability, the problem of rising groundwater in the eastern parts of the areas is causing concern.

b. *Soils, original salinity*

The soils of the Murrumbidgee Irrigation Areas are classed in three Great Soil Groups, the red brown earths, the grey and brown soils of heavy texture and the solonized brown (or mallee) soils. Soils in the latter Group occur mainly to the north of the irrigation areas, the natural salinity of these soils is somewhat higher than that of the soils in the other Groups. Because of their limited use for irrigation along the Murrumbidgee, these soils are not included in the present study.

Generally, the more permeable soil types (hydraulic conductivity, k_{sat} = 15–300 cm day^{-1} of the red brown earth Group are used for horticulture. The finer textured soils of this Group (k_{sat} = 5–30 cm day^{-1}), together with the grey and brown soils of heavy texture (k_{sat} = 5–30 cm day^{-1}) are used for large scale farming, with the main rotation of rice and pasture. Prior to irrigation there was no watertable and none of these soils was saline in the surface layers (VAN DIJK, 1961).

The maximum concentration of soluble salts in fine textured, virgin red brown earths occurs at about 100 cm depth in the profile, where the average concentration of total soluble salts is between 0.35–0.70 % and of NaCl between 0.08–0.24 %. Below this depth the soluble salt content decreases to around 0.15–0.35 % the NaCl content to 0.05–0.18 %. The grey and brown soils of heavy texture are generally non-saline to a greater depth, the maximum concentrations of soluble salt and NaCl are similar to those of the red brown earths, but these concentrations occur at about the maximum depth normally investigated (180 cm) or deeper. Figure 5 shows the variation in salt content with depth in some virgin soil profiles, the top part of the figure shows the distribution of total soluble salts and NaCl in red brown earth and transitional soil types, the lower part gives the distribution pattern in grey and brown soils of heavy texture.

2. CLIMATIC FACTORS

a. *General description of the climate and available records*

The Murrumbidgee Irrigation Areas are situated in the broad semi-arid area between two main climatic zones: the Mediterranean type characterized by winter rainfall to the south and south-west and the Monsoonal type, with the main incidence of rainfall in the summer period, to the north and north-east. Generally the rainfall in winter is of low intensity and longer duration than the summer rains, which chiefly occur during infrequent thunderstorms.

The main meteorological recording station is at present maintained by the Irrigation Research Station near Griffith, which is at Lat. 34° 17'S, Long.

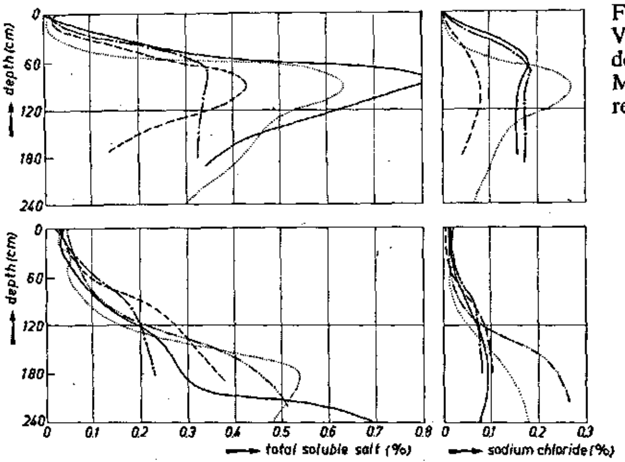


FIG. 5.
Variation of salt content with depth in virgin soils of the Murrumbidgee Irrigation Areas. After VAN DIJK (1961).

146°03′E and 128 m above sea-level. Continuous records of temperature and rainfall are available since 1924 and evaporation records from an Australian standard pan evaporimeter since 1932. This evaporimeter is of the sunken pan type, set in a well maintained lawn. It consists of a cylindrical copper tank, 3 feet in diameter and 3 feet deep, set in an outer concrete tank 4 feet in diameter and 2 ft. 10 ins. deep. The outer tank is set flush with the ground. The depth of water in the inner tank is allowed to vary from $1\frac{1}{2}$–3 inches below the rim.

Average monthly rainfall and evaporation, together with the average daily maximum and mean temperatures are given in figure 6. The annual rainfall is

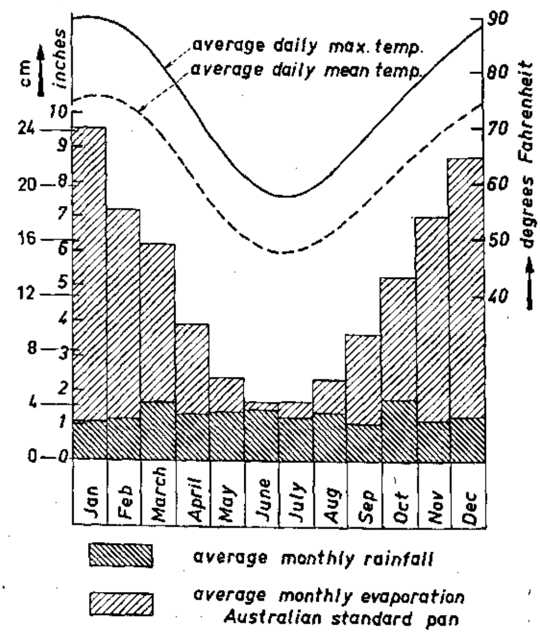


FIG. 6.
Climatic data for Griffith, N.S.W.

15.7 inches (400 mm), pan evaporation 59.8 inches (1520 mm). It is seen that the monthly averages of rainfall are evenly distributed throughout the year with slight peaks at the equinoxes. On the average, pan evaporation exceeds rainfall over the whole of the year. The mean monthly temperature ranges from 78 °F in January to 47.3 °F in July, the mean monthly maximum from 89.6 °F to 57.3 °F.

Because of the large interest in evaporation data in irrigated agriculture, records from other types of evaporimeter are available for various periods. Among these are a sunken pan of large diameter, a black-pan, Piche and U.S. Class A land pan evaporimeters. A description and thorough review of these records has recently been given by FLEMING (1962). The same author also examined the meteorological records available at Griffith to investigate relationships developed by PENMAN (1948) in England, KOHLER *et al.* (1955) in the U.S.A. and HOUNAM (1958) in Australia, for use in the Penman formula.

Since rainfall and actual evapotranspiration are the main climatic factors causing moisture and salt transport in the soil it is imperative to obtain reliable estimates of these quantities from the available records. Whilst rainfall data are relatively easy to interpret, provided there is no surface runoff, there are considerable difficulties in obtaining actual evapotranspiration rates from evaporimeter records. The former must be deduced from the latter through a series of relationships given by:

$$E_0 = c\, E_p \qquad (6)$$

$$E_t = f\, E_0 \qquad (7)$$

$$E_a = p\, E_t \qquad (8)$$

where E_0 = evaporation from a free water surface, E_p = pan evaporation, E_t = potential evapotranspiration, E_a = actual evapotranspiration and c, f and p are appropriate coefficients. None of these coefficients may generally be regarded as constant, *e.g.* the pan coefficient c has been shown to vary with the type of pan, condition of its surroundings and time of the year. The factor f appears to depend on the type of vegetation and also to be different as the crop grows in a cool, humid climate or under hot and arid conditions. Finally, it is now becoming increasingly clear that p depends not only on the amount of available water in the soil but also on the magnitude of E_t.

The pan coefficient c, may be elimated by using *e.g.* the Penman formula to estimate E_0 directly, provided suitable meteorological records are available for use in this formula.

An estimate of the magnitude of potential evaporation, E_0, for the Murrumbidgee Irrigation Areas will be made in the following section, by examining both HOUNAM's (1958) work on pan coefficients and FLEMING's (1962) investigation of the Penman formula. Subsequently the relations expressed by equations (7) and (8) will be investigated with the aid of the author's own measurements on a pasture plot and an examination of data presented by WEST and PERKMAN (1953) on evapotranspiration rates observed in a citrus orchard near Griffith.

b. *Potential evaporation,* E_0

FLEMING's calculations with the Penman formula have been made using the 1961 meteorological records. His data are particularly useful in the context of this paper, since the author's own observations on potential and actual evapotranspiration also extend for the greater part over this period. HOUNAM's (1958) paper on pan coefficients is of an earlier date and is best discussed with reference to the more recent work. Before examining FLEMING's results, his main conclusions are briefly summarized.

For the calculation of shortwave radiation R_s, the relations of PENMAN (1948): $R_s = R_a\,(0.18 + 0.55\,n/N)$, where R_a = radiation outside the atmosphere, n/N the ratio of actual to possible hours of sunshine, and the relation derived from two years records of mainly coastal stations in Australia by HOUNAM: $R_s = R_0\,(0.34 + 0.66\,n/N)$, where R_0 = clear day radiation on a horizontal surface, were compared with results from a simple shortwave radiation integrator. Neither of the above relations were found to be satisfactory for Griffith.

One of the errors made in applying these formulas is due to a bias in sunshine hour readings at Griffith, *e.g.* on a sample number of "clear" days the integrated records of a Fuess Actinograph showed a mean sunshine hour ratio of 0.87. The records of the radiation integrator were therefore used to estimate mean monthly solar radiation.

No data were available at Griffith to check the accuracy of the various equations used to estimate net longwave radiation R_l, which was therefore calculated from the Brunt equation as originally used by PENMAN:

$$R_l = \sigma T_a^4\,(0.56 - 0.09\sqrt{e_o})\,(0.10 + 0.90\,n/N) \tag{9}$$

where: n/N is as above,
σT_a^4 = the theoretical black body radiation at T_a°K.,
e_o = saturation vapour pressure at dew point in mm Hg.

The aerodynamic equation, used by PENMAN in the form:

$$E' = (0.35 + 0.0035\,u)\,(e_a - e_o) \tag{10}$$

where: E' = evaporation in mm water per day,
e_a = saturated vapour pressure in mm Hg, at mean air temperature,
u = miles run of wind per day at 2 m height,

was checked for Griffith with the aid of records available from an American Class A pan. The following effective approximation was obtained:

$$E'' = (0.35 + 0.0055\,U)\,(e_s - e_o^1) \tag{11}$$

where: E'' = pan evaporation (mm day^{-1}),
U = windspeed on the pan anemometer, at 0.5 m height (miles day^{-1}),
e_s = saturation vapour pressure determined from temperature means (mm Hg),
e^1_o = mean air vapour pressure, estimated from records (mm Hg).

Equation (11) was compared with the two forms of aerodynamic equation quoted by KOHLER *et al.* (1955) and found to be the same at Griffith for the mean monthly windspeed of 45 miles per day. It can also be seen that the relation is very similar to the value of the constants in equation (10) when the greater height is taken into account. FLEMING however, concluded that the value of the windspeed modifying coefficient is dependent on the dimensions of the evaporating surface and should decrease with increasing area.

In the application of the Penman formula to average monthly data of 1961 use was made of the 9 a.m. dew point data to obtain e_o, despite the fact that this procedure gives results which are sometimes too high. This is due to the fact that the standard method of observation is an unaspirated wet and dry bulb thermometer, which gives deviating results on days of light wind at 9 a.m. The reason for using these data is that they are frequently the only ones available.

The end result is that, apart from the uncertainty about the constants used in the net longwave radiation term, acknowledged errors occur both in estimating R_l and E'. However the effect of these errors on E_0 is not large as will be shown below.

The error in the longwave radiation term is due to errors in both e_o and n/N, the error in E' depends only on errors in e_o. Putting eq. (9) in logarithmic form, differentiating and taking absolute values gives the following expression for the relative error in R_l:

$$\frac{\Delta R_l}{R_l} = \frac{0.045}{0.56\sqrt{e_o} - 0.09\, e_o} \cdot \Delta e_o + \frac{0.90\, \Delta\, (n/N)}{0.10 + 0.90\, (n/N)} \quad (12)$$

Similarly for the relative error in E':

$$\frac{\Delta E'}{E'} = \frac{1}{e_a - e_o} \cdot \Delta e_o \quad (13)$$

From detailed examination of records it appears that errors in e_o are of the order of 1 mm Hg throughout the year, *i.e.* 8–10% in summer and 12–15% in winter, the maximum error in the sunshine ratio n/N seldom exceeds 10%. For periods of little wind at 9 a.m. and relatively high sunshine ratios these errors have their maximum effect and, when substituted in equations (12) and (13) give an error in R_l of about 14% and an error in E' varying from about 7% in summer to 25% in winter. Calculated values of both R_l and E' are low in this case, resulting in an overestimate of the net radiation term and an aerodynamic solution that is too low. In calculating E_0 with the Penman formula these errors are therefore compensating and the error in E_0 would hardly exceed 5% during periods in which the above conditions apply. It would appear therefore that over monthly periods the effect of these errors is not noticeable.

FLEMING's monthly values of E_0 are therefore considered to be reliable and his conclusion, that the Australian standard pan closely estimates E_0 permits use to be made of standard pan evaporation data for periods for which calculation of E_0 with the Penman formula have not been made. The elements of the Penman formula as calculated for Griffith for 1961 are set out in table 2, together with Australian standard pan evaporation data for 1961 and average pan evaporation data over the period 1932–1957.

Although it is, in principle, not necessary now to consider pan coefficients, it is informative to compare HOUNAM's (1958) work on pan coefficients for the Australian standard pan with the results discussed above. Various estimates of monthly pan coefficients are set out in table 3. The ratios E_0^*/E_{Aust} and E_o^{**}/E_{Aust} for 1961 have been calculated from data of FLEMING, the only difference being the method of calculating shortwave radiation, which was from direct measurement with a photocell meter in the case of E_0^* and calculated from HOUNAM's relation $R_s = R_0\,(0.34 + 0.66\, n/N)$, using values of R_0 calculated from Fuess Actinograph records, in the case of E_0^{**}. Using HOUNAM's relation results in lower pan coefficients, which tend to exhibit a minimum around June.

HOUNAM also observed this minimum in calculating average monthly pan coefficients E_0^+/E_{Aust}, where E_0^+ was obtained from the Penman formula using $R_s = R_0\,(0.34 + 0.66\, n/N)$, R_0 being estimated from the average annual vapour pressure and atmospheric pollution. This was in agreement with his values for Melbourne, where the above relation and the estimate of R_0 were more reliable, and with data concerning the relation of standard pan to large pan (12 ft diam.)

TABLE 2. Penman formula, 1961, Griffith, and Australian standard pan evaporation data (from FLEMING, 1962).

	Net solar radiation $0.95\ R_s$	Net longwave radiation R_l	Net radiation H_0	Δ*	Aerodynamic solution E'	Evaporation 1961 E_0	Evaporation 1961 E_{Aust}	Average E_{Aust}
	mm day^{-1}	mm day^{-1}	mm day^{-1}	mm Hg (F°)$^{-1}$	mm day^{-1}	mm day^{-1}	mm day^{-1}	mm day^{-1}
Jan.	10.2	3.1	7.1	0.78	8.8	7.6	7.9	7.8
Feb.	9.4	3.0	6.4	0.74	8.0	6.8	7.8	6.5
March	7.9	2.9	5.0	0.62	5.0	5.0	5.4	5.1
April	5.8	2.1	3.7	0.52	2.2	3.2	3.1	3.3
May	5.2	3.1	2.1	0.38	1.7	1.9	2.4	2.0
June	3.9	1.9	2.0	0.34	0.8	1.5	1.5	1.4
July	4.2	1.9	2.3	0.32	1.0	1.7	1.5	1.4
Aug.	5.3	2.4	2.9	0.34	0.9	2.0	1.8	1.9
Sep.	7.4	3.1	4.3	0.40	2.1	3.4	3.1	3.1
Oct.	9.0	3.3	5.7	0.55	4.4	5.2	5.2	4.3
Nov.	8.7	2.6	6.1	0.60	4.8	5.7	5.2	6.0
Dec.	9.4	2.8	6.6	0.68	6.9	6.7	6.8	7.2

* Δ = slope of the saturated vapour pressure-temperature curve, at air temp.

TABLE 3. Estimated pan coefficients for the Australian standard pan at Griffith.

Ratios	Jan.	Feb.	Mar.	Apr.	May	Jn.	Jl.	Aug.	Sep.	Oct.	Nov.	Dec.	Year
E_0*/E_{Aust}, 1961	0.96	0.87	0.92	1.03	0.83	1.00	1.17	1.14	1.08	0.98	1.04	0.99	0.97
E_0**/E_{Aust}, 1961	0.88	0.82	0.81	0.89	0.67	0.66	0.68	0.94	1.02	0.92	0.96	0.96	0.87
E_0^+/E_{Aust}, average	0.80	0.95	0.87	0.78	0.70	0.54	0.71	0.88	0.91	1.02	0.98	0.86	0.87
E^{12}/E_{Aust}, 1955	–	–	0.99	0.84	0.87	0.68	0.88	0.81	0.80	0.78	0.95	0.92	–

E_{Aust} = evaporation from Australian standard pan.
E_0* = evaporation calculated from Penman formula, with R_s calculated from photocell measurements.
E_0** = evaporation calculated from Penman formula, with $R_s = R_0\,(0.34 + 0.66\ n/N)$, R_0 obtained from Fuess-Actinograph records.
E_0^+ = evaporation calculated from Penman formula, with $R_s = R_0\,(0.34 + 0.66\ n/N)$, R_0 estimated from average annual vapour pressure and atmospheric pollution.
E^{12} = evaporation from a 12 feet diameter pan.

evaporation (see *e.g.* table 3). HOUNAM explains the low pan coefficients in the winter period by the fact that the difference between soil and pan water temperature is largest in this period, the soil being warmer. Extra energy is therefore available for evaporation from a sunken pan, resulting in relatively high evaporation rates. A comparison of evaporation figures from the American Class A pan (which is raised above the ground) and the standard Australian pan at Griffith supports this conclusion.

The high values of the pan coefficient (E_0^*/E_{Aust}) in the winter period as calculated from the data of FLEMING are probably due to an overestimate, at low levels of radiation, of solar radiation by the photocell meter (FLEMING, personal comm.). This trend would have little effect on the overall data and indeed would make the ratio E_0^*/E_{Aust}, and thus the pan coefficient, more constant. It is concluded that the low pan coefficients E_0^+/E_{Aust} for Griffith, as calculated by HOUNAM (1958) from his relationship $R_s = R_0\,(0.34 + 0.66\; n/N)$ are due to a large extent to an under estimate of shortwave radiation. The pan coefficients, E_0^*/E_{Aust}, calculated from the data of FLEMING (table 3) are very similar to those of HOUNAM (1958) for Melbourne.

c. *Potential and actual evapotranspiration, E_t and E_a*

The relationships $E_t = f\,E_0$ and $E_a = p\,E_t$ were investigated for the period Oct. 1960–Nov. 1961 on a field plot situated in a permanent pasture. During the winter period the growth on this plot was mainly Wimmera rye grass *(Lolium rigidum)* and sub-clover *(Trifolium subterraneum)* while during the summer, when the plot was not irrigated, the deeper rooting couch grass *(Cynodon dactylon)* dominated. The stand in both spring and autumn consisted of a mixture of these species.

Actual monthly evapotranspiration, E_a, of the pasture was computed from the waterbalance equation:

$$E_a = I + R + C - D \pm \Delta W \tag{14}$$

where: I = irrigation, R = rainfall, C = capillary rise, D = drainage and ΔW = the change in moisture storage of the profile. All the terms on the right hand side in eq. (14) were measured, except the quantity of irrigation water, I, which was the same in quantity and method of application as the surrounding pasture in order to ensure equal conditions in the environment of the plot. The only period of irrigation was during March, 1961, for which it was therefore not possible to solve eq. (14). The rainfall, R, was registered with a standard 8 inch rain gauge, ΔW was determined from periodic moisture samplings and C and D from changes in salt concentration in the root zone (see *e.g.* VERHOEVEN, 1950). To minimize errors due to variability of both moisture content and salt concentration, measurement of these quantities was in quadruplicate over each 10 cm depth interval. The salt concentrations in the root zone were generally too low (chloride from 1 gr l^{-} at 10 cm to 7 gr l^{-} soil moisture at 80 cm depth) to prevent or retard growth appreciably.

The results of this investigation are given in table 4, where the actual evapotranspiration is given as a fraction of the evaporation from an Australian standard pan. From equations (6), (7) and (8) it is seen that this fraction is equal to *p.f.c.* The composite coefficient *p.f.c.* is well correlated with average monthly pan evaporation ($r = -.84$) given in column 3 and percentage available water ($r = +.90$) shown in columns 5 and 6. Available water, calculated as a

TABLE 4. Relationship of actual evapotranspiration (E_a) to evaporation from an Australian standard pan (E_{Aust}) for pasture.

Month	$\frac{E_a}{E_{Aust}}$	Average E_{Aust}	Average watertable	Available water %	
1960–'61	–	cm month⁻¹	cm	0–40 cm	40–80 cm
Oct.	0.62	13.4	108	32	93
Nov.	0.38	18.0	149	11	45
Dec.	0.22	22.3	160	17	29
Jan.	0.29	24.2	171	2	0
Feb.	0.09	18.3	194	0	0
Apr.	0.69	10.0	102	78	104
May	0.63	6.1	121	75	111
June-July	0.77	4.3	122	89	114
Aug.	1.04	6.0	78	113	111
Sep.	0.75	9.4	87	67	105
Oct.	0.77	13.4	120	4	66

percentage of the amount of water occurring in the soil between 15 atm. and 80 cm suction, is seen to exceed 100% in some cases, when a high watertable (col. 4) existed. Distinction is made between available water in the layer 0–40 cm and 40–80 cm, to allow changes in effective root zone to be taken into account. For the period May–August, when the couch grass vegetation was practically dormant, average moisture content in the 0–40 cm layer only, was used in calculations. Before attempting a separation into the component coefficients p, f and c, the results of an experiment by WEST and PERKMAN (1953) concerning the effect of available water on the transpiration rate of citrus will be analysed.

WEST and PERKMAN (1953) investigated the relation between moisture withdrawal in part of the root zone of citrus (4–12 in.) and cumulative evaporation from a black pan evaporimeter. Their results are expressed in curves of the type $W_s = W_{wp} + W_{av}\, e^{-b\, E_{bp}}$, where W_s = water content of the soil, W_{wp} = soil moisture content at wilting point, W_{av} = available water, E_{bp} = black pan evaporation and b = an experimentally determined constant. Three different relationships were presented for varying surface treatments (tilled, bare soil and sod). Since the amount of available water and moisture contents at wilting point and field capacity respectively are rather poorly estimated by extrapolating these curves to both sides, these quantities will be calculated with the aid of additional data of GREACEN and PERKMAN (1953) and pF curves obtained by the author on similar soils.

Table 5 shows some physical properties of soils similar to the "citrus soil" of WEST and PERKMAN. This "citrus soil" is in fact composed of various soil types, in the experimental orchard concerned the main types are Banna sand and Hanwood loam. The main difference between the "citrus soil" and the soil types given in table 5 is in the aeration porosity, resulting in a lower field capacity for the citrus soil. The various estimates of field capacity are 13.9% from the data of GREACEN and PERKMAN (1953), 14.5% from the data of WEST and PERKMAN (1953) and 16.3% by averaging the 3 soil types in table 5. Wilting point estimates are 5.2% from the component soil types as well as from the data of WEST and PERKMAN. Taking wilting point at 5.2%, field capacity at 14.5% and average data from the 3 component pF curves at intermediate

TABLE 5. Some physical properties of soils with moisture characteristics shown in figure 7.

Soil type	Depth	Clay ($< 2\mu$)	Bulk density	Total porosity	Aeration porosity	Moisture	
						F.C	W.P.
	cm	%		%	%	% weight	% weight
Banna sand	30	13.3	1.53	42.2	12.8	15.0	2.2
Jond. loam	8	18.0	1.60	35.2	8.2	16.6	5.9
Hanwood loam	15	21.2	1.53	38.9	8.8	17.4	7.5
Citrus soil	15*	17.8*	1.65*	38.0*	15.0*	14.5+	5.2+

* From data of GREACEN and PERKMAN (1953).
+ From data of WEST and PERKMAN (1953).

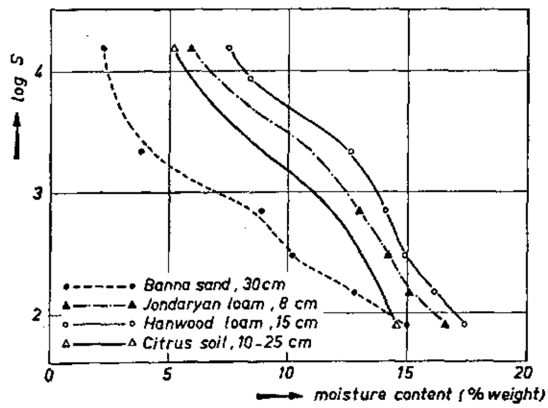


FIG. 7.
Field moisture contents versus suction.

moisture contents, gives a moisture characteristic for the citrus soil as shown in figure 7.

From the relationships given by WEST and PERKMAN it can now be concluded that field capacity was reached immediately after irrigation on the tilled and bare soil treatments and wilting point was approached only at the sod treatment. Relative evaporation rates may now be deduced from the above relationships, these are given in table 6. The relative rates for the sod treatment were calculated, assuming a cumulative evaporation of 1.2 in. (equal to the tilled treatment) from the black pan, as available moisture decreases from 100 to 75%. Average relative evapotranspiration rates, calculated from these data, are shown in table 7, in relation to moisture suction deduced from the composite pF curve of figure 7.

For a comparison of these data with the pasture plot (table 4) it is necessary to replace the black pan evaporation data with those of the Australian standard pan and to calculate moisture withdrawal in the whole soil profile from the losses in the main root zone of 4–12 in. Comparing records it is found that the average ratio of black pan to Australian pan evaporation for the months Jan., Feb. and March 1951 was about 1.37, slightly lower than the nominal value of 1.4. The average moisture loss on the tilled and bare soil treatment, between field capacity and 75% available water is 2.3% as against a mean loss of 1.25 in. from the black pan. Expressed as an amount equivalent to evaporation this

TABLE 6. Relative evapotranspiration rates for citrus as a function of percentage available water, calculated from data of WEST and PERKMAN (1953).

Treatment	Moisture loss from soil 4–12 in.	Cumulative evaporation	Difference	Ratio
	% of available water	inch	inch	
Tilled	0–25	1.2	1.2	1
	25–50	3.7	2.5	0.48
Bare soil	0–25	1.3	1.3	1
	25–50	3.2	1.9	0.68
	50–75	8.9	5.7	0.23
Sod	22–25	0.2	0.2	1
	25–50	2.1	1.9	0.63
	50–75	5.2	3.1	0.39
	75–90	±18	13	0.09

TABLE 7. Average relative evapotranspiration in relation to moisture suction

Percentage available water	Relative ratios	pF-range	Suction range (atm.)
100–75	1.0	1.9 –2.8	0.08– 0.6
75–50	0.60	2.8 –3.2	0.6 – 1.5
50–25	0.31	3.2 –3.57	1.5 – 3.6
25–10	0.09	3.57–4.09	3.6 –12.0

represents a loss of $8 \times 0.023 \times 1.65 = 0.30$ in., which is withdrawn by 60% of the active root system.

The remaining 40% of active roots occur mainly below 12 in., assuming moisture extraction by these roots to be equally effective gives an extraction of $0.4/0.6 \times 0.30 = 0.20$ in. below 12 in. depth. If no roots are assumed to occur in the 0–4 in. layer an extraction rate in this layer, by pure evaporation, equal to the 4–12 in. layer results in an additional loss of 0.15 in. water. A total loss of $0.30 + 0.20 + 0.15 = 0.65$ in. water is therefore found for the whole profile. However, it is very likely that some roots were present in the 0–4 in. layer and extraction below 12 in. may not have been as effective as in the main root zone due to decreasing root density. A total extraction by evapotranspiration of 0.60 in. would therefore appear to be a more realistic figure. The ratio E_a/E_{aust} is then given by $(E_{bp}/E_{Aust}) \times (E_a/E_{bp}) = 1.37 \times (0.60/1.25) = 0.66$.

The relative ratios of tables 6 and 7, multiplied by this figure therefore give ratios of E_a/E_{Aust}, comparable with table 4. These ratios are shown in table 8, col. 3, together with averaged ratios of the pasture experiment. The latter have been averaged over periods during which potential evaporation (E_0) and available water varied within rather narrow limits, *e.g.* for the Apr.–Sept. period, E_0 ranged from 1.5–3.4 mm day^{-1} and available water was always >75%.

We return now to a discussion of the constants c, f and p in equations (6)–(8). From the discussion of FLEMING's results it is seen that for 1961 the average pan coefficient, c, of the Australian standard pan at Griffith is close to 1, thus the ratio E_a/E_{Aust} is approximately equal to E_a/E_0. Applying this result ($c \cong 1$) to the 1951 citrus experiment, it is seen that, when available water is high $(E_a =$

TABLE 8. Calculated values of the constant f (eq. 7) and p (eq. 8) in relation to potential evapotranspiration and percentage available water.

Expt.	Period	E_a/E_{Aust}	E_t/E_0 ($=f$)	E_a/E_t ($=p$)	E_0 (mm day^{-1})	Av. water (%)
Pasture	Apr.–Sep. '61	0.78	0.76	1	1.5–3.4	90
	Oct. '60, Oct/Nov. '61	0.59		0.76	5.2–6.1	42
	Dec. '60–Feb. '61	0.20		0.26	6.8–7.6	8
Citrus	Jan.–March '51	0.66	0.66	1	7.5	75–100
		0.40		0.61	7.5	50– 75
		0.20		0.30	7.5	25– 50
		0.06		0.09	7.5	10– 25

E_t), $E_a/E_{Aust} \simeq E_a/E_0 = 0.66$, therefore $f = E_t/E_o = 0.66$. For the pasture it is found, by applying the average pan coefficient for the April–Sept. period (table 2) that $f = 0.78/1.03 = 0.76$.

Comparing these figures with values of f found elsewhere it is seen for the pasture that the factor f is close to the average annual value of f found by PENMAN (1948) for conditions applying to south England. Monthly values of f show a fairly large scatter, as may be seen from the April-Sept. ratios of E_a/E_{Aust} in table 4. The variations are of the same magnitude as those of PENMAN with the exception of the August ratio, here evapotranspiration of the pasture was found to be higher than pan evaporation. The latter result is compatible with results sometimes found from continually wetted lysimeters (see *e.g.* SLATYER and MC ILROY, 1961, p. 3-11). It is observed that during this month the moisture content on the plot was very high (table 4). However, it is considered to be doubtful that the results found for the remainder of the April–Sept. period have been influenced appreciably by reduced moisture availability in the upper root zone (the ratios shown would then not represent true values of f). Moisture availability in the upper root zone did not drop below 75% during this period and recent work by DENMEAD and SHAW (1962) has shown that reduction in transpiration did not take place to an appreciable extent if available moisture was reduced to 60% at potential transpiration rates (E_t) up to 4 mm day^{-1}.

The coefficient $f = 0.66$ for citrus is high compared to the value of 0.55 commonly quoted from results of consumptive use studies in the U.S.A. (*e.g.* BLANEY, 1952). Surveying data on consumptive use factors for a variety of crops under different climatic conditions, FLEMING (pers. comm.) considered that in an arid environment, where advection of energy is prominent, values of f should be higher for tall row crops, dependent on the way the crop interferes with wind flow. He suggested for citrus a value $f = 0.65$, which is close to the value found here.

Values of $p = E_a/E_t$, calculated from the relation $p = E_a/(E_{Aust} \times c \times f)$, are shown in col. 5, table 8. For the pasture experiment this implies that the value $f = 0.76$ found for the Apr.–Sept. period is the same during the remainder of the year, which may not be strictly valid. The accuracy of the values of p shown for citrus depends on the validity of the assumptions ($E_a/E_{Aust} = E_a/E_0$, total moisture extraction at high available water 0.60 inch) made in calculating f. It is realized therefore that the values of p, given in table 8, are approximate. Further work is necessary to verify these results and to establish similar relations for other crops and climatic conditions. However, the data presented here are sufficient for the purpose of showing (in section III–3) the influence of the factors crop, climate and soil physical conditions on the rate of salinization from saline groundwater.

3. Field experiments

a. *Site selection and description*

Field experiments were carried out during the period April 1960–Oct. 1961 on a number of sites where salinization had either occurred or where this was likely to take place due to an anticipated rise of the watertable. A further consideration in the choice of sites was to cover the range of soil permeabilities (from 5–100 cm day^{-1}) and soil physical conditions normally encountered in the Murrumbidgee Irrigation Areas, so as to evaluate their effect on salinization. Special attention was paid to soils of low permeability, mainly because of the conflicting views expressed in the literature concerning salinization and watertable control in such soils (see section I-3-f). A total of 5 sites was finally selected and small plots, 3 × 3 m in area, were established on these sites.

Plot 1 was placed on a permeable soil type, plot 2 on a soil of intermediate permeability and plots 3, 4 and 5 on slowly permeable soils. The sites containing plots 3, 4 and 5 are used for a variety of purposes and consequently a variation in watertable levels, amount of irrigation water applied and vegetative cover during the period of investigation occurred. This allows an evaluation, in the final analysis, of the effect of these factors on salinization. No different treatment, *e.g.* by means of extra irrigation or cultivation, existed between the test plots and the sites as a whole. Salinization was therefore studied under current management practice on all sites. A brief description of each site is given below.

Plot 1: was established on the Irrigation Research Station, Farm 459, Hanwood. The soil is a permeable loam (a red brown earth, type name: Yandera loam) situated in a natural depression. Prior to drainage the watertable fluctuated from the soil surface to a depth of 120–150 cm and surface salting was very apparent. The area has been drained since 1958 by tile drains at 180 cm depth, which has kept the watertable at approximately 160 cm depth. From 1958 onwards this site has been under observation to check whether reclamation without leaching and cropping was possible. This has proved not to be the case (see figure 4, section I-3-e).

The plot carried a sparse vegetation of wire weed *(Polygonum aviculare)*, goose foot *(Chenopodium atriplicinum)*, couch grass *(Cynodon dactylon)* and creeping salt bush *(Atriplex semibaccata)* for the duration of the experiment. This vegetation was removed by shallow cultivation on 18 October 1960 and again on 4 January 1961.

Plot 2: was established on a salt affected pasture of the Viticultural Nursery, Farm 217, Yoogali. The soil is of sandy texture at the top and changes abruptly at 50 cm to a clay texture. The clay content increases with depth and the profile becomes rather impermeable at about 180 cm (type name: Banna sand). The watertable has fluctuated in the past from 30 cm in winter to a maximum depth of 200 cm during dry summer periods when no irrigation was carried out.

The site has numerous salt spots and plot 2 was established on one of these. Because of the high salt content in the top layers the plot was practically bare; there was only a sparse vegetation of sea barley grass *(Hordeum hystrix)* during September and October 1960.

Plot 3: was established in a furrow irrigated apple orchard at Farm 95, Yoogali, on a grey clay soil of low permeability (type name: Camarooka clay loam). The original thin surface layer of clay loam has been mixed with part of the clay subsoil by cultivation and grading. The watertable was around 60 cm during the irrigation season and receeded to 120 cm during the winter periods. This has been the normal fluctuation over the last years, except in 1956 when the winter watertable was high due to excessive rainfall. Although the watertable is generally higher than at plots 1 and 2, surface salting has never been apparent but some trees show signs of salinity damage.

The orchard is not cultivated during the winter period, so as to encourage the growth of wintergrass *(Poa annua)* as a control against waterlogging. In the summer, during the irrigation season, a number of cultivations are necessary to control the vigorous growth of barnyard grass *(Echinochloa cruss-galli)*. The plot was treated in a similar manner.

Plot 4: was established on a drainage reserve at Farm 209, Koonadan, on a grey soil of heavy texture (type name: Tuppal clay). The site is a low lying area along side a drainage channel and is not irrigated. The groundwater is generally within 100 cm from the surface all the year round and is highly saline. Surface salinity is pronounced. A patchy vegetation of sea barley grass *(Hordeum hystrix)* existed on the plot, mainly from August to December, 1960.

Plot 5: was established on a permanent pasture, about 150 m from plot 4 (Farm 151, Koonadan). The soil here was a red brown earth (type name: Jondaryan loam) of low permeability. Watertable fluctuation depends on irrigation intensity and is of the order 30 cm–200 cm. The annual selfseeding winter pasture of Wimmera rye-grass *(Lolium rigidum)* and sub-clover *(Trifolium subterraneum)* received a late autumn irrigation and was carried through the winter period by natural rainfall. During the summer period the site was not irrigated and the main growth on the plot was the deep rooting, salt tolerant, couch grass *(Cynodon dactylon)*. Salinization of the top soil was not apparent on the site. This plot was established at the beginning of October 1960, for the dual purpose of obtaining data on salt movement and on actual evapotranspiration of pasture (see section II-2-c).

b. *Methods*

During the period May 1960–June 1961 regular measurements were made on the plots of moisture content, Cl^- content and suction gradients. In addition, records were kept of watertable levels and rainfall. At the end of this period undisturbed soil samples were collected for laboratory analyses and *in situ* measurements were made of hydraulic conductivity. Figure 8 shows details of the plot design.

Samples for moisture and Cl^- content were obtained each month by direct sampling with a 2.2 cm diameter Veihmeyer tube sampler, over 10 cm depth intervals, to 160 or 180 cm. Four replicates were obtained at each depth, these were taken at a random position on each of four concentric "sampling rings". After sampling each hole was filled with screened soil, subsequent sample holes were at least 25 cm from previous holes. Towards the end of the experiment no further sampling was possible on the inner ring and sample holes were distributed over the three remaining rings.

Initially, from May to August, 1960, each sample was analysed separately, to establish the variability of measurement results. During subsequent samplings

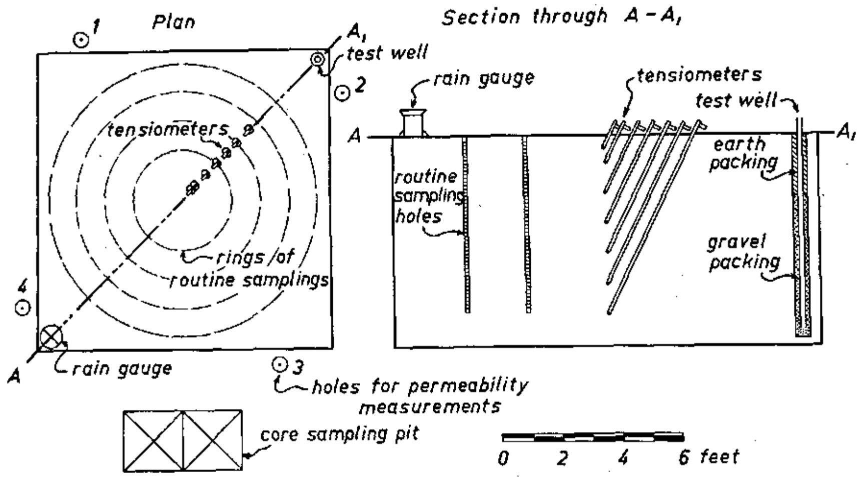


FIG. 8. Details of plot design.

the replicates for each depth interval were bulked together. Moisture content, expressed as gr. moisture/100 gr. dry soil (θ_w) was obtained by oven drying for 36 hours at 105°C. The Cl^- content (B) of these samples was obtained by electrometric titration with $AgNO_3$ on a 1:5 soil suspension.

Moisture suction was measured with tensiometers, installed at successive depth intervals of 1 foot (30.5 cm) to 5 or 6 feet below the soil surface. An additional tensiometer was located at 6 inches (15 cm) from the soil surface. All the porous ceramic tips of the tensiometers were located in a vertical undisturbed soil column at the centre of the plots, by installing the tensiometers at an angle (see figure 8). Suction was recorded on mercury manometers at plots 1, 2 and 5, and on Bourdon vacuum gauges at plots 3 and 4. Measurements of suction, together with rainfall and watertable level in an open test well (fig. 8), were recorded at least twice a week. Gauge and manometer readings were converted to express the suction, S, in cm water.

Moisture content – suction relationships were studied on undisturbed soil, consisting of core samples 8.5 cm in diameter and 2.5 cm high, for suction values up to pF = 3. For higher suctions, these cores were sub-sampled to give undisturbed samples 4.7 cm in diameter and 1 cm high. The samples were obtained in quadruplicate from a pit adjacent to each plot, at depths corresponding to the tensiometer depths. Four samples were thus obtained at depths of 15, 30.5, 61, 91.5, 122 and 152.5 cm.

The samples were saturated overnight, using filtered groundwater from the sampling site to prevent structure breakdown in the clay samples. After weighing the samples were transferred to an asbestos paper suction table for measurement of moisture content at suctions of 10, 20 and 40 cm water. Moisture contents at suctions of 80, 150, 300 and 700 cm water were measured using ceramic suction plates. Moisture at higher suctions, up to wilting point, was extracted in the pressure membrane apparatus.

Large core samples, 8.5 cm in diameter and 9 cm high were taken at the same depths, again in quadruplicate, to determine bulk density. These samples were subsequently used for particle size analysis. Percentages of clay ($< 2\ \mu$) and silt (2–20 μ) were determined by the hydrometer method, using sodium hexametaphosphate ("calgon") as dispersing agent. Samples with a high salt content were leached before dispersion. The sand fraction (20–2000 μ) was determined by decantation. None of the samples contained gravel (> 2000 μ) but the subsoil samples of plot 1 contained appreciable amounts of limestone rubble, which was separated from the soil by a 2 mm sieve and weighed separately. Results of the particle size analyses were corrected to give 100% sand + silt + clay.

Hydraulic conductivity was measured by the augerhole method, where possible over successive depth intervals. Some groundwater samples were taken to assess variability of salt content in the groundwater.

c. *Variability of measurement data*

The methods employed in this investigation have the advantage of studying moisture and salt movement on undisturbed soil profiles in their natural environment, but this advantage is partly offset by the large variability inherent in most field data. To reduce the error due to variability a large number of replicates are desirable, which however makes the investigation time consuming and also results in the early destruction of a field plot as an undisturbed site. The amount of replication at each sampling was therefore a compromise between the two requirements.

Variability of moisture, chloride content and hydraulic conductivity is indicated in table 9. For moisture and chloride contents coefficients of variation based on four replicates at each depth interval of 10 cm, were calculated and averaged over depth intervals at which the variability was reasonably uniform. The data are for the first four monthly samplings on plots 1–4 only, since analysis of individual samples over the whole period was prohibitive.

TABLE 9. Variability of moisture, chloride and hydraulic conductivity measurements, expressed as average coefficients of variation (100 × Standard deviation/Mean).

Plot No.	Layer (cm)	Moisture contents: tube samples (%)	Moisture contents: core samples (%)	Chloride (%)	Hydraulic conductivity: Depth interval (cm)	Hydraulic conductivity (%)
1	0–20	8.3	3.8	42	140–200	23
	20–180	6.5	4.2	18		
2	0–50	17.6	15.5	32	130–180	53
	50–180	8.0	2.5	21		
3	0–180	5.5	5.0	37	70–120	58
					140–200	51
4	0–30	9.8	6.7	45	20–70	44
	30–150	7.3	4.2	28	70–140	31
5					30–90	58
					90–150	12

Variability in moisture content, measured on Veihmeyer tube samples and on undisturbed core samples is comparable; the latter samples show a somewhat smaller variability due to the larger sample size. Chloride content is seen to be rather variable, especially in the top soil. The hydraulic conductivity is also quite variable, this is in agreement with previous work (TALSMA and FLINT, 1958, TALSMA, 1960).

Suction measurements at each depth were not replicated, but here an estimate of the reliability of results may be obtained by comparing suction values with watertable depth for tensiometers situated below the watertable during periods in which approximate steady state conditions exist. Also measurements of suction were made much more frequently than moisture and chloride, thus providing replication in time. The Hg manometers were accurate to $\pm$ 2 cm water suction, the Bourdon vacuum gauges to $\pm$ 7 cm.

d. *Physical properties of soils on plots 1–5*

The results of physical analyses on the core samples are given in table 10. All figures are averages of 4 replicates. Columns 3–6 are the results obtained from the large samples (8.5 $\times$ 9 cm), the data of columns 7–10 is derived from the moisture content-suction relationships obtained on the small samples (8.5 $\times$ 2.5 cm and 4.7 $\times$ 1 cm). The amount of water held at saturation has been used to indicate total porosity. A number of samples, especially clay samples from the deeper soil layers, exhibited swelling during saturation which caused excessive moisture uptake. This swelling process was not very pronounced since moisture content of the samples during sampling was generally high. Total porosity in such cases was calculated from particle density, dry bulk density and the known volume of sample rings.

Moisture percentages at saturation and in the low suction range were corrected in the case of samples exhibiting swelling by multiplying the amount of water held in the unconfined core sample by the ratio of porosity calculated from particle density to porosity deduced from saturation moisture percentage of the unconfined sample. This method of correction is empirical, a more satisfactory method is the one used by CRONEY and COLEMAN (1953) who showed that the suction, S, in cm of water in a confined sample (soil sample *in situ*) will be

$$S = S_s - a\,P \tag{15}$$

where S_s is the suction of an unconfined, "undisturbed" sample, P is overburden pressure, expressed as a height of equivalent water column and a is the fraction of overburden pressure affecting the suction. In non swelling soil $a = 0$, in entirely colloidal, saturated soil $a = 1$. In the latter case a soil of given water content, which has a suction of e.g. 200 cm in an unconfined sample has a suction of zero when in place at 100 cm below the soil surface if soil of wet bulk density of 2 g cm^{-3} lies above it.

Clay samples in the present study are not entirely colloidal and values of a are not known. In principle, these may be obtained from the other quantities of equation (15), provided hysteresis is eliminated by measuring both S and S_s at equal moisture content during *e.g.* a draining cycle. However, the results of the core sample analyses are for the draining cycle only, whilst moisture content-suction relationships obtained during the monthly samplings of moisture and chloride content have been obtained from sites subject to alternative wetting and drying. Consequently differences in moisture content obtained in the field and on core samples are to be expected, as may be seen from figure 9. The field data in this figure are seen to generally fall to the left of the curves drawn through the core sample data, although the latter have been corrected, be it in an empirical manner, for differences between the confined and unconfined state. The difference is of the order of 2–3 per cent by volume. In addition to the factors commonly contributing to hysteresis (*e.g.* "inkbottle" geometry

TABLE 10. Physical analyses of core samples.

Soil type and Plot number	Soil layer	Depth of sample	Sand > 20 μ	Silt 2–20 μ	Clay < 2 μ	Bulk density	Total porosity	Moisture contents			Available water	
								pF = 1.93	pF = 2.47	pF = 4.19	Column 1 cm²	
Dimensions and Column no.	cm	cm	%	%	%	gcm⁻³	%	gcm⁻³	gcm⁻³	gcm⁻³	cm	%
	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)
Plot 1 Yandera loam	0–30	15	67.5	11.3	21.2	1.53	38.9	.301	.263	.150	4.53	15.1
	30–50	35	53.8	11.2	35.0	1.39	47.5	.338	.306	.207	2.62	13.1
	50–70	61	60.7	15.1	24.2	1.46	44.9	.364	.317	.184	3.60	18.0
	70–110	91.5	66.0	12.3	21.7	1.61	39.2	.325	.283	.177	5.92	19.7
	110–140	122	60.5	15.8	23.7	1.50	43.3	.358	.318	.187	5.13	17.1
	140–160	152.5	60.7	12.7	26.6	1.56	41.2	.334	.295	.150	3.68	18.4
	0–160										25.48	15.9
Plot 2 Banna sand	0–25	15	81.6	4.5	13.9	1.53	42.2	.294	.196	.058	5.90	23.6
	25–50	30.5	83.6	3.1	13.3	1.53	42.2	.257	.185	.067	4.75	19.0
	50–80	61	50.9	10.3	38.8	1.42	46.3	.354	.315	.249	3.15	10.5
	80–110	91.5	45.3	11.5	43.2	1.41	46.7	.359	.330	.261	2.94	9.8
	110–140	122	51.3	8.8	39.9	1.47	44.5	.358	.326	.282	2.28	7.6
	140–160	152.5	44.5	7.8	47.7	1.49	43.7	.364	.341	.280	1.68	8.4
	0–160										20.70	12.9
Plot 3 Camarooka clay loam	0–20	15	51.1	9.7	39.2	1.51	45.4	.371	.347	.227	2.88	14.4
	20–45	30.5	39.5	12.4	48.1	1.32	49.3	.407	.367	.255	3.80	15.2
	45–80	61	43.9	9.0	47.1	1.48	42.9	.378	.354	.260	4.13	11.8
	80–110	91.5	42.8	8.1	49.1	1.56	41.9	.368	.346	.287	2.43	8.1
	110–135	122	43.2	6.4	50.2	1.62	38.9	.359	.339	.297	1.55	6.2
	135–160	152.5	45.2	6.4	48.4	1.62	38.9	.348	.327	.278	1.75	7.0
	0–160										16.54	10.3
Plot 4 Tuppal clay	0–20	15	37.1	13.0	49.9	1.44	42.9	.367	.338	.235	2.64	13.2
	20–45	30.5	36.1	16.7	47.2	1.49	42.9	.377	.350	.240	3.43	13.7
	45–70	61	41.0	14.6	44.4	1.47	44.7	.374	.356	.277	2.43	9.7
	70–100	91.5	42.2	11.8	45.8	1.56	41.4	.355	.338	.244	3.33	11.1
	100–130	122	44.1	12.9	32.0	1.58	40.3	.348	.320	.226	3.66	12.2
	130–160	152.5	36.6	16.4	47.0	1.59	40.0	.351	.328	.204	4.41	14.7
	0–160										19.90	12.4
Plot 5 Jondaryan loam	0–15	10	73.0	9.0	18.0	1.60	35.2	.266	.227	.094	2.58	17.2
	15–45	30.5	33.0	11.2	55.8	1.41	42.0	.350	.348	.279	2.13	7.1
	45–80	61	34.7	13.6	51.7	1.46	45.4	.406	.382	.283	4.30	12.3
	80–110	91.5	41.2	16.1	42.7	1.54	40.8	.348	.322	.262	2.58	8.6
	110–140	122	43.4	19.5	37.1	1.57	36.1	.330	.303	.222	3.24	10.8
	140–160	152.5	42.9	19.7	37.4	1.55	34.6	.305	.282	.211	1.88	9.4
	0–160										16.71	10.4

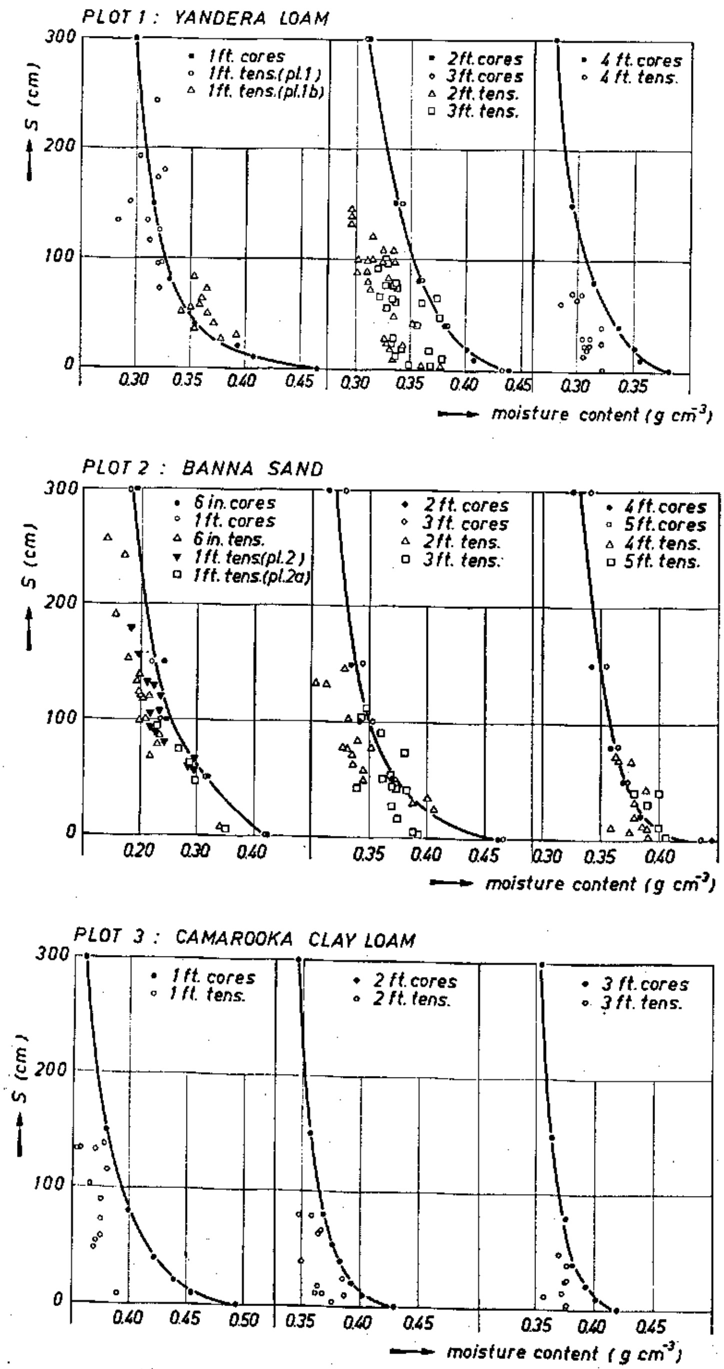


FIG. 9. Moisture content suction relationships, plots 1–.3

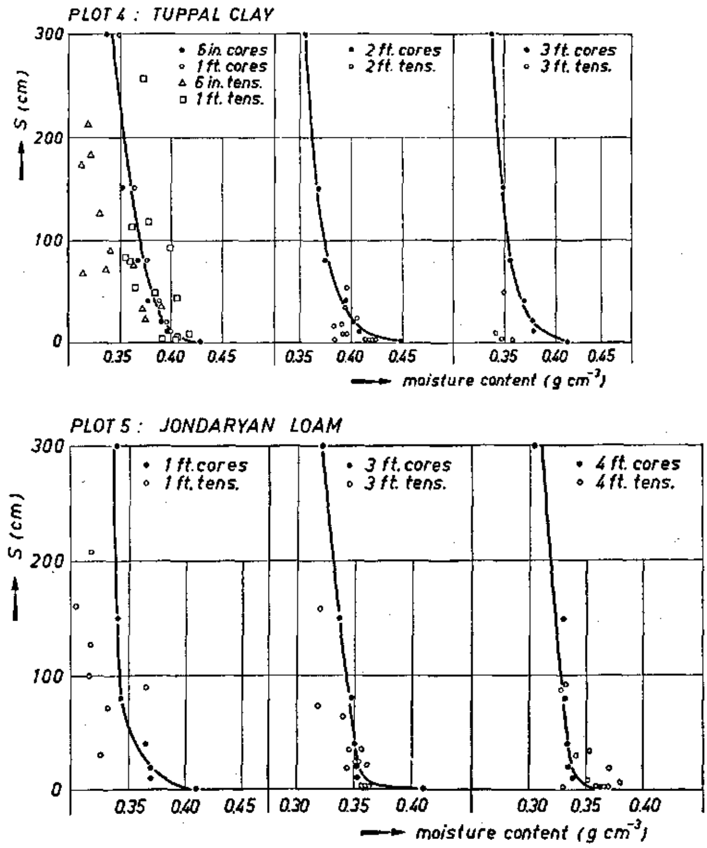


FIG. 9
Moisture content suction relationships, plots 4–5.

of pores, time factor effects), entrapped air must also considered to be a factor. In this regard it is of interest to note the position of field sampling moisture contents plotted along the X-axis in figure 9. These represent moisture contents obtained at or below the watertable, for which in place suction values were either zero or negative (positive hydrostatic pressure). It is seen that the majority of these are also to the left of the laboratory data. Air entrapment occupying from 1–5 per cent of the pore space below a permanent watertable may be deduced from these data.

Some discrepancy may further exist between core sample and plot moisture contents due to minor variations in soil properties between the plot and pit sites. In utilizing moisture content data in this study an indication will be given, where necessary, as to which set of measurements have been used. For example in figure 7 (section II-2-c) field moisture contents rather than core sample data have been used to plot the pF curves shown.

Continuing the discussion of table 10, available water (col. 11), expressed as the depth of water contained in a soil column of 1 cm^2, is obtained by multiplying the difference of columns 8 and 10 by the thickness of the soil layer for which the sample is representative. Column 12 gives the same quantity, expressed as a percentage of the total space. The moisture content af pF = 1.93 (80 cm suction) is taken here to represent field capacity. This is in agreement with data presented by GREACEN and PERKMAN (1953) on "aeration porosity" and TALSMA and HASKEW (1959) on "drainage porosity" values. The latter authors used a suction value of 60 cm to indicate drainable porosity but found from a comparison of values thus obtained and those deduced from watertable drawdown data that a slightly higher suction would estimate drainable porosity better. The 80 cm suction percentage (col. 8) is considerably higher than the moisture content at the more commonly used suction value of 300 cm (col. 9).

The data presented in table 10 is given in graphical form as figure 10, which shows the percentage of the total space occupied by sand, silt, clay, unavailable

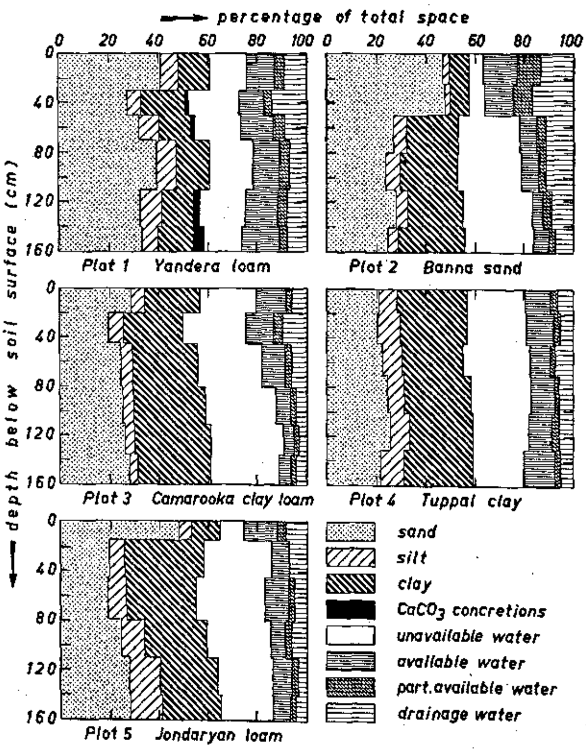


FIG. 10.
Diagrams of soils on plots 1–5.

water (held at suctions > 15 atm.), available water (held between the conventional limits of 300 cm and 15 atm. suction), easily or partially available water (between 80–300 cm suction) and drainage water (< 80 cm suction). The percentage of space occupied by carbonate concretions > 2 mm is included in the diagram for Yandera loam.

Table 11 gives the results of hydraulic conductivity measurements. At plots 1 and 2 data could only be obtained for the deeper layers, but the average values do not deviate much from average figures, obtained during the survey for potential test sites, over the depth interval of 90–210 cm. The latter values are entered in column 4.

There is generally a good relationship between hydraulic conductivity and

TABLE 11. Results of hydraulic conductivity measurements.

Plot no.	Depth interval (cm) (1)	Number measurements (2)	k_{sat} (cm day^{-1}) (3)	k_{sat} used in fig. 13 (cm day^{-1}) (4)	$EC.10^3$ groundwater (millimhos cm^{-1}) (5)	Cl^- in groundwater (%) (6)
1	140–200	3	69.2 ± 16.0	75.0	11.0 ± 1.5	.415 ± .057
2	130–180	4	26.8 ± 14.7	27.4	6.2 ± 1.8	.151 ± .051
3	70–120	4	8.1 ± 4.7	8.1		
	140–200	4	6.5 ± 3.3		10.7 ± 1.4	.339 ± .051
4	20–70	3	12.9 ± 5.7			
	70–140	3	17.6 ± 5.4		19.3 ± 5.7	.688 ± .245
	20–140	10	16.9 ± 5.1	16.9		
5	30–90	3	4.3 ± 2.5			
	90–150	3	7.6 ± 0.9	6.0		

drainable pore space (fig. 10, table 10, col. 7–col. 8). The average hydraulic conductivity of profile 4 is somewhat higher than expected for a soil of this clay content, this is seen to be associated with a higher drainable pore space (compare profiles 3 and 5 with profile 4, all these have about the same clay content). Profile 2 has a high drainable porosity in the top layers down to 110 cm, the hydraulic conductivity value of table 11 is strictly valid only for the lower layers and may be higher for the top layers at the plot, although the site average is not (table 11, col. 4).

Comparing the results for the various plots it is seen that profiles 1 and 5 have a clearly distinguishable zone of clay accumulation (the B-horizon of red brown earths) but the bulk density of this layer is low compared to the top and subsoil. Due to this these B-horizons are not considered to be restricting layers. The same trend in bulk density is apparent in profiles 2 and 3. The soil profile of plot 4 is rather uniform regarding its mechanical composition, bulk density increases gradually with depth and as a consequence porosity shows a gradual decrease.

Due to the high clay content and the relatively poor structure of the soils investigated, the amount of unavailable water, held above 15 atm., is high. Apart from profile 1 and the topsoils of profiles 2 and 5, unavailable water occupies more than half the total pore space (figure 10). The amount of available water (table 10, col 11–12, fig. 9 and 10) is highest in profile 1, especially in the subsoil, and in the sandy top layer of profile 2. The subsoil materials of profiles 2, 3 and 5 have low amounts of available water, profile 4 has reasonable storage capacity for a fine textured soil. Available water is also low in the B-horizons of profiles 1 and 5, the latter profile has hardly any storage capacity in this layer between 80 and 300 cm suction (fig. 9 and 10).

The amounts of available water, shown in table 10, col. 11 and 12, being calculated from desorption curves, are maximum values. Hysteresis and air entrapment may reduce these amounts in the field, reasonable field percentages for the clay soils of plots 3, 4 and 5 would be about 8 per cent for the whole profile down to 160 cm depth, 10 per cent for the Banna sand (plot 2) and 14 per cent for the Yandera loam (plot 1).

e. *Salinization on plots 1–5*

Chloride content and distribution in the soil profiles. – The distribution of chloride with depth at the plots is presented in figure 11. The chloride profiles are the average of two or three samplings, *e.g.* the winter curves are the average of samplings at the end of June, July and August. The average watertable depth for each season is shown in brackets, in the legends.

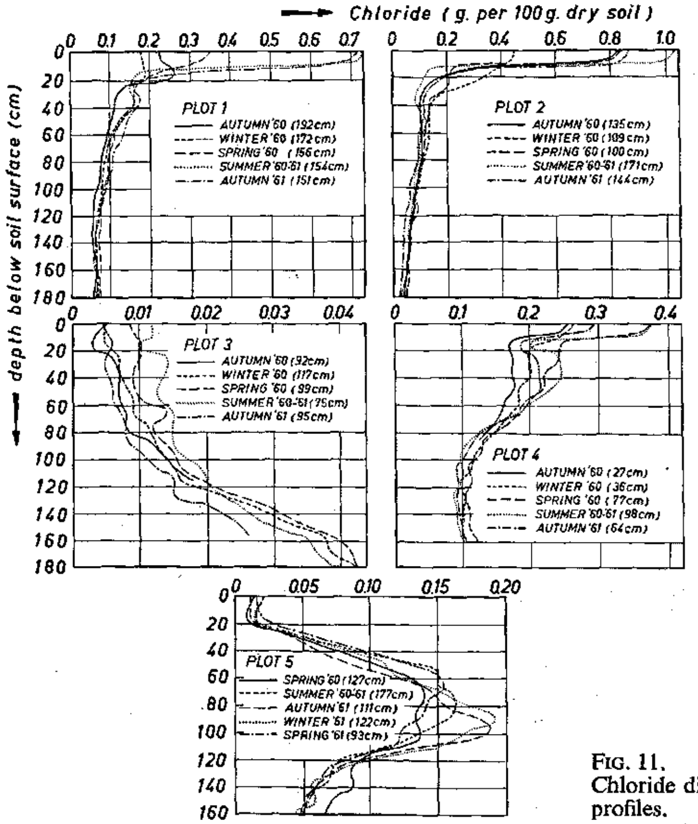


FIG. 11.
Chloride distribution in the soil profiles.

Table 12 gives the amounts of sodium chloride in the soil profiles down to 160 cm, for periods identical to those of figure 11. A note of explanation must be made in regard to the data shown in brackets (plot 2, spring and plot 4, summer). It is remembered that plots 2 and 4 were practically bare, which rendered them liable to wind erosion and runoff. These processes were probably responsible for removal of part of the saline crust, present on these sites during the periods concerned, beyond the boundaries of the plots. The quantities shown in the table, although correctly measured, do therefore not represent the total amounts, and chloride accumulation can not be deduced for these periods.

Calculation of capillary flow from moisture and chloride contents. – The measurements of moisture and chloride content at monthly in-

TABLE 12. NaCl content (kg) of a soil column 1 m² × 1.60 m.

Plot No.	Autumn 1960	Winter 1960	Spring 1960	Summer 1960–'61	Autumn 1961	Winter 1961	Spring 1961
1	4.07	4.27	4.97	5.53	6.04	–	–
2	5.86	5.94	(5.66)	6.05	5.53	–	–
3	0.58	0.61	0.69	0.72	0.46	–	–
4	5.68	6.08	6.81	(6.22)	5.99	–	–
5	–	–	3.57	3.67	3.64	3.89	3.60

tervals enable calculation of the rate of capillary flow from the watertable, by dividing the monthly salt increment of a layer by the average salt concentration of the inflowing groundwater (VERHOEVEN, 1950, RICHARDS *et al.*, 1956). For this purpose the salt content, given in per cent, of each 10 cm interval is converted to obtain *a*: the concentration of the soil moisture solution, and *b*: the amount of chloride stored in a soil column 10 cm long and with unit surface area (1 cm²).

Table 13 gives an example of this procedure, using the data of plot 1, for the period between samplings on 4-10 and 31-10-1960. The measurement data are given in columns 2–4, column 5 gives the initial and end concentration of chloride in the soil moisture $C_{Cl^-} = 1000B/\theta_w$ gl⁻¹) and column 6 gives the initial and final content of chloride in a soil column of 1 cm² surface area ($Z = B\ d_b/10$ gcm⁻²). The total increase, ΔZ is the sum of the increases over each interval to the watertable.

In the example it is found that $\Delta Z = 0.034$ gcm⁻² down to 150 cm, at this depth (col. 5) C_{Cl^-} has the constant value of 3.2 gl⁻¹. Therefore the total capillary flow over the period is $0.034/3.2 \times 10^4 = 106$ mm; the average daily flow, $q = 106/27 = 3.9$ mm.

TABLE 13. Conversion of measurement data for calculation of capillary flow. (Plot 1, period 4-10 to 31-10, 1960).

Layer cm	Bulk density d_b	Cl^- gr/100 gr soil (B)		Moisture cc/100 gr soil (θ_w)		Cl^- in soil moisture gr l⁻¹ (C_{Cl^-})		Cl^- per unit area gr cm⁻² (Z)		Cl increment gr cm⁻² (ΔZ)
(1)	(2)	(3)		(4)		(5)		(6)		(7)
		4-10	31-10	4-10	31-10	4-10	31-10	4-10	31-10	
0–10	1.53	.204	.376	13.9	10.2	14.7	36.9	.031	.058	+ .027
10–20	1.53	.214	.265	15.7	14.4	13.6	18.4	.033	.041	+ .008
20–30	1.53	.174	.164	18.5	17.0	9.4	9.6	.027	.025	— .002
30–40	1.39	.170	.183	22.9	21.2	7.4	8.6	.024	.025	+ .001
40–50	1.39	.148	.157	22.0	22.1	6.7	7.1	.021	.022	+ .001
50–60	1.46	.125	.120	20.7	21.8	6.0	5.5	.018	.018	–
60–70	1.46	.120	.114	20.7	20.8	5.8	5.5	.018	.017	— .001
70–80	1.61	.099	.091	20.3	19.8	4.9	4.6	.016	.015	— .001
80–90	1.61	.101	.095	21.2	19.9	4.8	4.8	.016	.015	— .001
90–100	1.61	.098	.097	20.9	20.8	4.7	4.7	.016	.016	–
100–110	1.61	.082	.085	20.6	20.3	4.0	4.2	.013	.014	+ .001
110–120	1.50	.078	.077	21.6	20.0	3.6	3.8	.012	.012	–
120–130	1.50	.078	.079	21.1	20.5	3.7	3.9	.012	.012	–
130–140	1.50	.070	.078	21.3	20.1	3.3	3.9	.011	.012	+ .001
140–150	1.56	.073	.070	22.8	22.2	3.2	3.2	.011	.011	–
150–180						3.0	3.3			

The concentration of chloride in the soil moisture solution was reasonably constant in time, at watertable depths occurring during the investigation, on plots 1 and 2 but showed variation on plots 3, 4 and 5. Here use was made of average concentrations, calculated from initial and end concentrations for the periods concerned (RICHARDS *et al.*, 1956). Where chloride concentration showed a strong variation with depth (*e.g.* plots 3 and 5) average concentrations found at about watertable depth were used.

Negative adsorption (BOWER and GOERTZEN, 1955) is to be taken into account when using chloride transfer in soils as a means for evaluating capillary flow rates (RICHARDS *et al.* 1956). In the present study, where soils with a large specific soil particle surface area and relatively high solute concentrations are studied, this factor is likely to be important. The effect of negative adsorption and possibly dispersion (DAY, 1956, DAY and FORSYTHE, 1957) effects (the data do not allow a distinction to be made between the two factors) may be seen from a comparison of chloride concentration measured on groundwater samples collected from auger holes and analyses of soil samples taken from the same holes or in the immediate vicinity. The results of this comparison are shown in table 14. In general it is seen that the solute concentration calculated

TABLE 14. Comparison of chloride percentages in groundwater samples and soil samples.

Site	1	2	3	4
groundwater	0.37*	0.34, 0.50, 0.89	0.30, 0.43, 0.071	0.75, 1.13, 0.28
soil sample	0.41*	0.26, 0.36, 0.67	0.23, 0.35, 0.055	0.45, 1.00, 0.20
Plot	1	2	3	4
groundwater	0.42*	0.15*	0.34*	0.69*
soil sample	0.32*	0.20*	0.23*	0.51*

* Average figures.

from soil analyses ($C_{Cl^-} = 100B/\theta_w\%$) is appreciably lower than the concentration in corresponding groundwater samples, that is in water held generally in the larger pores only. This is seen to be especially the case on sites 3 and 4, the data on sites 1 and 2 are not conclusive and no comparison is available for plot 5.

In the salt transfer process studied here, moisture conduction in the large and intermediate pores predominates and, where negative adsorption occurs, the solute concentration is therefore higher than indicated by analyses of soil samples. For the calculation of capillary flow rates, solute concentrations calculated from soil analyses have been adjusted wherever necessary to take account of differences in groundwater and soil sample concentrations.

Average daily rates of capillary flow, q, were calculated in this manner, for all periods during which salt accumulation occurred. These flow rates are shown in table 15. Leaching occurred during the periods for which no entry is shown: it is seen *e.g.* that leaching occurred in the wet month of July, 1960, on all plots investigated at that time. Surface removal of salt on plots 2 and 4, as already mentioned previously (discussion of table 12) prevented calculation of capillary flow on plot 2 for the periods of November and December, and on plot 4 for the December period. Due to the variability on the plots, especially of the chloride content in the surface layers (table 9), appreciable errors may exist in the calculation of some of these flow rates.

TABLE 15. Average daily rates of capillary flow in relation to watertable depth, rainfall and average suction in the top layers.

Plot	Period 1960–1961	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	April	May
	E_0	1.9	1.6	1.2	1.7	2.2	5.7	6.4	7.3	7.9	7.8	5.4	3.1	2.4
1	q (mm day^{-1})	–	1.03	–	1.92	2.16	3.91	1.29	1.29	0.13	–	5.00	2.64	
	Rain (mm)	38	20	54	40	50	4	42	46	3	42	45	45	
	Watertable (cm)	194	189	174	155	156	157	156	152	170	146	142	153	
	Suction at 15 cm (cm)	–+	230	140	163	233	336	?	?	> 700	> 700	–	–	
	Suction at 30.5 cm (cm)	123	162	130	131	168	201	146	114	?	?	–	–	
2	q (mm day^{-1})	–	0.9	–	0.9	1.7	0.8	?	?	2.24	2.24	2.1	–	–
	Rain (mm)	38	22	47	39	50	4	39	26	2	37	43	42	0
	Watertable (cm)	144	135	111	83	80	100	115	125	175	200	165	136	124
	Suction at 15 cm (cm)	123	131	91	68	71	93	116	115	184	241	195	151	–+
	Suction at 30.5 cm (cm)	115	122	79	53	55	74	96	97	151	196	165	116	119
3	q (mm day^{-1})	2.38	0.41	–	1.06	2.9	–	0.1	2.4	0.23	0.54	–	–	1.8
	Rain (mm)	12	20	55	36	53	5	45	22	2	47	41	44	2
							2×irr.*	2×irr.	2×irr.	2×irr.	2×irr.	1×irr.		1×irr.
	Watertable (cm)	92	117	128	111	130	103	67	66	82	73	72	96	116
	Suction at 15 cm (cm)	–+	457	147	205	179	234	81	227	195	215	210	152	513
	Suction at 30.5 cm (cm)	83	142	70	55	78	85	48	150	153	97	128	142	278
4	q (mm day^{-1})	2.4	0.9	–	2.2	2.3	2.4	0.14	?	1.9	–	–	1.1	–
	Rain (mm)	19	11	81	47	37	5	69	34	5	23	33	38	1
	Watertable (cm)	29	50	32	23	33	77	102	96	100	96	78	47	68
	Suction at 15 cm (cm)	32	75	27	15	40	470	502	282	258	253	123	59	102
	Suction at 30.5 cm (cm)	7	37	2	–7	10	156	194	107	107	103	76	35	65
	Period 1960–1961	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.
5	q (mm day^{-1})	0.37	1.08	–	1.04	0.29	–	0.34	1.33	0.55	0	0	0.10	1.55
	Rain (mm)	5	69	34	0	23	33	38	1	53	30	84	15	20
							1×irr.							
	Watertable (cm)	108	149	160	171	194	110	102	121	123	121	78	87	120
	Suction at 61 cm (cm)	61	127	180	384	314	43	42	68	64	51	4	34	142

* irr = irrigation.
\+ Tensiometer not installed or not operative.

General remarks on salinization. – The seasonal trends in salinization at each plot will be examined in the remaining paragraphs of this section. Figure 11, table 12 and the capillary flow rates shown in table 15 provide the basis for this discussion. A more complete analysis of the salinization process in relation to the variables watertable depth, suction at the root zone, rainfall, evapotranspiration and soil physical conditions will be made in sections II-4, a, b, c, and d.

Figure 11 shows that the general shape of the chloride distribution curves does not alter greatly from season to season. Noting the difference in scales for the chloride percentages on plots 1–5, it is seen that the largest concentrations of chloride were found in the topsoils of the permeable sites 1 and 2. Plot 4 also shows distinct accumulation in the surface layers, the irrigated plots 3 and 5 contain little salt in the surface horizon. It is perhaps of interest to note that the chloride quantities and distributions of the latter plots are not very different from the general curves for grey soils of heavy texture (plot 3) and red brown earths (plot 5) in the virgin state (see figure 5). Seasonal trends for each profile are discussed below.

Plot 1, Yandera loam: The average concentration of chloride in the groundwater remained fairly constant at about 3 gr l^{-1} throughout the period of observation. The lowest chloride content in the soil profile was found in the autumn of 1960 (table 12), when the watertable was deep (figure 11). In the winter period some chloride was leached out of the surface horizons but the total amount in the profile was higher than in autumn, due to an increase of chloride in the subsurface layers. Capillary transport in the winter period was therefore greater than the leaching action of winter rains. During spring, summer and autumn, increasing quantities of chloride were found in this profile, indicating a continual process of salinization. The watertable during these seasons was maintained at about 150 cm by a nearby open drain, which has a higher water level during the irrigation season (spring, summer and early autumn). This watertable level persisted to the end of the period of observation due to inadequate clearing of internal drainage water.

Plot 2, Banna sand: The seasonal trends in chloride content and distribution in the groundwater and soil profile are very similar to the trends observed at plot 1, except that desalinization occurred in the autumn (1961) period. Apart from a reduction in chloride at the surface, it is also seen (fig. 11) that the chloride concentration in the deep subsoil from 90 cm onwards is lower in this period. Associated with this is a lower concentration of chloride in the groundwater. It is not clear why this occurred but as a result of these low subsoil concentrations the total amount of chloride in the profile (table 12) during the autumn of 1961 was considerably lower than for any other period.

During winter and spring salt accumulation was less intense than at plot 1 (tables 12 and 15) even though the average watertable was higher. In the summer period more salt accumulated compared to plot 1, although the watertable was lower.

Plot 3, Camarooka clay loam: The chloride concentrations in this profile are low in the topsoil and increase gradually with depth. Groundwater salinity also increases rather rapidly below the watertable and is not at its maximum value at 180 cm. Salinization is seen to occur mainly during the spring period and was at its maximum rate just prior to the onset of irrigation early in October (table 15). The frequent irrigations in summer resulted in a high watertable and some further salinization occurred during this period. The less frequent early autumn irrigations (and rainfall) have resulted in a lowering of the watertable and desalinization occurred, here also combined with a reduction in groundwater salinity, similar to plot 2.

Compared to plot 1, considerably less salt accumulated in this profile, despite the fact that the watertable was much higher. This is partly due to the lower concentration of the groundwater (0.7 gr l^{-1} on the average just below the watertable), but even when corrected for concentration the quantities are much less. The winter and spring periods are valid only for this comparison, in these periods no irrigation water was applied to plot 3 and vegetative growth was comparable on both plots.

Plot 4, Tuppal clay: It is noted that, in comparison with plots 1–3 and 5, the watertable is much higher throughout the period of observation. The groundwater is also more saline (5 to 6 gr l^{-} at 100 cm depth). Salt accumulation was considerable during the winter and spring season, but there is no significant accumulation during the summer period (fig. 11). Figure 11 shows the average chloride profile for the summer period, excluding the data for December, when surface runoff caused desalinization. Desalinization occurred again during the autumn period.

Plot 5, Jondaryan loam: The groundwater salinity of this profile is variable between 3–6 gr l^{-1} and is therefore comparable to plots 1, 2 and 4. Chloride accumulation on a seasonal basis occurred during the summer and winter period. The variation in chloride content is found mainly between 40–110 cm depth. The amounts accumulated are small compared to plots 1 and 4, but are comparable to plot 3, when corrected for groundwater concentration.

To conclude, it is seen that chloride accumulation has occurred generally during the winter, spring and summer periods, and that desalinization has been confined mainly to the autumn period. The most intense salinization was on plot 1, where the watertable is relatively deep but the profile has a high permeability. The magnitude of chloride accumulation was of the same order on plot 4, where the watertable is high and the profile is slowly permeable. Plots 3 and 5 have profiles of very low permeability and the groundwater level was intermediate between profiles 1 and 4. Here the magnitude of salinization was much less. Plot 2 had a high amount of Cl^{-} in the surface from the onset and further salinization of this plot was slight.

f. *Measurement of suction and capillary conductivity*

Tensiometer records. – The course of soil moisture suction with time was plotted graphically for each tensiometer on plots 1–5. Complete records of the six tensiometers at plot 2 are shown in figure 12, together with rainfall and watertable recordings. Since a tensiometer below the watertable records watertable depth under non artesian conditions (a condition which applies here), the scale for the latter, on the right hand side in fig. 12, was chosen to coincide with the recordings of the tensiometer installed at 152.5 cm (5 feet) depth. It is seen that the agreement between the two is good. All tensiometers on plot 2 recorded low suctions throughout the period of observation and suction appears to be mainly determined by watertable depth and rainfall.

Suction measurements of the deeper tensiometers were very similar on the other plots, but higher suctions occurred in the top of these soil profiles. In summer the suction in the top soil was often outside the tensiometer range. An

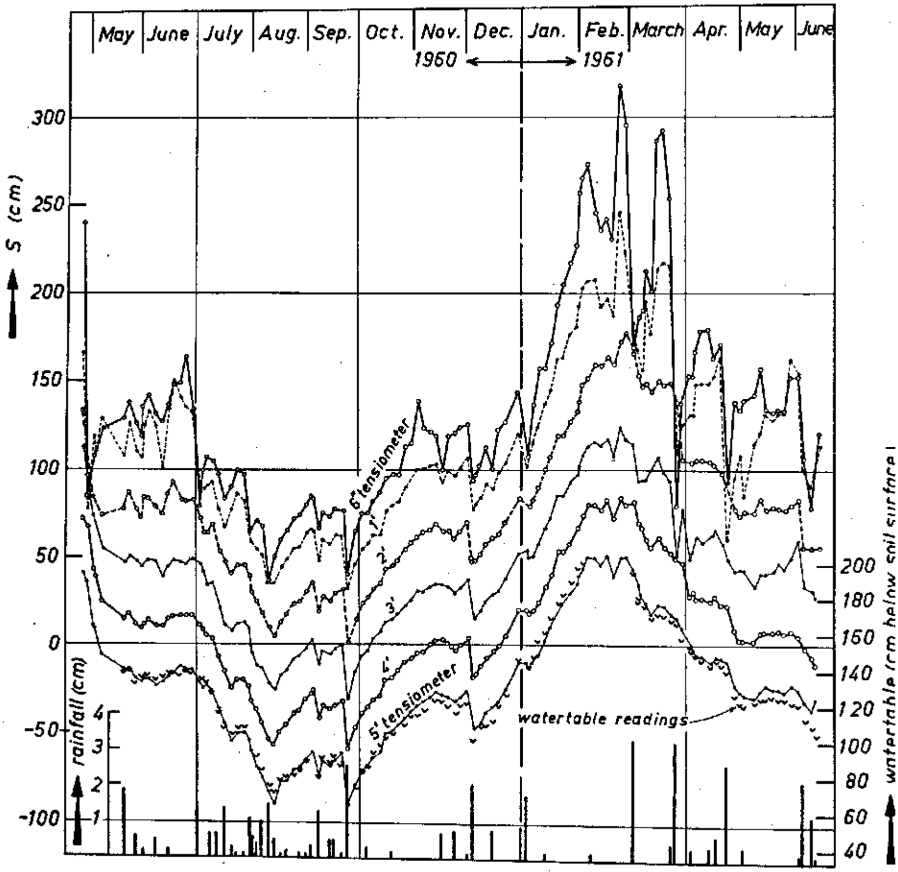


FIG. 12. Tensiometer recordings on plot 2.

indication of this is given in table 15, which presents average monthly suction values for tensiometers installed at 15 cm (6 inch) and 30.5 cm (1 ft) with the exception of plot 5, where the suction in the top soil was outside the tensiometer range for long periods. Here results are shown for the tensiometer at 61 cm (2 ft) depth.

It is seen that on plot 1 suction at 15 cm depth has increased beyond the tensiometer range (700 cm) during January and February, at 30.5 cm depth suction was outside the tensiometer range for part of the periods concerned. This was also the case for the 15 cm tensiometer during November and December. No figure is therefore entered in table 15 for average suction at these depths during these months. Sharp fluctuations in watertable level during March and April caused a large fluctuation in suction and average suction data are therefore not shown here. Plot 3 was irrigated in October, with the result that suction in the top layers fluctuated rather sharply and the average suction figures in table 15 have perhaps little meaning.

Calculation of capillary conductivity. – Using the unsaturated flow formula $q = -k$ grad Φ, the capillary conductivity, k_{unsat}, may now be calculated. Average daily flow rates, q, are given in table 15 and the potential gradient, grad Φ, may be derived from the tensiometer recordings (*e.g.* figure 12). The total potential for a tensiometer at height, z, above an arbitrary reference plane is $\Phi_1 = \psi_1 + z$, where ψ_1 is the potential in cm and, if the reference plane is chosen at the level of a lower tensiometer, the potential there is $\Phi_2 = \psi_2$, therefore grad $\Phi = (\psi_1 + z - \psi_2)/z$. The relation between potential (ψ) and suction (S) of a tensiometer is given by $S = -\psi$. The capillary conductivity thus calculated corresponds to the average suction $S = (S_1 + S_2)/2$.

Some examples of calculating capillary conductivity in this manner are given in table 16. Results of all such calculations are presented in figure 13, which

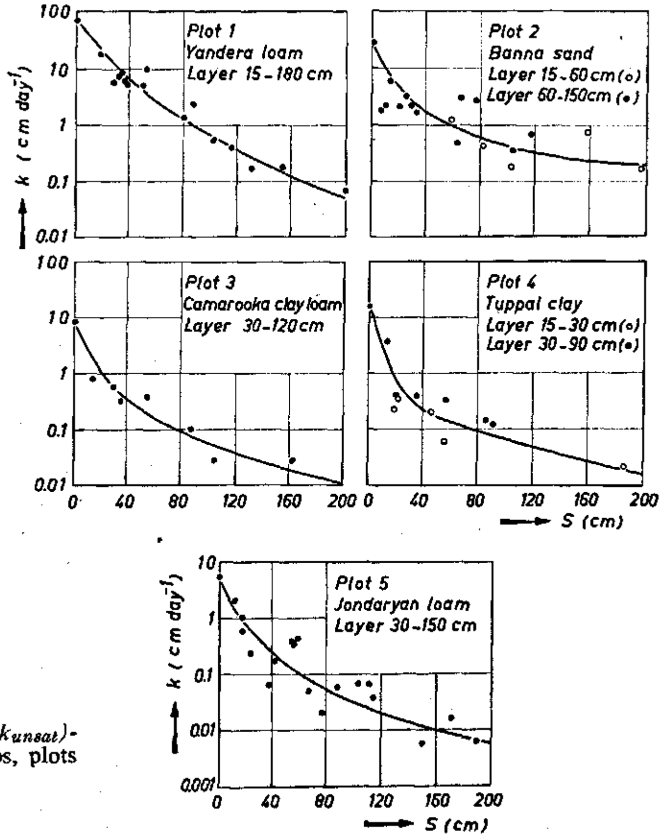


FIG. 13.
Capillary conductivity, (k_{unsat})-suction (S) relationships, plots 1–5.

gives the relationship between soil moisture suction and capillary conductivity for all plots. The hydraulic conductivity values of table 11, col. 4, have been used to extend the relationship to $S = 0$ (saturation).

Only periods during which approximately steady state conditions occurred have been used in computing data for figure 13. The main criterion in this choice was provided by the tensiometer records. Data have generally only been used whenever there was a continuous upward gradient for the period under consideration and suction did not vary over a wide range. These conditions were

TABLE 16. Calculation of capillary conductivity from tensiometer data and capillary flow rates.

Plot and period	Tensiometer below surface	Suction range over period S	Average potential ψ	Gravity potential z	Total potential Φ	Potential diff. $\Phi_1-\Phi_2$	Depth interval z	Gradient $(\Phi_1-\Phi_2)/z$	Capillary flow q	k_{unsat}	$(S_1+S_2)/2$
	cm	cm	cm	cm	cm	cm	cm	cm cm^{-1}	mm day^{-1}	mm day^{-1}	cm
Plot 1. Oct. 1960	30.5 61.0	156–262 99–122	–201.2 –104.6	+30.5 0	–170.7 –104.6	–66.1	30.5	–2.17	3.91	1.80	153
Oct. 1960	91.5 152.5	67– 75 0– 8	– 71.3 – 5.7	+61.0 0	– 10.3 – 5.7	– 4.6	61.0	–0.075	3.91	51.1	38
June 1960	91.5 122.0	93–105 55– 77	– 99.3 – 66.4	+30.5 0	– 68.8 – 66.4	– 2.4	30.5	–0.079	1.03	13.1	83
Plot 5. Feb. 1961	91.5 122.0	230–328 80–142	–266.5 –113.0	+30.5 0	–236.0 –113.0	–123.0	30.5	– 4.03	0.29	0.072	190
Nov. 1960	91.5 122.0	44– 84 11– 41	– 68.0 – 28.6	+30.5 0	– 37.5 – 28.6	– 8.9	30.5	–0.292	1.08	3.70	48
Oct. 1961	91.5 122.0	10– 65 –21– 14	– 32.9 + 1.9	+30.5 0	– 2.4 + 1.9	– 4.3	30.5	–0.14	1.55	11.1	17

not easily met in the 15–30.5 cm layer or on plot 3, which was frequently irrigated.

Distinction is made in fig. 13, plots 2 and 4, between capillary conductivity calculated for top soil and sub soil, since the profile diagram of plot 2 (fig. 10) and results of hydraulic conductivity measurements on plot 4 (table 11) indicate that the top layers on these plots may have conductivities differing from the subsoils. The results (figure 13) indicate that this is probably true at plot 4, where both hydraulic conductivity and capillary conductivity of the top soil tend to be lower than for the subsoil.

The main cause of the scatter in the data of figure 13 is due to the variability of the chloride data, with resulting errors in estimating average flow rates, q. Errors in calculating suction gradients are a contributing factor, especially where vacuum gauges were used (plots 3 and 4) and where small suction gradients existed (plot 2).

There is good agreement between hydraulic conductivity, obtained by the augerhole method and capillary conductivity, computed from chloride accumulation. For example it is seen in figure 13 that profile 1, with a high hydraulic conductivity has correspondingly high values of capillary conductivity at low suction values, profile 5 which has a low hydraulic conductivity has correspondingly low capillary conductivity values in the low suction range. A note of explanation must be made, however, regarding this comparison. The augerhole method measures largely the horizontal component of hydraulic conductivity but for measurements of capillary conductivity flow was entirely in the vertical direction. Comparison of the two sets of data is only reasonable in the case of isotropic soil.

The results of a study on anisotropy (TALSMA, 1960) at the sites containing plots 1 and 2 indicated that soils here were largely isotropic. Fine textured soils of the Murrumbidgee Irrigation Areas (plots 3, 4 and 5) show no characteristic layering indicating possible anisotropy and data on the shape of watertable between tile drains installed in such soils (TALSMA and HASKEW, 1959, their figure 7) were consistent with conditions of isotropy.

In relation to the hysteresis phenomenon it is pointed out that the calculation of capillary conductivity is based on results of field measurements only. No use has been made of the laboratory data and values of k_{unsat} obtained correspond therefore to the moisture content-suction relationships found for the field data (see discussion of figure 9, section II-3-d).

4. COMPARISON OF FIELD DATA WITH STEADY STATE THEORY

a. *Flow in the liquid phase*

The data of figure 13 may be used to solve the equation describing steady state upward movement of moisture in the liquid phase through unsaturated soil. This equation is of the form:

$$z = dS/(1 + q/k) \tag{16}$$

where z = watertable depth (cm), S = suction (cm), q = flow rate (cm day^{-1}) and k = capillary conductivity (cm day^{-1}). PHILIP (1957) assumed no special relation between S and k and, from experimental curves relating suction and conductivity, obtained solutions of equation (16) by numerical integration. GARDNER (1958) suggested that the relation between S and k can often be represented by a function of the type:

$$k = a/(S^n + b) \tag{17}$$

where a, b and n are constants. The ratio a/b is the hydraulic conductivity, the value of n appears to vary from 1.5 to 4, generally the coarser the texture of the soil the greater the value of n. GARDNER presented analytical solutions of (16) using eq. (17) with values of $n = 1.5, 2, 3$ and 4.

WIND (1955) obtained a solution of equation (16) for $k = aS^{-3/2}$. This solution was expanded by VISSER (1959) into a more general form by putting $a = D$ = the permeability of nearly saturated soil (the relation $k = aS^{-n}$ does not hold at saturation) and using values of n between 1.5 and 3. His solution is presented in nomographic form, the solutions of GARDNER may be obtained from the nomogram with the aid of a correction factor.

Relations of the type $k = a/(S^n + b)$ that fit the data of figure 13 reasonably well in the suction range of 10–200 cm are presented in table 17, together with the ratio a/b and values of k_{sat} determined by the augerhole method. The

TABLE 17. Relationship between k and S for plots 1, 2, 3 and 5 expressed in the form of equation (17).

Plot	Soil	$a/(S^n + b)$ (cm day^{-1})	a/b (cm day^{-1})	k_{sat} (cm day^{-1})
1	loam	$490000/(S^3 + 13000)$	37.7	75.0
2	loamy sand on light clay	$450/(S^{1.5} + 17)$	26.5	27.4
3	clay	$560/(S^2 + 80)$	7.0	8.1
5	clay	$400/(S^2 + 100)$	4.0	6.0

relationships presented for plots 2 (Banna sand) and 3 (Camarooka clay loam) are valid over the whole suction range investigated (0–200 cm) and are nearly indistinguishable from the smooth lines drawn through the data in figure 13. The relationship for plot 1 (Yandera loam) does not fit the experimental data well in the suction range 0–10 cm, as may be seen from a comparison of the ratio a/b and k_{sat}. The relation $k = 311250/(S^{2.9} + 4150)$ gives a value of $a/b = 75$, equal to k_{sat}, but overestimates capillary conductivity at low suctions (0–30 cm). For plot 5, the relation $k = 400/(S^2 + 100)$ is a reasonable approximation, capillary conductivity is slightly underestimated between $S = 0$–10 cm and slightly overestimated at $S > 150$ cm. A relation which fits the data better in the higher suction range is $k = 780/(S^{2.15} + 235)$. It is noticed that a small change in the value of n changes the values of a and b rather drastically. The data of plot 4 can not be represented by a relation of this type due to the sharp decrease of k in the low suction range (fig. 13).

Comparing our data with those of GARDNER (1958) and WIND (1955) it is observed that the relationship proposed by GARDNER, when plotted on log-log paper is approximately horizontal at low suction values (e.g. GARDNER and FIREMAN, 1958), the length and inclination of the nearly horizontal part of the curve depending on the magnitude of the constants a, b and n. At higher suctions the relationship becomes linear, the slope of the straight line giving, very nearly, the value of n. The relationship proposed by WIND (1955) and VISSER

(1959) gives a linear relationship over the whole suction range. The curves of figure 13 plotted in this manner in figure 14, are intermediate between these relationships. From the slopes of the straight portion of these curves values of n are found to be: 3.0, 1.5, 2.0, 1.9 (average) and 2.1 for the soils on plots 1–5 respectively.

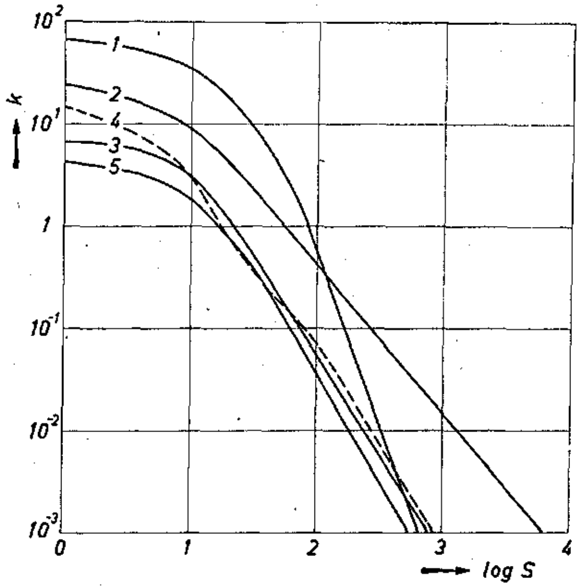


FIG. 14.
Relationships k — S, plots 1–5, on log-log paper.

A prerequisite for the GARDNER type of equation to be valid is a small suction range near saturation where k does not alter appreciably with S. This condition is often met in artificially packed soil columns as used by GARDNER and FIREMAN (1958) and NIELSEN and BIGGAR (1961) but is not always present in natural soil, where large cavities and root channels may be present which drain immediately suction is applied. In the present study for example GARDNER's relation holds strictly for plots 2 and 3 only. A study of NIELSEN *et al.* (1960) on undisturbed samples of two Iowa loess soils and two glacial till soils indicates that an equation of the GARDNER type can be fitted to the loess soils but not to the glacial till soils. The data on Pachappa fine sandy loam (GARDNER and FIREMAN, 1958) agree with field data of RICHARDS *et al.* (1956) for the same soil, the relationship found by the former authors therefore applies to the field soil also. Thus the condition of slow decrease of k in the low suction range appears to apply to about half the field data described here. From the rest of these data (plots 1, 4 and 5, glacial till soils) it can be seen that k decreases rapidly at low suctions and relationships intermediate between those of GARDNER and WIND are found.

Solutions of equation (16) are presented in figure 15 for values of q between 0.05–1.0 cm day^{-1} and values of S up to 15 atm. This equation was solved numerically for plots 1, 4 and 5 and for plots 2 and 3 with the aid of equation (13) resp. (14) of GARDNER (1958). The solutions are shown in the form of suction distribution above the watertable at various flux rates, the

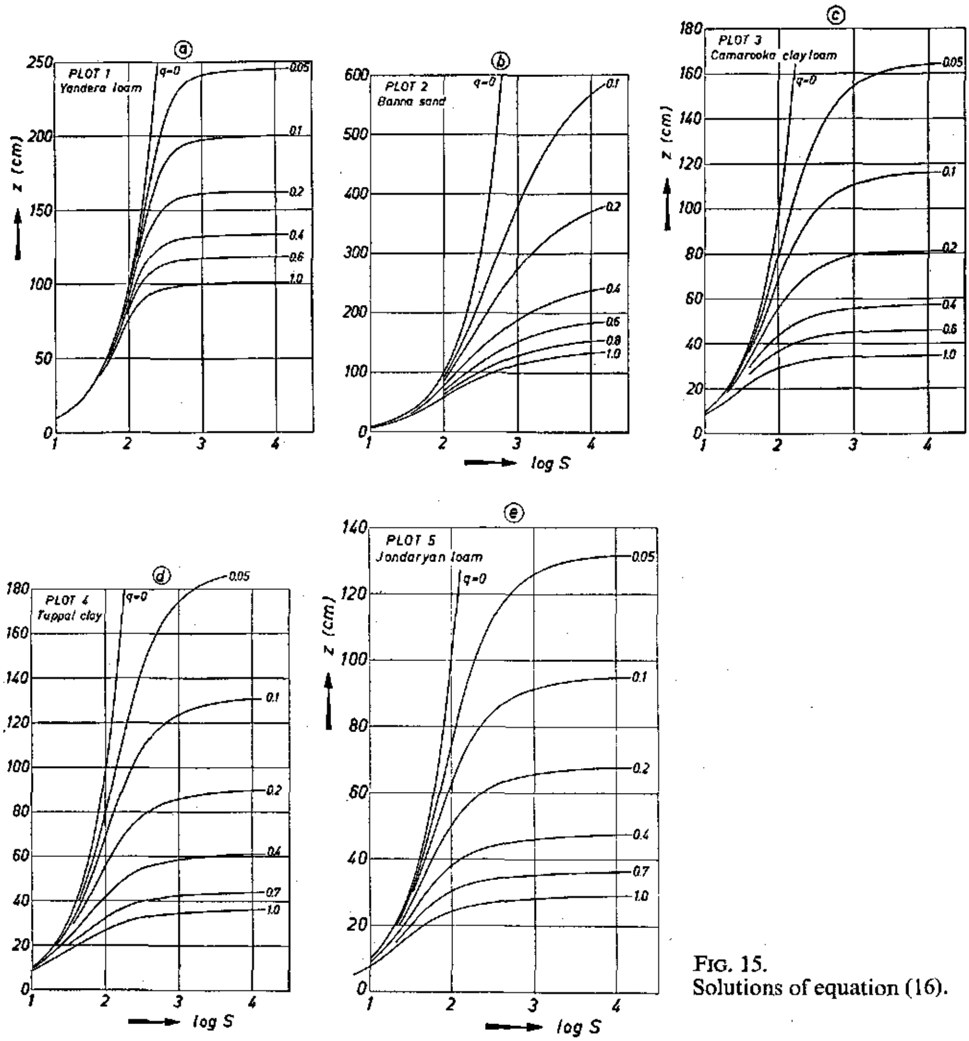


FIG. 15.
Solutions of equation (16).

solution for $q = 0$ represents the suction distribution above the watertable in the absence of evaporation. Noting the difference in scales it is seen that *e.g.* a flux rate of 0.1 cm day^{-1} can be maintained, if the suction at the soil surface is 15 atm., when there is a permanent watertable at 200 cm at plot 1, 585 cm at plot 2, 116 cm at plot 3, 131 cm at plot 4 and 95 cm at plot 5.

Solving equation (16) for suction values to 15 atm. implies that the capillary conductivity – suction relationships (fig. 13, 14, table 17) are valid over the range of moisture contents up to wilting point. A few data available on plots 1, 4 and 5 indicate that these relationships may be extended to suction values of about 500 cm but no data are available to check the validity of this assumption at higher suctions. This objection is, however, not as serious as it might appear to be, as may be seen from figure 16, which shows the relationship of flux rate

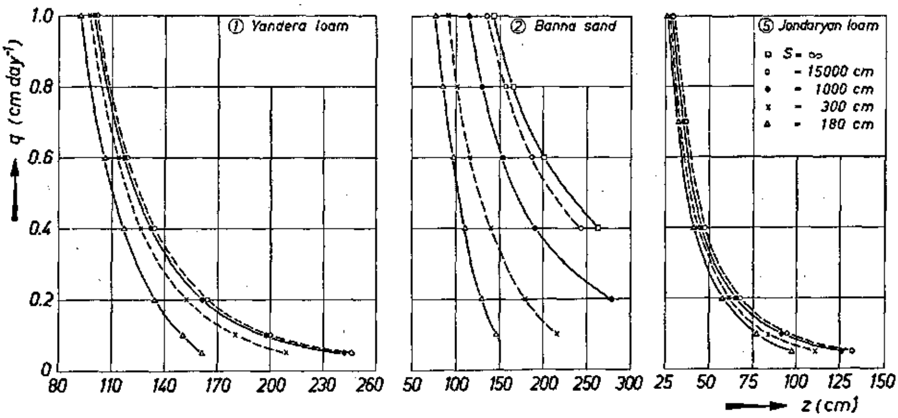


FIG. 16. Relationships of flux *(q)* and watertable depth *(z)* at various values of suction *(S)* at the soil surface.

and watertable depth at various values of suction at the soil surface, for plots 1, 2 and 5. These relationships are not shown for plots 3 and 4, they are very similar to plot 5. It is seen that the curves giving the relation between flux and watertable depth (for plots 1 and 5, and thus also for plots 3 and 4) at suctions S = 15000 cm and 1000 cm nearly coincide and even that the curve for S = 300 cm gives only a slightly different relationship. In other words, the limiting flux rate possible through the soil profile is already approached while suction at the surface remains relatively low. The exact relation between S and k at high suction values is therefore not very important. GARDNER (1958) reached a similar conclusion. An exception is the soil at plot 2, where k decreases relatively slowly with increasing suction, S. Here it is desirable to know the exact relationship over a larger range of suction, preferably to about 2000 cm.

Flow resistance and therefore loss of potential is low close to the watertable. This is easily seen from figure 15, where solutions at various flux rates nearly coincide with the curve for $q = 0$ at pF = 1. For instance, the drop in potential in profile 3 (Camarooka clay loam) at 40 cm above the watertable is 6 cm at $q = 0.1$ and 20 cm at $q = 0.3$ cm day^{-1}. It may therefore be concluded that, similarly to the capillary conductivity-suction relationship at high suction values, a precise knowledge of this relation near saturation is not very important. Thus, the approximate relations given in table 17 for plots 1 and 5 yield solutions of equation (16) that do not deviate appreciably (in the range q = 0.05–1.0 cm day^{-1}) from those given in figure 15. For instance, at a flow rate of 0.1 cm day^{-1}, the approximate relations give for S = 15000 cm at the surface a watertable depth of 205 cm for plot 1 and 98 cm at plot 5, compared to 200 and 95 cm found from the numerical solutions presented in figure 15. Similarly, no serious objection can be made to fit an equation of the type proposed by VISSER (1959) to the capillary conductivity-suction relationships.

b. *Meteorological conditions*

In the steady state case the flux, q, through the soil profile equals the evaporation rate, E, from the soil surface. Considering the influence of meteorological conditions on the evaporation rate and therefore on the flux through the soil, it is observed that the evaporation rate, E_s from nearly saturated soil is an adequate specification for our purpose of the meteorological conditions. Three possible cases can then be considered.

Firstly, when the watertable is reasonably close to the surface, we have in general the condition that evaporation from the soil is not limited by the water transmitting properties of the soil, or $E_s < E(z_w)$, where $E(z_w)$ is the maximum flux rate with the watertable at $z = z_w$. In this case $E = E_s$, in other words the flux through the soil is controlled solely by meteorological conditions. Secondly, when $E_s = E(z_w)$, the flux through the soil profile is just sufficient to maintain the evaporation rate $E = E_s$, the only difference with the first case being a different suction (or moisture) distribution in the profile above the watertable. Thirdly, when E_s exceeds $E(z_w)$ the moisture transmitting properties of the soil become the limiting factor in evaporation and now $E = E(z_w)$. The moisture or suction distribution in the soil remains virtually the same as in the previous case, apart from a very small increase in the dry zone at the soil surface through which moisture is transmitted partly in the vapour phase (PHILIP, 1957).

To illustrate the various stages consider the soil profile of plot 1 during a period when E_s, the potential evaporation rate given by meteorological conditions is 0.4 cm day^{-1}. Then if the watertable is at 100 cm it is seen (figures 15 and 16) that $E(z = 100) = 1.06$ cm day^{-1} and therefore $E = E_s = 0.4$ cm day^{-1}. The suction at the soil surface, $S_0 = 123$ cm. At a watertable depth of 119 cm, $E(z = 119) = 0.60$ cm day^{-1}, $S_0 = 206$ cm and at a depth of 134 cm $E = E(z = 134) = E_s = 0.4$ cm day^{-1}, $S_0 = 15$ atm. With a further lowering of the watertable, to *e.g.* 200 cm we find that $E = E(z = 200) = 0.1$ cm day^{-1}.

The analysis of PHILIP (1957) further suggests that the dependence of E on the heat flux entering or leaving the soil is especially of importance when $E_s > E(z_w)$, a condition applying in the present study in the summer period when E_s is large and the watertable relatively deep. It was found that a downward heat flux inhibited evaporation more than an upward one increased it. Thus a net flux $E < E(z_w)$ results. This point is of interest in examining the data of table 15.

In the preceeding paragraphs it has been tacitly assumed that the soil surface is bare and E_s therefore relates to this condition. The relation between E_s and the potential evaporation E_0 (section II-2-b) is given by PENMAN (1948) as $E_s = 0.9\ E_0$ but for Phoenix, Arizona, FRITSCHEN and VAN BAVEL found a relation $E_s = 1.18\ E_0$. For the present purpose it appears reasonable to equate E_s with E_0. Where a site supports vegetation, in this study specifically plots 3 and 5, E_s is identified as E_t in the case of a well watered vegetation (low suction values in the soil) and, as the soil in the root zone progressively dries out, E_t is replaced by appropriate values of E_a as *e.g.* given in table 8. Evaporation in this case is not confined to the soil surface but occurs over the whole depth of the root zone. Values of suction, appropriate to upward movement of moisture from the watertable are now taken at the lower root zone.

c. *Movement in the vapour phase*

So far movement of water in the liquid phase only has been considered but the contribution of flow in the vapour phase is of interest when the maximum flux through the soil profile is equal to or smaller than the evaporative demand, in other words $E_s \geq E(z_w)$. In the present study movement of water in the liquid phase has been considered possible to moisture contents corresponding to 15 atm. suction. The assumption underlying GARDNER's work is that maximum flux in the liquid phase occurs by specifying $S = \infty$ at the soil surface. PHILIP (1957) concludes from an analysis of the influence of relative humidity, h_0, at the soil surface on the flux through a profile of Yolo light clay that the flux depends strongly on h_0 only when $h_0 \leq 0.99$, this value corresponding to a suction of 14000 cm. At values of h_0 between 0.99–1, corresponding to suctions 0–14000 cm, vapour pressure gradients are so low that flow in the vapour phase may be neglected. When S increases beyond 14000 cm flow in the vapour phase soon becomes dominant and occurs through a small "vapour bottleneck" created at the soil surface, the length of this zone increasing as the difference between E_s and $E(z_w)$ increases.

GARDNER (1958) studied, in a similar but more simplified manner, the influence of vapour movement through this small region on the maximum flux $E(z_w)$ calculated for flow in the liquid phase only. He found that $E(z_w)$ increased by a fraction equal to: $nD_v\,(p_1 - p_2)/z_w$, where n = as in equation (17), D_v = vapour diffusion coefficient, p_1 = saturation vapour pressure of water and p_2 = vapour pressure at the soil surface. From data, presented in this study and by PHILIP (1957), this correction was found to amount to some 5–10%.

The reduction of evaporation by a surface mulch is well known. Defining for this purpose a mulch as a medium which conducts moisture in the vapour phase only, one has from simple diffusion theory the relation $E = D_m\,(p_1 - p_2)/L$, where E, p_1 and p_2 are as previously defined, D_m = vapour diffusion coefficient for the mulch and L = mulch thickness. The reduction in evaporation rate predicted by this expression is proportional to the thickness of the mulch, a result which has been confirmed by several experiments (see *e.g.* WIEGAND and TAYLOR, 1961). At small values of L, the formal relationship given above is invalid and the energy available for evaporation rather than the mulch thickness becomes limiting. The minimum mulch thickness must be of the order of a few millimeters to reduce evaporation through the entry of vapour diffusion as the limiting process. From the review of WIEGAND and TAYLOR (1961) it appears that the necessary mulch thickness increases with the coarseness of the mulch material, a fact which is also evident from results presented by HOLMES *et al.* (1960) on evaporation rates from soil with different tilth conditions at the surface.

An important conclusion may be drawn from a comparison of data relating to evaporation control by means of mulches (introduction of a vapour barrier) as against limiting flow in the liquid phase by lowering of the watertable. In both cases the curves of evaporation rate versus depth of watertable (fig. 16) or depth of mulch are of similar shape, that is hyperbolic on linear coordinates. However, the results of GARDNER and FIREMAN (1958) on Pachappa fine sandy loam show that the depth scale of the mulch is only about 1/100 that of the watertable depth scale, which means that a small increase in depth of a surface mulch has as much influence on the evaporation rate as a large increase in watertable depth. The shallow cultivation on plot 1 during October, 1960, resulted in a mulched surface and the effect of this mulch on the capillary flow rate will be ascertained from the relevant data on this plot. This is done in the following section.

d. *Analysis of the data of table 15*

The data of table 15, where capillary flow rates, together with potential evaporation, rainfall, watertable depth and suction in the top layers, are shown for plots 1–5, may now be compared with the analysis given in the preceeding sections (4-a, b and c). Bearing in mind that individual flow rates may be appreciably in error due to the variability of the field data, it appears reasonable to combine these results for periods during which conditions were about the same and to draw conclusions on the basis of the overall data rather than for individual entries.

Taking for instance the period May-Sept. 1960, for the, during that time, practically bare plots 1, 2 and 4 it is seen that the potential evaporation rate was fairly low, whilst the average watertable generally was sufficiently high to have the conditions: $E_s = E_0$ (bare plots) and $E_s < E(z_w)$, thus the soil physical properties are not limiting the flow rates. The data for plot 1 during May and June are an exception, here $E_0 = 1.6–1.9$ mm day^{-1} and $E(z_w) = 1.2$ mm/day^{-1} (figure 16, plot 1, watertable = 190 cm). Capillary flow rates at these plots during this period are seen to be about equal to or smaller than E_0. Rainfall during this period has sometimes resulted in leaching, *e.g.* during July, when rain was fairly evenly distributed over the month with an intensity of about 2–3 mm day^{-1}, thus on the average higher than E_0 (= 1.2 mm day^{-1}). The average suction distribution above the watertable should therefore result in the establishment of downward potential gradients. This is seen to be the case, for example the equilibrium suction at plot 1, 15 cm depth would be 174 — 15 = 159 cm, whereas the average suction was 140 cm.

Rain during June was less than the potential evaporation and watertable depth and moisture content at the beginning and end of this month was about the same (see *e.g.* figure 12). Considering therefore that all the rain evaporated again after entering the soil profile, it is seen that the evaporation rate at *e.g.* plot 2 has been about 0.9 + 0.6 = 1.5 mm day^{-1}, very close to the value of $E_0 = 1.6$ mm day^{-1}. During August and September considerable rain fell, this was partly stored in the soil profile as may be seen from a rise in the average watertable, some runoff occured on plots 1 and 4, especially after a heavy fall in September (see fig. 12) and the remainder evaporated. This quantity is difficult to establish, but when added to the capillary flow rates for these months, results in evaporation rates about equal to (plot 2) or exceeding (plots 1 and 4) the potential evaporation rates. However, these differences generally appear to be within the error limits (table 9) of about 30–40% due to variability of chloride and 10–15% due to variability of moisture content. It may be concluded therefore that during the period May–Sept. 1960, the evaporation rate at these plots has been essentially the same as the potential evaporation rate, E_0.

Potential evaporation increased sharply during the dry month of October, 1960, with the result that in this period the condition $E_s < E(z_w)$ gradually changed to $E_s > E(z_w)$. At plot 2, $E_s < E(z_w)$ still applied, but it is seen that here also the capillary flow rate, which represents evaporation from the plot since rainfall was negligible, is much less than E_0. This result can not be interpreted in terms of the analysis given in the preceeding sections, reduction in evaporation here is probably due to the formation of a salt crust at the surface which, according to data of MOROZOV and VERNIKOVSKAYA (1954) reduces evaporation considerably. The capillary flow rate at plot 4 during October is about equal to $E(z_w)$ and, though no great reliability can be placed upon a single observation, is therefore in agreement with the theory.

The measurements at plot 1, during October, will be analysed in more detail.

From late September onwards weed growth was vigourous on the plot. On the 18th October this vegetation was removed by shallow cultivation to about 5 cm depth, resulting in a rather dry cultivated top soil. The effect of this cultivation on the suction distribution in the subsoil is shown in figure 17, which shows the suction recorded by tensiometers at 30.5, 61 at 90.5 cm below the soil surface. Suction recorded at 15 cm is not shown, since the tensiometer at this depth

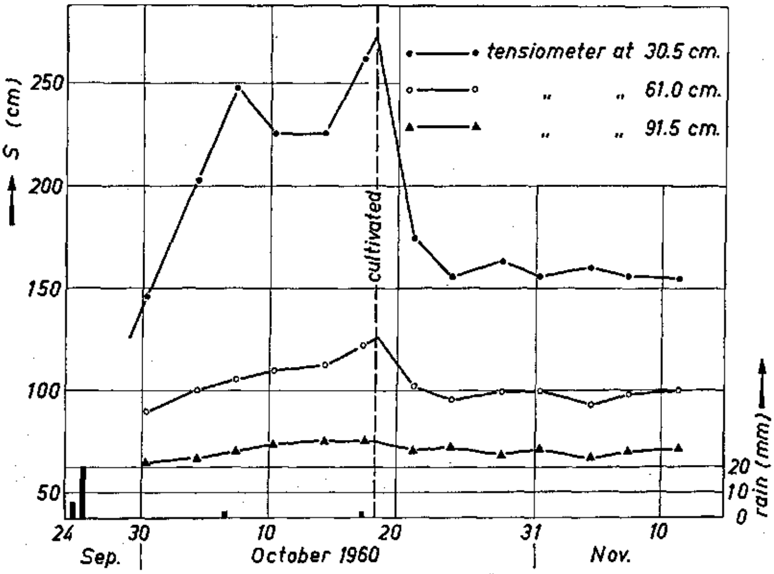


FIG. 17. Effect of shallow cultivation on suction in the subsoil, plot 1, period Oct. 1960.

broke down on 31 Oct. Suction recorded by this tensiometer on the day before cultivation was 472 cm, after cultivation 180 cm. Steady state flow rates deduced from the suction distribution in the soil profile prior to ($z_w = 157$, average S at 30 cm $= 250$ cm) and after cultivation ($z_w = 157$, S at 30 cm $= 160$ cm) are about 0.4 and 0.2 cm day^{-1} respectively, thus the flow rate is reduced by a factor 2 due to cultivation. However, this result may not be simply interpreted as the effect of a dry surface mulch of about 5 cm depth, since prior to cultivation the plot supported vegetation and moisture withdrawal took place over the whole depth of the root zone, estimated to be about 30 cm (depth of the A-horizon) while after cultivation evaporation took place just below the surface mulch. At this depth (152 cm above the watertable) the maximum transport in the liquid phase, $E(z_w) = 0.27$ cm day^{-1} and, if additional transport in the vapour phase of 5% is assumed (see section c) $E(z_w) = 0.284$ cm day^{-1}. The effect of cultivation is therefore seen to be twofold, firstly the site of evaporation in the soil is changed, reducing evaporation from about 0.40 to about 0.28 cm day^{-1} and secondly the effect of the mulch itself, reducing evaporation further to about 0.20 cm day^{-1}. The influence of this type of mulch, although still important for practical purposes, is by no means as pronounced as the effect of the more ideal mulches (separated from the soil by a screen) used under laboratory conditions. Cultivation establishes a mulch placed directly on the soil and thus liquid movement is still possible, especially if the soil is rather wet. This was found to be the case on this plot in January when cultivation, a few days after a 37 mm rainfall had no appreciable effect on the suction distribution in the soil profile.

During the mid-summer period (Jan.-Feb., 1961) the condition $E_0 > E(z_w)$ applied to the three bare plots 1, 2 and 4. Actual evaporation, E, calculated from the monthly salt and water balance at these plots (equation 14), together with values of E_0, $E(z_w)$ and suction at 5 cm depth are shown in table 18. The suction

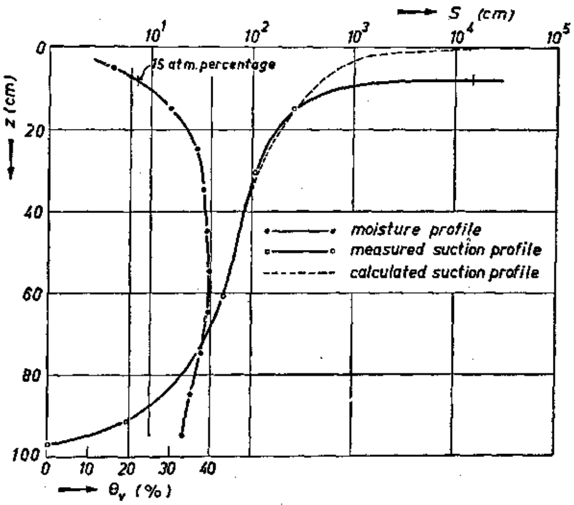
TABLE 18. Values of E_0, $E(z_w)$ and E, plots 1, 2 and 4, during the summer, 1961.

Plot	Period	E_0 (mm day^{-1})	z_w (cm)	$E(z_w)$ (mm day^{-1})	E (mm day^{-1})	S at 5 cm (atm.)
1	Jan.	7.9	170	1.75	1.3	2.5
2	Jan-Feb.	7.9	188	6.6	3.5	0.3
4	Jan-Feb.	7.9	98	1.7	0.4	> 15

values have been deduced from moisture content measurements of the 0–10 cm layer and the moisture characteristic of the 10–20 cm layer, they are therefore approximate only. It is seen that the actual evaporation, E, from all plots has been lower than the predicted value $E(z_w)$. Considering the probable causes contributing to this result it is observed that the values of $E(z_w)$ would appear to be reasonably accurate. Taking vapour movement into consideration would increase the value of $E(z_w)$ by 5–10% (see section II-4-c), but the effect of the diurnal heat flux results in a net decrease of $E(z_w)$. Net shortwave radiation, R_s, for the period in question is about 10^{-2} cal cm^{-2} sec^{-1} (FLEMING, 1962) during day-time. At this value of R_s and values of z_w and E_0 appropriate to our conditions, the reduction in $E(z_w)$ given by PHILIP (1957) is of the order of 10–20%. It is not possible to estimate the increase in $E(z_w)$ during the night but it may be concluded that the total correction for $E(z_w)$, due to vapour movement and heat flux, should be reasonably small.

The reduction in evaporation at plots 1 and 2 is probably due to the formation of a salt crust (MOROZOV and VERNIKOVSKAYA, 1954) which, after being largely removed from these plots by rainfall during November and December, was formed again during January. This encrustation was more of a patchy nature on plot 1, where the reduction in evaporation was found to be less than at plot 2. Suction values at 5 cm depth are low, indicating that formation of a natural mulch did not take place to an appreciable extent, if at all. In contrast to this is the soil at plot 4. Here salt encrustation was much less pronounced, as may be seen from figure 11, but the soil, being of clay texture to the surface, showed numerous shrinkage cracks, thus facilitating the establishment of a natural mulch. The approximate depth of this mulch is seen in figure 18, which shows the average moisture and suction distribution in the soil of plot 4. The moisture percentages shown are the average of moisture samplings on 4 Jan., 2 Feb. and 7 March. The measured suction profile represents the average of suction values recorded during January and February by the tensiometers at 15, 30.5, 61 and 91.5 cm depth, above 15 cm this line was extended with the aid of the moisture content curve, to 15 atm. The calculated suction profile, corresponding to $E(z_w) = 0.17$ cm day^{-1}, $z_w = 98$ cm, is shown for comparison.

FIG. 18.
Moisture and suction profiles, plot 4, period Jan.–Feb., 1961.



If the mulch depth is taken to correspond to that part of the profile where $S > 15$ atm., this self mulching effect is seen to extend to about 8 cm depth. The reduction in evaporation is equivalent to lowering the watertable to 210 cm ($E(z_w) = 0.04$ cm day^{-1} for $z_w = 210$ cm). Therefore, if the whole of the reduction in evaporation may be attributed to the mulch, this mulch has had the same effect as would be achieved by a lowering of the watertable from 98–210 cm.

The results at plots 3 and 5 will not be discussed. Conditions at plot 3 generally have not been favourable for a comparison with steady state theory, vegetation in the form of weed growth was regularly removed by cultivation, whilst the plot was frequently irrigated during the spring, summer and autumn period. The overall effect of irrigation on salinization has been discussed in section II-3-e. Plot 5 supported a pasture vegetation for the whole of the period of observation and, whilst values of E_a are available and could be used to specify the evaporative demand, these have been obtained from the same set of data as are those of table 15. Thus no independent means of checking the results is available.

Summarizing the discussion presented above it may be concluded that the agreement between the field data and the theory of steady state flow in the liquid phase is satisfactory when no limit is set on evaporation by the moisture conduction properties of the soil profile. Where the potential evaporation rate exceeded the maximum flux rate possible through the soil profile, evaporation from the soil has been equal to, or lower than, the predicted maximum rate through the profile. Under high evaporative conditions reduction in evaporation was caused either by the establishment of a natural mulch or by the appearance of a salt crust at the soil surface. The effect of removing an actively growing vegetation by cultivation is twofold, the site of evaporation in the profile moves in upward direction and the tilled layer acts as a mulch. This mulch may not be very effective when the soil surface is still in a moist condition. Establishment of an ideal much, which conducts water in the vapour phase only, is difficult to achieve under field conditions.

III. ASSESSMENT OF THE SALINITY HAZARD

The data presented in the previous sections enable an assessment to be made of potential salinization as determined by soil physical properties, climatic conditions, groundwater salinity and depth of watertable. The concept of critical depth of watertable will be examined by considering the effect of both soil physical and climatic conditions, assuming groundwater salinity to be constant. Finally the effect of groundwater salinity will be shown by way of some examples of salt accumulation, which will also illustrate the principles involved and enable some remarks to be made on possible procedures of salt control.

1. Critical depth of watertable

From the data and discussion in sections II-4-b and d it is apparent that for any given soil profile the depth of the watertable, z_w, where the maximum flux rate through the soil, $E(z_w)$, equals the evaporation rate, E_s, is an important depth. Since it was shown (section II-4-c) that the main movement was in the liquid phase we then have a maximum upward movement of salt dissolved in the water. If the watertable is higher than z_w the flux rate is equal to this maximum rate and at watertable depths exceeding z_w the flux rate E drops below E_s to the value of $E(z_w^*)$ corresponding to the appropriate watertable depth, z_w^*. This is illustrated in figure 19 and by the data of table 19.

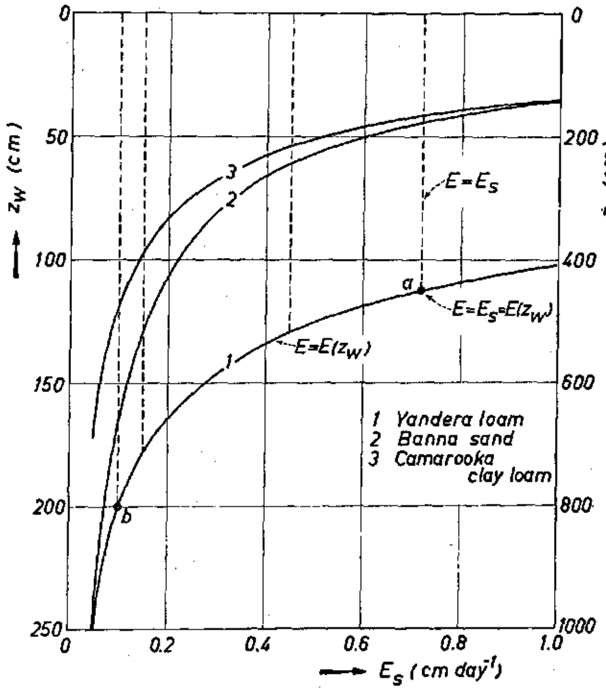
Table 19. Values of z_w (in cm) for the soils of plots 1–5, corresponding to seasonal evaporation rates.

Period	E_0 (cm day^{-1})	Yandera loam	Banna sand	Camarooka clay loam	Tuppal clay	Jondaryan loam
	0.10	201	660	117	132	95
Winter	0.15	178	504	96	106	78
Spring	0.45	128	241	55	58	45
Summer	0.72	112	178	43	44	35
Autumn	0.35	139	286	63	67	52

Table 19 gives values of z_w for the soils studied here and for flux rates corresponding to seasonal potential evaporation rates found under average climatic conditions of the Murrumbidgee Irrigation Areas. Values of z_w for $E_s = 0.1$ cm day^{-1} are also shown in this table. Figure 19 shows maximum flux rates, $E(z_w)$, in relation to watertable depth for three different soil types. The vertical broken lines indicate the average summer, spring and winter evaporation rates and the evaporation rate of 0.1 cm day^{-1}, as given in table 19. Values of z_w plotted along the ordinate on the left in figure 19 apply to curves 1 and 3, those along the ordinate on the right apply to curve 2. The values of z_w at the intercepts of the $E(z_w)$ curves and the vertical broken lines, *e.g.* those marked by *a* and *b* on curve 1, correspond to values of z_w given in table 19.

Considering now the potential salinization of the soil on plot 1, Yandera loam, under average climatic conditions of the Murrumbidgee Irrigation Areas, it is seen that during the summer period salinization will proceed at the maximum rate as long as the watertable remains at 112 cm (point *a* in figure 19) or higher. If in summer the watertable is lowered to 134 cm the evaporation rate will drop to 0.4 cm day^{-1} and salinization will occur at a rate of 0.56 × the maximum

FIG. 19.
Maximum flux rates in relation to watertable depth for the soils on plots 1–3.



rate as determined by meteorological conditions. A decrease of evaporation rate to 0.2 cm day^{-1} is achieved by lowering the watertable to 164 cm, and to 0.1 cm day^{-1} by lowering the watertable to 201 cm. Further reduction in evaporation and thus salt transport is achieved only by lowering the watertable to a much greater depth, as may be seen from the shape of the $E(z_w)$ curve at values of $E_s < 0.1$ cm day^{-1}. Considering the spring period, it is seen that the maximum flux rate in this period is maintained as long as the watertable is above 130 cm, at greater depths the flux rate is reduced in exactly the same way as above and is again determined by the shape of the $E(z_w)$-curve.

Curves 1, 2 and 3 (figure 19) are very similar in shape, as are those of Tuppal clay (plot 4), Jondaryan loam (plot 5, see fig. 16), the river basin clay (fig. 2) of WIND (1955) and (see also fig. 2) Pachappa sandy loam (GARDNER and FIREMAN, 1958). Therefore, for all field soils investigated up to date, it is seen that a relatively small increase of the watertable depth, z_w, results in a large reduction of the flux rate down to values of about 0.2 cm day^{-1}. At values of E_s between 0.2 and 0.05 cm day^{-1} a greater depth increase is necessary to lower the flux rate and at $E_s < 0.05$ cm day^{-1} a further lowering of the watertable affects the flux rate very little. The critical range on all these curves is found at watertable depths corresponding to evaporation rates of about 0.1 cm day^{-1}, thus providing a physical basis for the conclusion drawn from experimental results in the USSR (see section I-3-b).

However, it is also seen from figure 19 and table 19 that the evaporation rate of 0.1 cm day^{-1} is smaller than the lowest average potential evaporation rate found in the winter period. Thus if the watertable is lowered to the depth, z_w, corresponding to $E(z_w) = 0.1$ cm day^{-1}, it is found that the condition $E(z_w) <$

E_s generally applies during the whole year. This facilitates, on bare soil, the establishment, *e.g.* by suitable tillage operations, of an artifical mulch which over a period of time acts as a vapour barrier and thus reduces the danger of salinization. The length of time over which the mulch is effective depends on a number of factors, such as the velocity of the advancing wetting front from below, the rainfall distribution and properties of the mulch material (*e.g.* the consolidation-time relationship). It is observed that, where these factors are favourable and thus an effective mulch is possible, a higher watertable may be tolerated during periods of high potential evaporation and values of z_w corresponding to $E(z_w) = E_s$ for the periods concerned may be regarded as critical depths.

It is concluded, when salinity control is achieved by limiting flow in the liquid phase, that the critical depth corresponds to the watertable depth at which the flow rate through the soil profile is reduced to values of about 0.1 cm day^{-1}. The reason is seen to be mainly a soil physical one and the conclusion rests on the observation that at values < 0.1 cm day^{-1} a great decrease of the watertable depth is necessary to change the flow rate appreciably. In the case of evaporation control under conditions of fallow a further reason is that the flow rate of 0.1 cm day^{-1} is sufficiently small to permit under favourable circumstances the maintenance of a mulch at any desired period of the year in most irrigated areas of the world. In practice the latter method is widely applied (see *e.g.* KRIMGOLD, 1945, and MARSHALL, 1959, p. 2–3). Were salinization is reduced by the establishment of a vapour barrier one has the, seemingly paradoxical, conclusion that saline groundwater may be tolerated at a higher level if potential evaporation is high.

2. ESTIMATION OF POTENTIAL SALINIZATION FROM SOIL PHYSICAL PROPERTIES

The importance of the soil physical properties in determining the critical depth of watertable introduces the question whether various soils can be ranked, for the purpose of estimating potential salinization, on the basis of some easily determinable properties. For example, the data discussed in section I-3-c (figure 3) suggest a possible relationship between salinization and hydraulic conductivity. For this purpose, the critical depth of watertable z_w, for $E(z_w) =$ 0.1 cm day^{-1}, used as an index of potential salinization, is shown in table 20 together with the hydraulic conductivity, k_{sat}, and the constants a, b and n of equation (17) for various soil types.

It is seen from these data that the hydraulic conductivity alone is not sufficient to indicate the salinity potential of a soil profile, *e.g.* the sand profile of WIND (1961) and the Banna soil of the present investigation have similar values of k_{sat} but the sand is seen to be the safest, whilst the Banna soil is the most dangerous of all soils listed in table 20. Capillary conductivity decreases very rapidly as suction increases in the sand ($n = 4$) but this decrease is slow in the case of Banna ($n = 1.5$). Therefore an improved estimate of potential salinization is obtained if both k_{sat} and n are known. However, even this is not sufficient, as may be seen by comparing Yandera loam and Pachappa sandy loam. The values of z_w and n are not very different for these soils, but k_{sat} of the Yandera loam is considerably higher. Capillary conductivity decreases rapidly at low suction values in the Yandera loam soil but drops only after a certain suction value is exceeded in the Pachappa sandy loam. The same is found when com-

TABLE 20. Critical depth of watertable in relation to soil physical properties.

Soil type	k_{sat}	a	b	n	z_w for $E(z_w) = 0.1$ cm day^{-1}
	(cm day^{-1})	(cm^{n+1} day^{-1})	(cm^{n+1})	–	(cm)
Yolo*	1.0	400	400	2	99
Pachappa*	12.3	32.10^4	26.10^3	3	178
Chino*	2.8	1600	565	2	198
basin clay+	∾5.0	–	–	1.5	95
sand++	40.0	4.10^6	10^5	4	88
Yandera	75.0	–	–	∾3	201
Banna	27.4	450	17	1.5	660
Camarooka	8.1	560	80	2	117
Tuppal	16.9	–	–	∾1.9	132
Jondaryan	6.0	–	–	∾2.1	95

* From data of GARDNER and FIREMAN (1958).
\+ From data of WIND (1955).
++ From data of WIND (1961).

paring *e.g.* Camarooka clay loam and Jondaryan loam with Yolo light clay, here z_w, a and n are about the same but values of k_{sat} and b differ considerably.

Potential salinization is therefore not easily derived from a single physical property but depends on the magnitude of all factors listed in table 20. Although these may; in principle, be obtained from an analysis of relatively simple laboratory measurements of capillary conductivity, it was shown in section II-4-a that if these laboratory measurements represent field conditions it may not be possible to express the results in the form of equation (17). It is however simple enough to state the requirement, namely a knowledge of the relation between conductivity and suction, for soils with values of $n > 2$ in the suction range 0–500 cm and, when $n < 2$ over a somewhat larger range of suctions (see section II-4-a).

The salinity hazard on the fine textured clay soils Camarooka, Tuppal and Jondaryan is not very great, the critical depths (tables 19 and 20) are considerably smaller than for soils of intermediate texture. It also appears that coarse textured soils, in table 20 represented by the sand profile of WIND (1961), are not very liable to salt accumulation. These results are in agreement with the conclusion drawn in section I-3-c. It is finally pointed out that the critical depths, derived for the soils studied here, are all within the depth range in which the relation $k = f(S)$ was studied, with the exception of Banna sand. It is observed that the deep subsoil of this soil type, from 200 cm downwards, has a very low permeability, therefore watertable control to 660 cm would in practice not be required.

3. GROUNDWATER SALINITY AND EXAMPLES OF ACTUAL SALINIZATION

In the preceeding sections an assessment was made of potential salinization of the various soils by considering maximum possible salt transport as dependent on soil physical properties and potential evaporation rate only. Actual salinization may be less, due to the effects of mulching, rainfall, irrigation and actual evapotranspiration rates by crops on the flow rate through the profile. For comparative purposes equal groundwater salinity has been assumed. It is immediately apparent however that different groundwater salinity has a pronounced effect on actual salinization and it is *e.g.* easily shown that, even when the watertable is below the critical depth, harmfull amounts of salt may accumulate provided sufficient time is allowed for this process. The effect of all these factors is shown by way of some examples, using as far as possible, actually occurring conditions.

Consider for example the profile of Camarooka clay loam, where in winter groundwater with a chloride content of 0.1% is present at a depth close to 117 cm. The maximum flow rate through the profile is then 0.1 cm day^{-1} (table 19) and under climatic conditions of the Murrumbidgee Irrigation Areas (evaporation and rainfall distribution) it is quite possible for this rate to be maintained over a 30 day period. A total amount of chloride equal to 3.10^{-3} gr cm^{-2} month^{-1} then accumulates to the surface (in the case of bare soil), but from groundwater containing 1% chloride the total amount would be 3.10^{-2} gr cm^{-2} month^{-1}. If, at the end of this period, this chloride is distributed, by cultivation or rainfall, over a layer of 20 cm depth, these amounts, expressed as a percentage of dry soil are $100 \times 3.10^{-3}/(20\,\text{cm}^3 \times 1.4) = 0.0107\%$ and 0.107% respectively. The latter concentration is sufficiently high to require some leaching before cropping can commence. Since the watertable is at the critical depth, further lowering would have little effect, *e.g.* at watertable depth of 165 cm the amount accumulating from groundwater containing 1% chloride would be 0.053%, still high enough to require some leaching. Salinity control by mulching during the dry period is possible however ($E(z_w) < E_s$) and this would be simpler than the installation of a deep expensive drainage system, if such dry periods do not occur frequently.

The salinization rate of the Jondaryan loam soil supporting a winter pasture may be assessed from the data on plot 5 (see tables 4, 8, 19 and figures 7, 16). Taking the roots to be mainly concentrated in the top 20 cm and available water in the root zone maintained at 90%, we have during the months June and July the conditions: $E = E_t = 0.76\ E_0 \cong 0.1$ cm day^{-1} (table 8), suction at the lower root zone (fig. 7) corresponding to 90% available water is about 125 cm. The flow rate from an assumed static watertable at 100 cm depth to the lower root zone at 20 cm depth is 0.06 cm day^{-1} under these conditions (fig. 16c). If available water in the root zone is 80%, then E remains at 0.1 cm day^{-1}, suction at the lower root zone is about 230 cm and the flow rate is about 0.1 cm day^{-1}, that is equal to the evapotranspiration rate. Further drying of the root zone is therefore not possible in this case.

It is seen that the salinization rate under these conditions is reduced by maintaining a high moisture content in the root zone and that the maximum salinization rate occurs when available moisture in the root zone is still relatively high. Salinization therefore can only be reduced by maintaining a very high moisture content in the root zone. If this soil was bare, then the maximum flow rate from a watertable at 100 cm would be 0.09 cm day^{-1} (fig. 16), slightly lower than for the pasture vegetation.

Table 21 shows the amount of chloride accumulated in the root zone of citrus, during the interval between 2 successive irrigations in summer, in relation to available water percentage. Data relating to the moisture characteristic and thickness of the main root zone, as well as the relation between actual and potential evaporation, have been taken from section II-2-c (tables 7, 8 and fig. 7), whilst the moisture conducting properties of the subsoil are assumed to be identical to the Yandera loam (fig. 16a). In the example the watertable is assumed to be at 226 cm, the groundwater containing 0.3% chloride. The results shown are approximate, in so far that the moisture extraction by roots is assumed to take place at a constant depth of 25 cm and is assumed to occur in successive stages during which flow is in the steady state. The following observations may, however, be made.

TABLE 21. Salinization of the root zone of citrus in a single irrigation period.

Available water	Average suction	Flow rate	$E_a = E_0.f.p.$	Period	Chloride in root zone
(%)	(cm)	(cm day^{-1})	(cm day^{-1})	(days)	(% dry soil)
100–75	340	0.06	0.50	2	0.00090
75–50	1050	0.095	0.30	5	0.00356
50–25	2250	0.098	0.15	20	0.01470

It is clear that the salinization rate (col. 3) increases rapidly during the initial stage of moisture extraction but approaches a limiting value when moisture content in the root zone is still quite high. Most of the chloride accumulates when the root zone is rather dry (50–25% available water); in this case actual evapotranspiration (col. 4) is low, the contribution of capillary flow to evapotranspiration is high and thus the time interval for moisture extraction from 50 to 25% available water is large (col. 5). The amount of chloride accumulated in the root zone (col. 6) is sufficiently high to require leaching during the following irrigation and provides an estimate for determining the leaching requirement (section I-2). In the present example the total contribution of capillary flow to the consumptive use of water by the citrus crop is about 45%.

SUMMARY

The requirements for adequate drainage in irrigated areas must often be based on the prevention of serious salinization of the root zone. The salinity factor is to be taken into account in designing the capacity and in determining the desirable depth of the drainage system. It is shown that the required discharge capacity may, in principle, be derived from the leaching requirement and to be dependent on the water requirements and salt tolerance of the crop and on irrigation water quality. Where saline groundwater is present the desirable watertable depth depends on the water conducting properties of the soil, on climatic factors and on the salinity of the groundwater. From a survey of literature relevant to this problem it was seen that drainage to depths of 180 cm and more is usually recommended to control the process of salinization from saline groundwater, but no sound analysis exists to support such recommendations. The dependence of the watertable depth on the above factors was therefore studied for conditions existing in the Murrumbidgee Irrigation Areas, Australia.

Natural drainage of the Murrumbidgee Irrigation Areas is generally poor due to their unfavourable physiographic position on the large inland plains of

south-east Australia, but the irrigation water is of high quality. Salinity problems that have developed are therefore caused by redistribution of salts present in the subsoil, due to rising watertables.

The climate of this area is semi-arid, the average rainfall amounts to 400 mm annually and is fairly evenly distributed over the year. Average potential evaporation, E_0, calculated from the Griffith meteorological records with the aid of the Penman formula and estimated from Australian standard pan evaporimeter records, varies from 0.14 cm day^{-1} in July to 0.78 cm day^{-1} in January. On the average potential evaporation exceeds rainfall over the whole of the year. The difference between calculated and measured monthly values of E_0 during 1961 was generally no more than 15%, therefore the Australian standard pan at Griffith estimates E_0 rather closely over periods of a month or longer. Potential evapotranspiration, E_t, of a pasture, from late autumn to early spring, was found to be 0.76 E_0 and of citrus, during the summer, 0.66 E_0. Some values of actual evapotranspiration, E_a, are given for these crops, in relation to available moisture percentage and potential evaporation.

The effect of soil physical conditions and watertable depth on salinization was studied under field conditions on five test plots. Sites were selected so as to cover the range of soil physical conditions normally encountered in the Murrumbidgee Irrigation Areas. Special emphasis was placed on fine textured soils of low permeability, mainly because of the conflicting views expressed in the literature concerning salinization and watertable control in such soils.

The process of salinization was studied by monthly samplings of moisture and chloride content. These data were used to calculate average daily capillary flow rates for each month between May, 1960–Oct. 1961. Potential gradients were calculated from suction measurements recorded by tensiometers installed at regular intervals between the soil surface and the watertable. Capillary conductivity-suction relationships were obtained by calculating capillary conductivity from average flow rates and average potential gradients where tensiometer records indicated reasonably steady state conditions. At the end of the investigation, undisturbed soil samples were collected from the plots to characterize the soil profile in terms of bulk density, porosity, particle size distribution and moisture content-suction relationships. Hydraulic conductivity was obtained by in place measurements with the augerhole method.

Due to the high clay content and relatively poor structure of the soils investigated the amount of unavailable water, held above 15 atm., was generally high. The moisture content-suction relationships were generally found to be different in the laboratory and the field due to hysteresis and air entrapment in the case of field samples and swelling phenomena in the unconfined core samples. Reasonable field percentages of available water vary from about 8–14% for the soils studied. Good agreement was found between values of hydraulic conductivity measured by the augerhole method and capillary conductivity computed from chloride accumulation although the variability of both the chloride data and hydraulic conductivity was rather high. In using chloride as a tracer to study water movement it was found necessary to take negative adsorption into account.

The agreement between the field data and the theory of steady state flow through unsaturated soil was satisfactory when no limit was set on evaporation by the moisture conducting properties of the soil. Where the potential evaporation rate exceeded the maximum flow rate possible through the soil profile evaporation from the soil was equal to, or less than the predicted maximum flow rate through the profile. Reduction of evaporation under these conditions was caused either by the establishment of a natural mulch or by the appearance of a salt crust at the soil surface.

The concept critical depth of watertable was used to characterize the potential salinization of a soil profile. It was concluded that the critical depth corresponds to the watertable depth at which the flow rate through the soil profile is reduced to values of about 0.1 cm day^{-1}, since at values < 0.1 cm day^{-1} a great increase of the watertable depth is necessary, in all soils investigated, to change the flow rate and therefore salt transport appreciably. In the case of controlling evaporation from bare soil a further reason is that the flow rate of 0.1 cm day^{-1} is sufficiently small to permit under favourable circumstances the maintenance of a dry surface mulch at any desired period of the year. Where salinization is reduced by a mulch it is seen that saline groundwater may be tolerated at a higher level if potential evaporation is high.

Generally the watertable should be kept at a deeper level in soils of intermediate texture than in either fine or coarse textured soils. The salinity hazard of the fine textured soils of the Murrumbidgee Irrigation Areas is markedly reduced if the watertable is kept at about 120 cm below the surface of a bare soil or, in the case of a cropped surface, 120 cm below the root zone.

SAMENVATTING

Bij het opstellen van een drainage-advies in irrigatiegebieden moet men dikwijls rekening houden, zowel bij het vaststellen van de capaciteit van het drainagesysteem als ook bij de beoordeling van de gewenste drainagediepte, met het bijna altijd aanwezige verzoutingsgevaar. Voor het vaststellen van de gemiddelde afvoer capaciteit kan in principe gebruik gemaakt worden van de gewenste uitspoelingsnorm, welke afhangt van de waterbehoefte en zoutresistentie van het betrokken gewas als ook van het zoutgehalte van het irrigatiewater. Bij de aanwezigheid van zout grondwater hangt de gewenste grondwaterstand af van de fysische eigenschappen van het bodemprofiel die bepalend zijn voor het vochttransport in onverzadigde grond, van klimatologische factoren en van het zoutgehalte van het grondwater. Uit de literatuur blijkt dat een gewenste grondwaterstand van 180 cm of dieper gewoonlijk aanbevolen wordt, vaak zonder de beschikking te hebben over voldoende gegevens. Het verband tussen de grondwaterstand en bovengenoemde factoren werd daarom onderzocht in de irrigatiegebieden langs de Murrumbidgee rivier in Australië.

Het drainageprobleem in deze gebieden is vooral ontstaan door de ongunstige fysiografische ligging van het bevloeide terrein, waardoor de natuurlijke afvoer van drainagewater, zowel boven- als ondergronds, bemoeilijkt wordt. Het verzoutingsverschijnsel wordt niet veroorzaakt door het voorkomen van zout in het irrigatiewater maar door het opstijgen van zouthoudend grondwater.

Het klimaat van dit gebied is semi-aried, de gemiddelde jaarlijkse regenval bedraagt 400 mm en is vrij gelijkmatig verdeeld over het jaar. De gemiddelde potentiële verdamping, E_0, berekend uit meteorologische gegevens met behulp van de formule van Penman en afgeleid uit gegevens van een Australische standaard verdampingspan, varieert van 0.14 cm dag^{-1} in juli tot 0.78 cm dag^{-1} in januari. De gemiddelde potentiële verdamping is gedurende het gehele jaar hoger dan de gemiddelde regenval. Het verschil tussen de berekende en gemeten waarde van E_0 bedroeg in 1961 niet meer dan 15% en er mag daarom ook aangenomen worden dat de Australische standaard verdampingspan in zijn huidige opstelling de potentiële verdamping, E_0, vrij nauwkeurig meet, althans gedurende perioden van een maand of langer. De gemiddelde potentiële evapotranspiratie, E_t, van grasland bedroeg 0.76 E_0 gedurende het najaar, winter en vroege voorjaar van 1961. Voor citrus werd een waarde van $E_t = 0.66\ E_0$ berekend voor de zomerperiode van 1951. Enkele waarden van de werkelijke verdamping, E_a, van beide gewassen werden berekend.

Op een vijftal proefveldjes werd het verband tussen verzouting, bodemfysische eigenschappen en grondwaterstand nagegaan. Deze proefveldjes werden voornamelijk geselecteerd op basis van verschillen in het bodemprofiel. Speciaal werd aandacht besteed aan gronden met een hoog kleigehalte en een slechte doorlatendheid, hoofdzakelijk omdat er nogal tegenstrijdige opvattingen bestaan in de literatuur omtrent de gewenste grondwaterstand in dergelijke gronden.

Het verzoutingsproces werd bestudeerd door middel van maandelijkse vocht- en zoutbepalingen. Deze gegevens werden gebruikt voor het berekenen van gemiddelde dagelijkse capillaire opstijgsnelheden voor maandelijkse periodes tussen mei 1960 en oktober 1961. De in de profielen optredende potentiaal verschillen werden berekend uit metingen van de zuigspanning met tensiemeters, die op regelmatige afstanden tussen het bodemoppervlak en de waterspiegel geplaatst waren. Het verband tussen het capillaire geleidingsvermogen en de zuigspanning werd gevonden door het capillaire geleidingsvermogen te berekenen met behulp van de gemiddelde capillaire stijgsnelheden en het gemiddelde potentiaal verschil gedurende periodes waarin de capillaire opstijging ongeveer stationair was. Dit laatste werd beoordeeld met behulp van de tensiemeterregistraties. Na afloop van het verzoutingsonderzoek werden de proefveldjes bemonsterd teneinde gegevens te verkrijgen aangaande het soortelijk gewicht, de porositeit, de korrelgrootte-verdeling en het verband tussen vochtgehalte en zuigspanning van de verschillende bodemprofielen. De doorlatendheid werd bepaald met de boorgatenmethode.

Het percentage water dat niet beschikbaar is voor plantengroei was in het algemeen groot vanwege het hoge kleigehalte en de betrekkelijk slechte structuurtoestand van de onderzochte grondsoorten. Duidelijke verschillen waren vaak merkbaar bij het vergelijken van de pF-krommen enerzijds bepaald aan ongeroerde monsters in het laboratorium en anderzijds afgeleid uit de resultaten van het veldonderzoek. Dit is te wijten aan het hysteresisverschijnsel en het voorkomen van lucht-insluiting in de veldmonsters en aan het opzwellen van de ongeroerde monsters in het laboratorium. De hoeveelheid water beschikbaar voor plantengroei varieerde in het veld van ongeveer 8 tot 14% voor de verschillende bodemprofielen. De waarde van de doorlatendheid, gemeten met de boorgatenmethode, sloot goed aan bij de gemeten waarden van het capillaire geleidingsvermogen in alle onderzochte profielen, hoewel de variabiliteit van de doorlatendheid zowel als van de chloorbepalingen in het algemeen groot was. Het bleek nodig de negatieve adsorptie van het chloride-ion in rekening te brengen bij het berekenen van vochttransport.

De overeenstemming tussen de experimentele resultaten en de theorie van stationaire opstijging in onverzadigde toestand was bevredigend in die gevallen waar de verdamping van de grond niet gelimiteerd werd door bodemprofiel-eigenschappen. Waar dit wel het geval was was de verdampingsintensiteit of gelijk aan de voorspelde maximale opstijgsnelheid door het bodemprofiel, of, vaak aanmerkelijk, lager. Deze lage verdampingsintensiteiten waren te wijten aan de vorming van een zoutkorst of een sterk ingedroogde laag (een z.g.n. "mulch") aan het bodemoppervlak.

De zogenaamde critische grondwaterstand werd gebruikt voor het karakteriseren van het potentiële verzoutingsgevaar van een bodemprofiel. Deze critische diepte stemt overeen met die grondwaterstand waarbij de maximale stijgsnelheid naar het bodemoppervlak gereduceerd is tot een waarde van ongeveer 0.1 cm dag^{-1}, omdat voor alle onderzochte bodemprofielen gevonden werd dat een grote verlaging van de waterspiegel noodzakelijk is om de capillaire opstijging en dus ook het zouttransport in belangrijke mate te beperken in het traject < 0.1 cm dag^{-1}. Indien het gaat om de verdamping (en dus vocht- en zouttransport) van kale grond te beperken kan men tevens als argument aanvoeren dat de opstijgsnelheid van 0.1 cm dag^{-1} over het gehele jaar kleiner is dan de potentiële verdamping, waardoor de mogelijkheid tot het in stand houden van een effectieve mulch begunstigd wordt. In dit geval kan men zich op kale grond een hogere grondwaterstand permitteren naarmate de potentiële verdamping hoger is.

In het algemeen moet de grondwaterstand op grotere diepte gehouden worden in zavelgronden dan in klei- zowel als zandgronden. In de irrigatiegebieden langs de Murrumbidgee is het verzoutingsgevaar niet erg groot in gronden met een hoog kleigehalte, wanneer de grondwaterstand bij kale grond niet hoger oploopt dan ongeveer 120 cm beneden het maaiveld of, indien een gewas geteeld wordt, 120 cm beneden de bewortelingszone.

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