CROP GROWTH AND WATER USE FROM SALINE WATER TABLES

by

Md. Humayun Kabir Khandker
B. Sc. Agric. Engg. (Hons.) (BAU, Bangladesh)
M. Sc. Agric. Engg. (Newcastle University, UK)

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ABSTRACT

How much water can a crop abstract from below a saline water table and how does the salinity affect yield? These questions are important because shallow groundwater may represent a substantial resource in flat, low-lying areas, but may also represent a threat to sustainability where salinity is high. A series of experiments in a glasshouse aimed to elucidate irrigation management practice under salinity conditions and to develop a root uptake model under both osmotic and matric stresses.

The extraction of soil water and groundwater by lettuce and perennial ryegrass crops were measured in three instrumented lysimeters. Water table depths were 0.6, 0.9 and 1.2 m below the soil surface. The lysimeters were initially saturated with saline water (electrical conductivity 4.5 dS m^{-1} for lettuce, 9.4 dS m^{-1} for the first crop of ryegrass and 0.4, 7.5 & 15.0 dS m^{-1} for the second crop of ryegrass) and drained until an equilibrium soil water profile was attained. Water with the same electrical conductivity was then supplied by Mariotte siphons to maintain the constant water table. The water table contribution was recorded and water losses from the soil profile were estimated from daily readings of soil water potential using tensiometers and gypsum blocks. Solute samples were extracted periodically for salinity measurement. The cropping period of lettuce was 90 days from sowing and the 1st & 2nd cropping periods of ryegrass were 223 & 215 days respectively.

The first ryegrass experiment showed that the water table depth (60, 90 and 120 cm) did not have significant contribution (37, 36 and 36 mm) on either total soil moisture use or groundwater contribution. Similar results were found for total soil moisture use for lettuce, though the groundwater contribution varied significantly. The second ryegrass experiment showed that salinity at the water table strongly influenced total soil moisture use, but the total groundwater contribution varied only slightly.

The overall crop experiments show that the groundwater contribution was within the range of 25-30% of the total water use, except for the 15 dS m^{-1} treatment where the contribution was greater than the soil moisture use. Groundwater contribution rate was higher when the plants were subjected to more osmotic and matric stresses. Yield component data show that increasing salinity leads to a reduction in total yield, but the drymatter proportion was higher. Higher salinities occurred in the upper 15 cm of the root zone, because of the greater soil moisture depletion. Below that depth the salinization rate was smaller, because of the greater groundwater contribution in the later part of the

season. There is reasonable agreement between measured and estimated (based on convective transport theory) values soil salinity. Salinities increased in the root zone by about 3-fold of initial salinity for lettuce and around 4-fold for ryegrass in the top 5 cm depth, but below 15 cm depth it was less than 2 fold.

Finally, a simplified model was developed to describe the interaction of root-zone salinity and water uptake, considering salinity and water stress as additive. The model shows that the higher the root-zone salinity stress, the higher the predicted water uptake while plant uptake considered -1.5 MPa. This variation is ranged from 4 to 17% for 0.4 to 9.4 dS m^{-1} and 30 % for 15 dS m^{-1}.

The model was developed in a climate with low atmospheric demand, but needs testing in a more severe environment.

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BASIC NOTATIONS

([ ] Indicates dimensions)

a = dimensionless fitted parameter (see Equation 3.14)

α = root radius (see Equation 2.5 [L]), and

a_1 = empirical constant (see Equation 6.5)

AI = Willmott's index of agreement (see Equation 6.10)

B = empirical constant to represent a specific root activity factor (see Equation 2.7)

b = dimensionless fitted parameter (see Equation 3.14)

β = radius of effective cylinder of soil surrounding the root (see Equation 2.5 [L]),

B_1 = dimensionless coefficient to represent the geometry of the flow (see Equation 2.9)

b_1 = empirical constant (see Equation 6.5)

C = concentration of solute (see Equation 3.17 [ML^{-3}])

c_1 = empirical constant (see Equation 6.5)

C_{cr} = concentration of solute in the upward flow solution (see Equation 3.23 [ML^{-3}])

C_r = capillary rise from water table (see Equation 3.12 [LT^{-1}])

C_{sm} = concentration of solute in soil moisture (see Equation 3.23 [ML^{-3}])

CT = cumulative potential transpiration (see Equation 6.7 [L])

CT_a = cumulative stress-adjusted potential transpiration (see Equation 6.7 [L])

Δ = Gradient operator

D = hydraulic diffusivity (see Equation 2.6 [L^2T^{-1}])

d_1 = empirical constant (see Equation 6.5)

D_c = hydrodynamoc dispersion (see Equation 3.21 [L^2T^{-1}])

D_i = root density (see Equation 2.10 [LL^{-3}])

D_m = the mechanical dispersion coefficient (see Equation 3.19 [$L^2 T^{-1}$])

D_p = molecular diffusion coefficient in porous media (see Equation 3.18 [L^2T^{-1}])

Ψ_{fi} = loss in potential due to friction (see Equation 2.10 [L]).

E = the maximum possible evapotranspiration (see Equation 2.12 [LT^{-1}])

E_a = actual evapotranspiration (see Equation 6.5 [LT^{-1}])

EC = electrical conductivity of the solution (see Equation 3.24 [dS m^{-1}])

E_o = potential evaporation (LT^{-1}),

E_p = potential evapotranspiration (see Equation 6.5 [LT^{-1}])

E_s = maximum possible soil evaporation (see Equation 2.12 $[LT^{-1}]$)

ET = evapotranspiration [L]

Φ = wall extensibility (see Equation 2.13 $[L^{-1}(L)^{-1}]$)

g = acceleration due to gravity (see Equation 2.3 $[LT^{-2}]$)

H = hydraulic potential (L)

H_{root} = root water potential at the soil surface (see Equation 2.8 [L])

ϑ = velocity of water relative to the solid phase (see Equation 2.3 $[LT^{-1}]$),

J_c = flux of solute (see equation 3.17 $[ML^{-2}T^{-1}]$)

$J_{c\text{-}d}$ = convective-dispersive transport of solute (see equation 3.22 $[ML^{-2}T^{-1}]$)

J_d = diffusive flux (see equation 3.18 $[ML^{-2}T^{-1}]$)

J_h = dispersive flux (see equation 3.19 $[ML^{-2}T^{-1}]$)

K = hydraulic conductivity $[LT^{-1}]$

K_a = stress-adjusted crop coefficient (see Equation 6.7)

K_{cr} = crop coefficient (see Equation 6.7)

K_i = root-soil system permeability (see Equation 2.10 $[LT^{-1}]$)

K_{sat} = saturated hydraulic conductivity is equal to 'a / b' (see Equation 3.16 $[LT^{-1})$

L = root length (see Equation 2.13 [L])

L = rooting depth (L)

λ = dispersivity (see Equation 3.20 [L])

L_{ar} = actual depth of root zone (see Equation 2.12 [L])

L_{er} = effective rooting depth (see Equation 2.12 [L])

L_i = initial root depth (see Equation 6.6 [L])

L_{max} = maximum root depth (see Equation 6.5 [L])

L_t = Root depth at day d (see Equation 6.6 [L])

M = soil moisture extraction (see Equation 3.12 $[LT^{-1}]$)

M_e = external pressure of soil resistance (see Equation 2.13 [L])

n = dimensionless fitted parameter (see Equation 3.14)

n_1 = empirical constant in Equation 3.20

p = pressure of water (see Equation 2.3 $[ML^{-1}T^{-2}]$)

p_a = reference pressure being the atmospheric pressure and it is assumed to be uniform (see equation 2.3 $[ML^{-1}T^{-2}]$)

P_t = turgor pressure (see Equation 2.13 [L])

q = flux (see Equation 3.12 $[LT^{-1}]$)

θ = volumetric moisture content (L^3L^{-3}), and

θ_s = the saturation moisture content (see Equation 3.9 $[LT^{-1}]$)

R = drag coefficient (see Equation 2.3 [dimensionless])

ρ = density of water (see Equation 2.3 $[ML^{-3}]$),

r = radial distance from the axis of the root (see Equation 2.6 [L]),

RDF = proportion of total active roots (see Equation 6.8)

RDF = proportion of total active roots in depth increment dz,

RMSE = root mean square error (see Equation 6.10)

RNA = correction depth (L), to account for non-active roots (see Equation 2.12 [L])

R_o = reduction factor in water uptake due to osmotic potential in the root zone (see Equation 3.26)

RRES = dimensionless coefficient for longitudinal water flow in the root xylem (see Equation 2.8)

R_{root} = root resistance (see Equation 2.7 [TL^{-2}])

R_{soil} = soil resistance (see Equation 2.7 [TL^{-2}])

R_u = specific root resistance per unit length (see Equation 2.7 [TL^{-1}])

S = sink term (see Equation 2.7 [$L^3\ T^{-1}$])

S_{rz} = solute content in the root zone (see Equation 3.23 [M])

T = transpiration rate per unit area of the soil surface (see Equation 6.4 [LT^{-1}])

t = time (T)

t_{inc} = time increment from the planting date (see Equation 6.5 [T])

t_j = present day (Julian day) (see Equation 6.6 [T])

t_m = day from planting to reach maximum root depth (Julian day) (see Equation 6.6 [T])

t_{max} = number of days to root profile maturity (see Equation 6.5 [T])

t_p = time considering planting date equals 1(see Equation 6.5 [T])

t_{pj} = planting day (Julian day) (see Equation 6.6 [T])

U_i = inflow to roots (see Equation 2.10 [L^3T^{-1}])

V_{cr} = volume of upward flow into the root zone (see Equation 3.23 [L^3])

V_i = volume of soil in the layer (see Equation 2.10 [L^3])

v_p = Pore water velocity (see Equation 3.20 [LT^{-1}])

V_{sm} = volume of soil moisture evaporated from the root zone (see Equation 3.23 [L^3])

x = distance (L)

Y = minimum turgor pressure for expansion (see Equation 2.13 [L])

Ψ_L = Leaf water potential (L),

Ψ_m = matric potential (see Equation 3.1 [L]),

Ψ_n = pneumatic potential (see Equation 3.1 [L])

Ψ_o = osmotic potential (see Equation 3.1 [L])

Ψ_r = water potential at the root surface (see Equation 2.5 [L]),

Ψ_{Si} = soil water potential in the layer (see Equation 2.10 [L]),

Ψ_{soil} = Total potential of soil water (L),

Ψ_t = total potential (see Equation 3.1 [L]),

Ψ_z = gravitational potential (see Equation 3.1 [L]),

Ψ_{Zi} = loss in potential due to elevation (see Equation 2.10 [L])

z_1 = vertical distance from soil surface down to the water table (see Equation 3.14 [L])

Part 1: PRELIMINARY

Chapter: 1

INTRODUCTION

1.1 Salinity: a threat to agriculture

The use of water is increasing faster than the increase of human population. As water is being used at increased rates, the trend of degradation of water quality is also increasing. The quality of water for agriculture is continually degrading as irrigation requirements in the world are extending.

Increasing demand for water resources is forcing a reassessment of management practices as farmers and communities deal with the use of water of decreasing quality. Farmers are faced with balancing the benefits of leaching and drainage to reduce soil salinity against the damage from pesticide and other chemical residues that are transported by drainage water. This balance is being shifted as irrigation water quality changes or, in some cases, when regulations concerning groundwater pollution affect irrigation management practices (Cardon, 1990).

Irrigation with saline water and soil salinity problems are widespread. Millions of hectares of land throughout the world are too saline to produce an economic crop yield and more land becomes non-productive each year because of salt accumulation. Szabolcs (1985) reported that about one-third (estimated by FAO and UNESCO) of all existing irrigation systems (totalling about 250 million hectares) was seriously affected by salinity and waterlogging and that 10 million hectares of irrigated land was abandoned annually. The salinity problem is more widespread and acute in arid and semi-arid regions, because of extensive irrigation, low rainfall and the relative scarcity of good quality water (Yaron, 1981). Tanji (1990) indicated that 23% of the cultivated land in the world is saline, another 37% is sodic and that the salt-affected soils are not only limited to semiarid and arid regions. In several other regions, the climate and the mobility of salts produce saline water and soil seasonally.

AFRC news supplement (1990) documented that about 950 million hectares of the earth's surface is affected by salt, an area which is continuing to increase. Between 30-50 % of irrigation agriculture is affected by salinity. The distribution of salt-affected soils in the continents and sub-continents is illustrated in Table 1.1 (from Szabolcs, 1985).

Table 1.1: Salt affected soils in the continents and sub-continents of the world.

Continents / Sub-continents	Million ha
North America	15.755
Mexico and Central America	1.965
South America	129.163
Africa	80.608
Southern Asia	87.608
North and Central Asia	211.686
Southeast Asia	19.983
Australia	357.330
Europe	50.832
Total	954.832

The reported figures of salt affected area clearly shows that the salinity problem already has spread over a large area and that it is continually increasing.

Salinity and its associated problems have greatly emerged from irrigation practices along with naturally salt-containing water resources.

All natural waters contain dissolved salts. Typical salt concentrations encountered in agricultural operations range from 5 to 10 mg l^{-1} in rain water to 10,000 mg l^{-1} or higher in some highly saline waters. In coastal areas, where sea water intrusion occurs, salt concentrations in shallow groundwater may approach 35,000 mg l^{-1}. Water may acquire additional salts during percolation through soils. These salts are derived from the residual soil solution, from highly soluble minerals precipitated previously and from weathering of other soil minerals (McNeal, 1977).

Many irrigation projects throughout the world operate with 25 to 40 percent overall efficiency. Thus, perhaps, only one-third of the water released at the project headwork is actually beneficially used for evapotranspiration by crops. The remaining two-thirds of water is not only wasted, but enhances waterlogging and salinity problems (Salazar et al., 1984).

Irrigation of valleys and plains typically entails the development of a high water table, resulting inevitably from the application of water amounts greater than the amounts used by the crops. As all irrigation waters contain salts, and as the concentration of salts tends to increase when soil moisture evaporates from the surface or is extracted and transpired by plants, the maintenance of favourable conditions in the root zone requires the application of an extra amount of water to leach out the excess salts. This extra amount of water percolates downward and tends to raise the level of the water table (Hillel, 1990).

The severity of water table rise in irrigated agriculture can be anticipated from Table 1.2. Soil salinity and sodicity continue to be major problems in food producing areas of the world. This is true despite advances made in our knowledge of the physics and chemistry of saline soils and the institution of modified management practices. As the future of irrigated agriculture is assessed, it is possible that existing problems could worsen because of the reduction in the sources of good quality water, increasing populations and possible degradation of groundwater by leachate from cropped soils. Disposal of drainage or agricultural wastewater that is collected to prevent resalinization presents an additional problem that must be addressed now. Not only salts, but particular constituents such as selenium and boron in drainage water, intensify the problem of disposal and present new environmental hazards (Biggar et al., 1990).

Table 1.2: Observed irrigation induced water table rises (from Smedema, 1990).

Areas of the world	Original water table depth (m)	Water table rise (cm / year)
Bhatinda / Punjab (India)	40-50	45
N. Pakistan	15-30	30-50
Punjab (M. Pakistan)	10-15	20-40
Khaipur Command/ Sind (S. Pakistan	4-10	10-30
State Farm 29 / Xinjang (China)	5-10	35-70
Murray-Darling Basin / N.S. Wales (Australia)	30-40	50-150
Noubaria/Western Desert (Egypt)	15-20	200-300
East Ghor North / Jordan Valley (Jordan)	10-15	nil
Beni Amir (Morocco)	15-30	150-300
Gezira Scheme (Sudan)	20-50	nil
Salt Valley / Arizona (USA)	15	60
Amibara / Middle Ahwaz Valley (Ethiopia)	10-15	100

It is increasingly recognised that irrigation and drainage research should be broad and far-sighted in scope to deal with water related problems that are developing on irrigated lands and associated supplies. The continued use of water supplies to produce irrigated crops requires new water management strategies and technologies that are realistic in terms of biological, economic, social, and environmental constraints. New research is needed to provide for optimum use of water resources and maximum protection of water quality.

1.2 Irrigation and salinity hazard

In irrigated areas, salinity is an almost universal threat because irrigation waters contain hundreds or thousands of mg per litre of salts. Besides the salts introduced by irrigation, soil in arid regions may contain salts of geological origin. When soil water is derived from irrigation, the soil water is at least as saline as the irrigation water and usually more so. The increase in salinity results from evaporation and consumptive use of water by plants, both of which concentrate the salts in the residual soil water. When water application exceeds evapotranspiration, the excess water carries the concentrated soil solution out of the root zone, which may cause waterlogging and subsequently increase the movement of salts into the upper root zone.

Apart from the limitations, the problems associated with drip irrigation are the accumulation of salts at the periphery of the wetted circles surrounding each emitter that can hinder the growth of subsequent crop, and the excessive through-flow and leaching which can take place directly under the drip emitters (Hillel, 1987).

Saline waters are not always suitable for sprinkler irrigation. According to Ayars and Westcot (1985), water containing 3 meq per litre of Na or Cl should not be sprayed onto plant leaves because of the damage to the foliage. Also, soil salinity around the surface of the cone of soil water developed by the drippers in the drip irrigation method requires flooding by fresh water or by water of low electrical conductivity (EC) to leach the accumulated salts on the soil surface (Balba, 1990).

In the state of Hariana in north-western India, the water table in the large area, that is underlain by brackish and saline groundwater, is rising owing to percolation losses from irrigation. Disposal of surplus water that cause the water table to rise can be solved by providing an outfall drain, but the cost of outfall drain becomes so high that its economic feasibility seems not encouraging (Boumans et al., 1988).

Plant growth is a function of salinity and the matric potential of soil water. Salinity control is a major objective of irrigation management, even though the primary objective of irrigation is to maintain soil matric potential in a range suitable for optimum crop yield. These objectives are often closely related because salinity is almost a universal threat where irrigation water typically contains significant amounts of dissolved salts (Shainberg and Oster, 1978; Jensen, 1983).

Frequent irrigation increases the average soil water content, but it develops much higher salt concentrations in the lower root zone. Proper drainage to improve salt management requires that the water table should remain low enough so that leaching can be efficiently accomplished and that salinity does not rapidly increase due to upward capillary movement (Salazar et al., 1984).

Bernstein and Fireman (1957) stated that salt accumulation 5 to 10 fold greater than originally present in the plough layer may be found in furrow-irrigated ridges after a single irrigation. Such zones of intensified salinity have frequently been the immediate cause of germination failures of row crops. Salt concentrations in ridges often greatly exceeds the salinity of the drainage waters. If washed either by irrigation or rain into the root zone the crops may be killed unless the salts are promptly leached out.

Soil water availability to crops can be improved in saline soils by the selection of appropriate management procedures. The management procedures that require minor change are frequent irrigation, selection of salt tolerant crops, additional leaching, pre-plant irrigation and seed placement. The alternative management procedures that require a significant change are changing the irrigation methods, altering the water supply, land grading, modifying the soil profile and installing artificial drainage. Salt control by irrigation practice is widely accepted, but no irrigation method is considered a permanent and hazardless way of controlling salinity (Jensen, 1983).

Balba (1990) reported that, when soil is irrigated with saline water, the following salinity hazards take place:

a) The soil retains an amount of the added salt equal to the retained water times its salt concentration.

b) As the water flows through the soil profile, it displaces the soil solution. Thus the soil loses part of its soluble salts.

c) At steady state, the amount of salt removed from the soil equals the amount of salt retained with the retained water of each irrigation.

d) The salt precipitates originally present in the soil are subject to dissolution and movement by the applied water.

e) The salt applied with the water is subject to precipitation due to reactions with constituents of the irrigated soils including the soil air.

f) The exchange reaction takes place simultaneously with the water flow into the soil.

g) In the case of plants growing in the irrigated soils, the situation is further complicated because the plants absorb water at a much higher rate than they absorb salts.

h) Rainfall, water-logging or presence of impermeable layers in the soil profile may accentuate or hinder one or more of the above processes, thus changing the expected results.

Irrigation is unavoidable because agricultural productivity is directly dependent on it, but the proper management of irrigation waters or the development of a new irrigation method(s) or the modification of an old system(s) should always be concerned to combat the inherent problems that may develop.

1.3 Water table management: an alternative to crop water supply

Although the need has been accepted for centuries to lower the water table in waterlogged soils and leach down salts by drainage, recognition of the use of shallow water tables or controlling optimum water table depth or the combined approach of drainage/sub-irrigation for crop production purpose is much more recent. The sub-irrigation concept is drawing much attention nowadays to consider it as an irrigation method or a subsidiary to the existing methods. The extent of this interest can be appreciated from the following individual pieces of information.

Water table control for both drainage and sub-surface irrigation represents a potential management alternative on soils with adequate drainage. In the Netherlands, drainage is essential during the spring to remove excess water and, during the summer, capillary rise from the water table is an important source of water for plant growth (Ratts and Gardner, 1974).

Using the water table as a source of water to the plants has several advantages such as: reducing irrigation needs, lowering costs and decreasing the amount of water that needs to be removed by artificial drainage (Torres, 1987).

A shallow water table can contribute significantly to water use by crops. Several reports (Misra et al., 1969; Sharma and Singh, 1971) suggest that crops responded very little to irrigation mainly because of the shallow water table in the experimental area. Hassan (1990) indicated that a properly managed water table can be a resource irrespective of shallow, medium and deep-rooted crops.

Although a third or more of the total water requirement may be available from groundwater at shallow depths (Salazar et al., 1984), it is important to consider the detrimental effects of a shallow water table on crops. Shallow water tables may prevent adequate root development and this will result in small available moisture capacity in the root zone if the water tables drop (Salazar et al., 1984).

One management strategy, which is prompted in areas of low salinity, is to control the water table depth by encouraging use by the crop of the shallow groundwater thus reducing irrigation requirements and drainage volumes (Grismer and Gates, 1987). Studies in California and Texas have shown that some tolerant crops (cotton, alfalfa, barley) are capable of utilizing a significant portion of their evapotranspiration demand from shallow brackish or saline water tables (Grimes et al., 1984; Ayars and Schoneman, 1986).

Utilization of shallow groundwater holds great potential for reducing subsurface drainage flows. Research is needed on crops to assess the extent of their use of saline groundwater. Timing of irrigation may further control groundwater use. Improved irrigation scheduling techniques need to be developed that allow the farmer to utilize a portion of the high water table to meet the crop evapotranspiration while minimizing the upward flow of salt into the root zone (Westcot, 1988).

Though the introduction of groundwater pumping for lowering the water table or reuse is a water management system for salinity control, the inherent problem is aquifer salinization (Heuperman, 1988).

The rate of capillary upward movement from the groundwater depends on the depth of water table below the root zone, soil moisture content and gradient, soil texture and structure, capillary properties and evaporative conditions. Generally, in coarse textured soils, rapid movement can occur over short distances with large moisture gradients. Water can move greater distances in fine textured soils, but movement is slower (Salazar et al., 1984)

Torres (1987) reported that a shallow water table may be either beneficial or harmful depending on the type of crop, soil texture and quality of the groundwater. During the last decade, much attention has been devoted to studies of the water table as a source of water for irrigation (sub-irrigation) rather than considering it as a drainage problem, as was the approach in the previous years. In this regard, two limiting depths to the water table must be considered: i) an upper limit determined by the aeration requirements of the crops and ii) a lower limit where an adequate water supply can obtained.

Bradford and Letey (1992a) reported that the cost of installing drainage may be avoided by altering irrigation management and allowing the crop to draw water from the water table.

A shallow saline water table has developed throughout the irrigation region of northern Victoria, Australia with commensurate rises in soil salinity. The re-use of the moderately saline groundwater is one salinity management strategy under consideration. However, this option is only applicable if productivity losses are minimal (Smith et al., 1993).

Gupta and Abichandi (1970) reported that saline groundwaters, having an EC range of 4 to 10 dS m^{-1}, occur widely in Western Rajstan, India. Some of these areas are used for growing salt tolerant Kharchi wheat.

Melvin et al. (1990) reported that sub-irrigation with water-table management is a growing concept in selected humid areas of the USA, e.g. in the Midwest and Southeast regions. A dual purpose sub-irrigation/drainage water-management system appears to be best suited to flat, poorly drained soils in humid and semi-humid climates where annual excess precipitation slightly exceeds annual irrigation requirements. Cooper et al. (1992) reported that a new concept in Midwestern irrigation, USA, is the use of the same drain lines for both sub-irrigation and drainage to provide a total water management system and water table control. Shirmohammadi et al. (1992) also reported that water management alternatives for humid regions may be categorised as drainage (surface and subsurface drainage) and drainage with water table control (controlled drainage and controlled drainage-sub-irrigation). Subsurface drainage, alone, mainly lowers the water table during the wet period until an equilibrium condition exists, governed primarily by drain depth. Controlled drainage is achieved by placing a control structure, such as a flashboard riser, in the outlet ditch or subsurface drain outlet to control the rate of subsurface drainage. Controlled drainage-sub-irrigation may be the most economical and feasible method for the shallow water table conditions in the eastern United States. This system is similar to the controlled drainage system, except that supplementary water is pumped into the system to maintain the water table at its present level during drought periods.

Benefits of using shallow groundwater include reduced irrigation, lower production costs, moderation of groundwater moving to deeper aquifers, and minimization of groundwater requiring disposal through drainage systems (Hoffman et al., 1990).

Bradford and Letey (1992 b) reported that excess water during any irrigation caused a rise in water table, but this water remained available for later crops which lowered the water table. Under water table conditions, higher simulated yields were achieved by applying less irrigation during the crop season and more during the pre-irrigation for salt leaching purposes. It is suggested that having low salinity in the upper part of the root zone during the initial stages of production is important for better crop establishment (Letey, 1993).

From the view of irrigation management, it may be possible to incorporate sub-irrigation in the irrigation management systems, especially where there is a shallow saline water table. Two major factors should be appraised when considering sub-irrigation as a potential source: how much water can be made available to crop water demand, and how much salt accumulates in the root zone due to capillary movement.

1.4 Water table management concept

Sub-irrigation in its natural or artificial form is the process by which water is supplied to the plants' root zone. Water table management is a natural process of sub-irrigation. The ratio of water supply to crop water demand depends on the position of the water table with respect to the rooting depth. In usual practice, the saline water table is exploited as a supplementary source of water supply and the rest of the crop water demand is fulfilled by accompanying irrigation or artificial sub-irrigation with non-saline or relatively low salinity water to maximize the production potential.

Attempts are also made to lower the water table by drainage to minimize the rate of salt accumulation and thus reduce the salinity hazard. The impact of the water table depth and soil properties on the rate of upward movement must be known to evaluate what depth to the water table should be maintained in irrigated agriculture. This information is also desirable when estimating the amount of water available to plants due to upward movement of groundwater, thereby reducing the irrigation requirement (Hoffman et al., 1990).

A climate with high atmospheric demand may be a major constraint to using a highly saline water table as a single source of crop water supply. However, in a climate with low atmospheric demand, the potential of utilizing highly saline water table for crop production has not yet been tested or modelled. The inherent problems with traditional irrigation are forcing us to make such a effort. The present view of saline water management for crop growth is to make water available in the soil by pre-irrigation with saline water and thereafter supplement the crop water needs from the saline water table. Water table management enables us to reduce the usual crop water requirements through the reduction of evaporation loss from the soil surface if water table depth can be optimally maintained.

1.5 Problem to be investigated

There is a little scope for using water of degraded quality for domestic and industrial purposes and hence, the first target for using degraded quality water is agriculture. By this time, the use of saline water has proven beneficial to agricultural production, but the best possible way(s) to use saline water in agriculture is still unclear. Hanson and Kite (1984) stated that the beneficial effects of water and the detrimental effects of salt from a water table on crop production have not been completely quantified.

In spite of general acceptance and awareness of the global hazard of salinization and alkalinization, these processes have not been arrested or diminished. On the contrary, they are expanding. This fact is reflected in a great number of books, papers and reports devoted to the subject. The weakness is not always the lack of sufficient study, but rather the lack of sufficient knowledge of how to adopt methods that are both technically sound and economically feasible (Szabolcs, 1985).

A immense amount of work has been done on management strategies like drainage and leaching, irrigation with nonsaline water above saline water tables, blending of saline and nonsaline water, or deficit irrigation, etc. for management of salty soils. The question may arise of how much or to what extent sweet water is available in saline-prone zone. Sometimes a shallow aquifer of sweet or less saline water above highly saline aquifer may be available, but the abstraction of such a shallow aquifer is dangerous because of the subsequent adverse effects on that aquifer or neighbouring aquifers. However, the fact is that the problem of high salinity is becoming an increasing concern for irrigation and drainage management.

So far it has been understood from the literature that little work has been done on saline water-table management for sub-irrigation and possibly no clear-cut protocol has been put forward for proper sub-irrigation management. So, it is a question of to what extent the sub-irrigation or subsurface irrigation in the saline-prone areas can take the place of the widely used practices of simultaneous irrigation and drainage methods. Hoffman et al. (1990) specifically, mentioned that, the relationships among crop water use and the depth and salt content of groundwater are not well understood. Several experiments have been conducted, but generalization are difficult to make based upon these results.

Therefore, the following information(s) may be needed to develop and intensify the idea of saline water sub-irrigation method on crop production mechanics:

a) what is the maximum salt load developed in the root zone by capillary rise by growing a longer-duration crops (e.g.. perennial crops) solely from the water table management to determine the actual potential of sub-irrigation for crop production under high salinity;

b) in what way can the groundwater uptake be maximized and what is the pattern of soil moisture and groundwater use from a highly saline soil profile and water table;

c) what is the magnitude of the sub-soil water use below the root zone; and

d) to what extent can the lower root zone extract water especially when the upper root zone becomes inactive. This has been brought to my attention because irrigation agronomists are usually very much concerned to keep the salinity as low as possible in the upper root zone rather than in the lower.

It is hoped that, the present study, 'Crop Growth and Water Use from Saline Water Tables' will elucidate the future direction of irrigation management practice and, thus, it is aimed to conceptualize the mechanics of water utilization by crops when dealing with highly saline waters.

1.5.1 Specific objectives

Experiments will be conducted in lysimeters with controlled water tables as an indicative model for field practice with the following specific objectives:

a) to measure the groundwater contribution to crop water requirements from shallow saline water tables;

b) to assess the behaviour of crop water use under sub-irrigation and the soil salinization impact on the crop production function for different combinations of salinity and water tables; and

c) to develop a simulation model for water uptake by roots from a salinized soil profile and saline water table.

Chapter: 2

LITERATURE REVIEW

2.1 Solute movement

In order to control the salinity of soil water, it is necessary to understand the mechanisms of solute movement into and away from the root zone.

2.1.1 Solute transport: general phenomena

The transport of solute or solute ions related exclusively to water movement is called convective or mass transport. The dissolved ions tend to move with the liquid water whenever soil water is dynamic, such as during infiltration, redistribution or evaporation.

In the solution phase, ions move at random by thermal motion, the mechanism of this movement is called molecular diffusion. The molecular diffusion is proportional to the gradient of chemical potential of solute in a soil solution. The rate of diffusion also varies with the diffusion coefficient modified by a tortousity factor, the viscosity of solution, and the electrostatic interaction. All these factors account for a reduction in the ion velocity.

Hydrodynamic dispersion occurs when soil solution flows through a soil volume. On the microscopic scale, the soil solution does not move at the same rate throughout the soil volume. It is not a driving force like convection or diffusion to cause solute movement, rather it represents the variability of flow velocity within an individual pore or relative to other pores or both.

The flow velocity can be decomposed into two parts: i) the average velocity, and ii) the deviation. Hence, the average flux of a solute can be considered equal to the sum of a) the convective flux, which is the flux carried by water at the average velocity; and b) the dispersive flux, which is the flux produced by spreading or dispersion caused by the fluctuating velocity.

Unfortunately, different names are often given to the diffusion and dispersion parameters: a) diffusion as molecular diffusion or, diffusion or, diffusion coefficient; b) dispersion as mechanical dispersion or, dispersion or, dispersion coefficient; and hydrodynamic dispersion as diffusion-dispersion or, diffusion-dispersion coefficient or, apparent diffusion coefficient.

Solute transfer in a porous medium is associated with several mechanisms, such as molecular diffusion, hydrodynamic dispersion, thermal diffusion, mass flow of water (convection), salt sieving, and gravitational descent (Diestel, 1976). The gravitational descent mechanism was ignored by Diestel because differences in densities of soil solution are small.

During the transport process, solutes tend to disperse. Two principal factors that produce solute dispersion are i) molecular diffusion, i.e., random movement of particles by thermal motion, and ii) mechanical dispersion. The mechanical dispersion is caused by the following factors: a) velocity distribution across individual soil pores, b) different average pore-water velocity existing within different size pores, c) tortuous branching pore sequences that promote spreading of a solute as displacement proceeds (Beer, 1979).

Diffusion can take place in any of the solid, liquid or gaseous phases as a result of the random thermal motion (Brownian motion). If a concentration gradient exists in the solution, then solute will diffuse (or move on average) from a region of higher concentration to one of a lower concentration (Elrick and Clothier, 1990).

Nasser and Horton (1992a) reported that salt sieving develops when solutes are more restricted than the water in their movement through the soil. The restriction of solute movement in soil drier than the field capacity (at about -3.4 m matric pressure) is generally caused by an electrical double layer. They also reported, through Boast (1973), that three aspects of the erratic flow of soil solution through a soil volume cause hydrodynamic dispersion. First, within a given pore, the flow rate of soil solution is lower near the walls of the pores than in the middle. Second, flow is faster in large pores than in smaller pores. Third, soil solution flow does not flow simply in the direction of soil volume; soil solution flows in some pores at an angle to the mean direction of the soil solution flow.

Depending on the experimental conditions, the measured concentration is flux-averaged or volume-averaged. Flux-averaged concentration, which represent the mass of solute per unit volume of fluid passing through a unit cross-section during an time interval, usually refers to effluent (Kreft & Zuber, 1978); whereas volume-averaged represents the average concentration of solute within a finite representative elemental volume of porous media or mean local values of soil pore-water solute concentration (Bear, 1972). Bowman and Rice (1986) and Jaynes et al. (1988) assumed that the concentration measured by suction samplers is volume-averaged.

In practice, convection and diffusion interact to produce a complex phenomenon, called hydrodynamic dispersion, as a result of which the solute movement exceeds that expected from summing convective and dispersive fluxes (Rose, 1977).

2.1.2 Causes of root zone salinization

When water moves through the soil, soil water carries its solute load in its convective stream, leaving some of it behind to the extent that the component's salts are absorbed, taken up by plants, or precipitated whenever their concentration exceeds their solubility (e.g., at the soil surface during evaporation) (Hillel, 1980).

By far the most common cause of high salinity is however called salinization, i.e., the accumulation of salts in the upper layers of the soil and especially in the root zone from some outside source. Root zone salinization is mainly caused by irrigation with inadequate leaching capillary salinization from groundwater and by evapotranspiration (Smedema and Rycroft, 1983).

Capillary salinization is the salinization due to evaporation from the groundwater table. For this type of salinization to take place, saline groundwater must occur within such a depth that upward capillary flow is able to reach the evaporation zone. Capillary salinization mostly depends on capillary water flux & diffusion in response to concentration gradients (Smedema & Rycroft, 1983; Hillel, 1980).

2.1.3 Variability in field-level solute transport

The convection Dispersion Equation (CDE) has been used for many years to describe solute flow through soil columns and is the foundation for various models of vertical solute flow. Unlike small soil columns, field soils are heterogeneous and display variations in their hydraulic properties. To quantify the solute transport in an actual field situation the CDE approach is modified, e.g. i) the Stochastic approach (Sposito et al., 1986), ii) the Transfer Function model (Jury, 1982), and ii) Markov Chains approach (Knighton and Wagenet, 1987). These are not discussed here, being beyond the scope of this investigation.

Coefficients of variability for solute concentration in soil ranging from 60% to 130% have been reported in different studies (Jury and Sposito, 1985). Solute velocity and hydrodynamic dispersion are assumed to be log-normally distributed (Nielsen et al., 1973). Solute velocity and hydrodynamic dispersion vary not only laterally across the field, but also with depth (Biggar and Nielsen, 1976; Jury, 1982). The concept of dispersivity was originally used for saturated conditions and now is accepted as valid for unsaturated flows with some reservations (Dagan and Bresler, 1979; Sposito et al., 1986). Under unsaturated conditions, dispersivity depends on the soil water content as the pore-water velocity depends on it (Kirda et al., 1973).

The reliability of the prediction of solute movement depends on the quality of the input data (e.g. the water content at the beginning of the experiment) and the quality of the (measured) soil characteristics (e.g. hydraulic conductivity) [Knighton and Wagnet, 1987; Costa et al., 1991]. Bresler (1972) indicated that for many practical salinity control purposes it may be assumed that, under transient conditions, the overall diffusion-dispersion term contributes very little compared to the macroscopic-average viscous flow. Ayars et al. (1977) reported that salt transport due to dispersion in partially saturated soils is negligible compared to convective transport of solute. El-Hassey (1991) found under lysimetric experimental conditions that the measured salinity was on average 16.95% higher than the predicted salinity based on convective flow only. He also mentioned that the error in the prediction could be due to possible by-pass flow around the edge of the lysimeter or due to the dissolving of salts from the soil matrix or from neglecting dispersion in the numerical model. Ismail (1990) showed that there is a strong relationship between actual evapotranspiration and the salt balance of the soil, which makes it a vital factor to account for salt balance.

2.1.4 Conclusion

Salt movement through the soil is mainly a consequence of water movement, i.e. convective transport. The effect of hydrodynamic dispersion (diffusion + dispersion) depends on the flow velocity of soil water. When leaching of salts down from the root zone is an objective, this effect can be significant. During evaporation when the hydraulic conductivity reduced successively with soil drying, the extent to which, this effect will contribute to the salt movement compared to convective transport needs to be evaluated separately.

2.2 Water movement

2.2.1 Water movement mechanisms

Water movement from soil to roots may be as liquid, as vapour or as both and the degree of movement depends on local soil-water environment.

The availability of soil moisture in the unsaturated zone depends on the thermodynamic state of the water in soil and plants. The extraction of soil moisture by plants therefore follows the thermodynamic laws of water movement. Unsaturated soil moisture flow is governed by evaporation , when there is no internal drainage. The moving force in an unsaturated soil is subject to a matric potential, which is equivalent to negative pressure potential. The gradient of this potential constitutes a moving force (Klausink, 1969 ; Hillel, 1980).

The matric potential is affected by i) adsorption, ii) attraction between water molecules and ions in the electrical double layer of clay particles, and iii) a small deviations in the soil air pressure from the existing atmospheric pressure (Stroosnijder, 1976). In the saturated region, the attraction of soil matrix is negligible. Pressure merely results from the hydrostatic pressure, so that value for the pressure potential (Ψ_p) is positive. The pressure potential in the saturated zone has been termed 'submergence potential' (Rose, 1966).

Rose (1963a) reported that it is not always easy to specify the relevant potential for liquid flow in porous materials, but for vapour flow the mechanism is doubtless one of molecular diffusion under a vapour pressure gradient which may be set up by a temperature gradient, by a solute concentration gradient, by a matric potential gradient or by a combination of all three. His experimental results (Rose, 1963b) showed that vapour transfer becomes significant when soil becomes very dry.

The osmotic or solute potential reduces the total potential energy. The question is to what extent, the soil matrix itself— (particularly clay layers within the soil) can restrict the passage of various solutes which in effect acting as a selective membrane, can make an osmotic potential gradient be effective as a hydraulic gradient in inducing convective water flow (Hillel, 1980). Warrick (1990) more clearly reported that, the ability of the soil matrix to restrict flow and serve as a partially semipermeable membrane becomes the crucial factor as to whether an osmotic potential can be sustained. The expert consensus is that, except for where biological membranes are present (such as plant roots), hydraulic forces dominate water flow in soils.

Letey et al. (1968) studied the effect of osmotic pressure gradients on water movement in unsaturated soil with clay and fine sandy loam soils and concluded that, at low suctions (up to 0.05 MPa), for practical purposes the osmotic potential gradient can be neglected in the consideration of water flow through soil. They also reported that the coefficient relating water flux to osmotic pressure gradients is low at high suctions (up to 1.5 MPa) so the amount of water moved by an osmotic gradient is not great. Of course, the consideration of osmotic potential gradient is only applicable when a semi-permeable membrane present, e.g. root cell membrane, or an air-water meniscus.

Abd-El-Aziz and Taylor (1965) found that the osmotic pressure gradient in an unsaturated soil system had a very small effect upon water movement. The salt concentrations they used were 0.2 N and 0.3 N. The soil water suction must be greater than 5 bar before osmotic pressure gradients would be very effective in causing water movement. Their data would indicate that the effect is greater at lower average salt concentrations compared to higher salt concentrations (Letey et al., 1968).

Raats and Gardner (1974) described an expression for flux, θ or $\theta\vartheta$ from a consideration of the forces acting upon the water, either from a microscopic, or macroscopic point of view. Macroscopically, they consider flowing water to be subject to three forces : (i) a force arising from a spatial variations of the water pressure, (ii) the gravitational force, and (iii) a drag force associated with the movement of the water relative to the solid phase. The differential balance of forces may then be written as:

$$\theta\vartheta p + \theta\rho g \Delta z + R\vartheta = 0 \qquad \text{......................(2.1)}$$

Dividing equation (2.1) by R and multiplying by θ, it becomes,

$$\frac{\theta^2\vartheta p}{R} + \frac{\theta^2 \rho g}{R} + \theta\vartheta = 0 \qquad \text{....................(2.2)}$$

$$\text{or,} \theta\vartheta = -\frac{\theta^2\vartheta p}{R} - \frac{\theta^2 \rho g}{R}$$

$$= \frac{\theta^2\vartheta p}{R} - K\Delta z \qquad \left[\begin{array}{l} \because K = \dfrac{\theta^2\rho g}{R} \\ \text{or,} \dfrac{\theta^2}{R} = \dfrac{K}{\rho g} \end{array}\right]$$

$$= -\frac{K\vartheta p}{\rho g} - K\Delta z$$

$$= -K\Delta\Psi_m - K\Delta z \quad \left[\because \Psi_m = \frac{p - p_a}{\rho g}\right]$$

$$= -K\Delta(\Psi_m + \Psi_z)$$

$$\text{or,} \theta\vartheta = -K\Delta H \quad [\because H = \Psi_m + \Psi_z] \qquad \text{.........(2.3)}$$

which is the form of Darcy's law,

where,

θ = volume of water per unit volume ($L^3 L^{-3}$),

ϑ = velocity of water relative to the solid phase (LT^{-1}),

K = hydraulic conductivity (LT^{-1}),

p = pressure of water ($ML^{-1}T^{-2}$),

p_a = reference pressure being the atmospheric pressure and it is assumed to be uniform ($ML^{-1}T^{-2}$),

Ψ_m = matric potential (L),

ρ = density of water (ML^{-3}),

g = acceleration due to gravity (LT^{-2}),

R = drag coefficient,

Ψ_z = gravitational potential (L), and

H = hudraulic potential (L).

2.2.2 Conclusion

The driving force for water flow in the soil-plant-atmosphere continuum is a difference in water potential. The osmotic potential in soil water, which causes the reduction of plant uptake is a well established phenomenon. But the significant effect of gradient in this potential in water movement only has been recognised in clayey soils, though the magnitude of the effect is not known.

2.3 Hydraulic conductivity

Hydraulic conductivity is the flux of water per unit hydraulic gradient per unit time in the porous medium (usually known as proportionality constant), which depends on the driving force (hydraulic potential) and the transmitting properties of the flow medium.

The hydraulic conductivity of the soil is important in determining the maximum infiltration rate, the resistance to flow to plant roots and the rate of drainage of saturated soils. Infiltration rates and drainage of saturated soil are determined by the saturated hydraulic conductivity of the soil and resistance to flow to plant roots are determined by unsaturated conductivity of the soil, (Campbell and Mulla, 1990).

Kablan et al. (1989) reported that, although methods do not exist to easily predict the values of hydraulic conductivity over a wide range of moisture contents from fundamental soil properties, a number of methods have been reported to estimate hydraulic conductivity using macroscopic determinations of water flow and the corresponding gradient of hydraulic head. They commented that the statistical methods developed by Childs and Collis-George (1950), Millington and Quirk (1959), Brooks and Corey (1964), Green and Corey (1971), Mualem (1976) and Van Genuchten (1978, 1980) are not adequate for the complex structure of most soils, because their predictions of $K(\theta)$ depend upon on assumption of random distribution of pore space with little or no regard to the shape and arrangement of pores. They also added that the methods may be acceptable for use in hydrological models where prediction of $K(\theta)$ in the wetter range would be important, but would not be suitable for plant water-uptake models over ranges of lower θ.

Vereecken et al. (1990) tested the equations of Gardner (1958), Gilham et al. (1976) and Wind (1955) to estimate unsaturated hydraulic conductivity on 127 soil cores from a wide variety of Belgian soil series and found Gardner's three-parameter model to be the best performer. They showed that K_{sat} and b parameters were relatively insensitive to estimate the unsaturated hydraulic conductivity. The equations are as follows:

$K = a / [(-\Psi_m)^n + b]$ ....(Gardner, 1958);

$K = K_{sat} / (-\Psi_m)^n$ ....(Wind, 1955); and

$K = a\theta^n$ ....(Gillham, 1976).

where,

K	unsaturated hydraulic conductivity (LT^{-1}),
K_{sat}	saturated hydraulic conductivity (LT^{-1}),
Ψ_m	matric potential (L),
θ	actual soil water content ($L^3 L^{-3}$), and
a, b & n	fitted constants.

Eching and Hopmans (1993) pointed out that the optimized soil hydraulic functions as determined from soil cores do not necessarily represent the behaviour of soil in-situ. Field methods are generally considered more reliable for determining hydraulic conductivity but are restricted to high water contents (Arya et al., 1975).

Wendroth et al. (1993) re-evaluated the evaporation method for determining hydraulic conductivity functions on unsaturated soils having textural classes of sandy loam, silty loam & clay and concluded that the evaporation method is an elegant, simple & inexpensive technique. They also added that a major limitation of this method is caused by the estimation of near-zero hydraulic gradients close to soil water saturation.

Ilyas et al. (1993) conducted experiments to improve the hydraulic conductivity of saline-sodic soils (a fine loamy) in the field and concluded that it is difficult to improve the physical properties of this type of soil in a short period of time. Sub-soiling and drainage were not very helpful in increasing the permeability, whereas the combination of straw, gypsum and crops showed promising results than the use of one individual technique.

Reddi and Danda (1994) compared the soil hydraulic properties equation (Van Genuchten, 1980) with laboratory and field experimental results and found a large differences between the techniques; even the n parameter varied. For example, they found n values for a particular sandy loam soil with a constant saturated hydraulic conductivity of 300 cm d^{-1} as:

$n = 2.1$ measured in the laboratory;

$n = 2.99$ measured in the field, and

$n = 2.70$ estimated.

They added that the equation does not adequately simulate the physics of the problem. Again, when they used the Van Genuchten's (1980) hydraulic properties relationships to predict the recharge from rainfall employing the UNSATI model (Van Genuchten, 1978; a finite element numerical model) and LPM (Danda and Reddi, 1992; a lumped parameter model which assumes that the entire unsaturated zone can be lumped together as one homogeneous unit). They conclude that both the available sophisticated (UNSATI) and the simple model (LPM) on water movement predicted equally poorly, which indicates that the importance of these parameters far outweighs the sophistication of the water flow models.

Nielsen et al. (1973) reported that the variability of hydraulic conductivity in field situatios is much greater than measured in the laboratory, which was found to be in the range of 50% to 200% .

Hassan (1990) measured diffusivities by two methods, soil water depletion (a field method, see section 4.2.5.2) and the one-step outflow method, and found that conductivity values obtained from diffusivity and soil water depletion agreed well up-to almost 100 kPa soil moisture suction and also that conductivity values from the drainage flux (another field method, see section 4.2.5.1) are slightly higher compared with those found from the two methods aforesaid. He also mentioned that hydraulic conductivity in the field can be determined from upward flux and hydraulic gradient data when there is a loss of water from evaporation and plant use. The equation is :

$$q = C_r + M = -K\frac{dH}{dz} \qquad \text{................}(2.4)$$

where,

q = upward flux (LT^{-1}),

C_r = capillary rise (LT^{-1}),

M = soil moisture extraction from below the root zone ($L^3\,L^{-3}$),

K = unsaturated hydraulic conductivity (LT^{-1}), and

dH/dz = hydraulic gradient ($L\,L^{-1}$).

2.4 Water table contribution to crop water use

Exploiting a shallow water table for crop production has received considerable attention since the 1980's. Capillary rise from a water table (popularly known as groundwater contribution in irrigation) and its rate are governed by soil and plant factors, evapotranspiration demand, water table depth and salinity.

2.4.1 Saline water table contribution

Namken et al. (1969), after a four-year study using lysimeters with moderately-saline shallow water tables, found that estimates of crop water use from soil moisture depletion lead to errors in the estimation of evapotranspiration due to the capillary water contribution. They also found that cotton grown on deep permeable soils obtained a substantial portion of its water needs from a static water table at depths of 91, 183 and 273 cm.

Kruse et al. (1986) used lysimeters to measure the water use by alfalfa from a shallow saline water table in Colorado State University, USA. Water tables were maintained at 60 cm and 105 cm, the salinity for irrigation water was 0.66 dS m^{-1}, and the salinities of the solutions used to create and maintain the water table throughout the growing season were 6 dS m^{-1} and 0.66 dS m^{-1} for two different water table treatments. The percentage of groundwater used by the crop (average of three years) was 62% for the 6 dS m^{-1} groundwater and 76% for 0.66 dS m^{-1} groundwater. The alfalfa appeared to use slightly more water from the less saline water treatment. He also reported that with the water table maintained 60 cm below the soil surface, established alfalfa can obtain water at potential rates for evapotranspiration from this source. The authors added that visual observation indicated that alfalfa growth above a water table was noticeably better than in lysimeters with no water table.

El-Hassey (1990) and Kruse et al. (1985) conducted experiments in Colorado State University, USA, on maize growth with saline water table depths of 60 to 105 cm having salinity of maximum 6.0 dS m^{-1}; the proportion of groundwater used is shown in Table 2.1.

2.4.1 Nonsaline water table contribution

Cambell et al. (1960) found that alfalfa in a semiarid region of USA, produced nearly the same yield with and without six irrigations applied per year when the water table was between 5 and 9 feet below the soil surface.

Follet et al. (1974), growing maize, alfalfa and sugar beets on a sandy soil with a declining water table in USA, obtained maximum yield when the water table was 69 cm deep at the beginning of the cropping season.

Sub-irrigation from water tables at 30, 60 and 90 cm was used by Stewart et al. (1980) to irrigate clover in Melbourne, Australia. The water table depth did not have any effect on the yield but a more active root zone was observed in 60 and 90 cm water table treatments when the soil moisture content was between 0.15 and 0.25 cm^3 cm^{-3}. They added that there is not an optimum position of the water table because the root system is continually growing and the crop needs for water are constantly changing. Thus, the water table needs to move closer to the root zone as the crop develops to compensate the increasing crop demand for water.

McMullin and Read (1983) conducted experiments in Alberta, Canada, using constant water table lysimeters filled with loamy sand and clay loam soils. Results indicated that nearly 50% of the consumptive use of barley was provided by a water table at 1.2 m deep.

Evapotranspiration studies relating crop yield and water use of sweet corn were conducted by Shih (1985) using organic soils of Florida, USA. The water table was maintained at 30, 60 and 90 cm which resulted in corresponding average evapotranspiration rates of 3.6, 3.0 and 2.3 mm/day respectively.

An experiment with maize planted on non-weighing lysimeters containing a sandy soil showed that 63% of evapotranspiration was provided by a water table at 155 cm (Benz et al., 1985) in North Dakota, USA. They also reported that groundwater contribution decreased with an increase in surface water applications, either irrigation or rainfall.

Ayars and Schoneman (1986) showed that cotton can extract significant amounts of water from a perched saline water table in California, USA. Tovey (1964) indicated that a high water table can be an asset to forage production if water table fluctuation can be controlled.

Torres (1987) reported that the water table contribution to evapotranspiration of wheat crop decreased linearly with depth. Independently of soil type, water table contributions for lysimeters with water tables at 50 cm was well above 90%, lysimeters with water tables at 100 cm had a contribution ranging from 40 to 70%, and lysimeters with water tables at 150 cm had a contribution 10 to 34% in Utah State University, USA.

Shih (1988) conducted an experiment in Florida, USA, on 'Drip irrigation and sub-irrigation of sugarcane' and found that the soil moisture content in the profile with the sub-irrigation system was much greater than with the drip irrigation. Before canopy closure of the plant cane, ET from drip irrigation was significantly lower than from sub-irrigation; by contrast, after canopy closure for the two ratoons, the ET from drip irrigation was significantly higher than from sub-irrigation.

Hassan (1990) conducted water use experiments in Newcastle University, UK, with bean, barley & lettuce above 60, 90 & 120 cm sweet water tables and the 60 cm water table showed the highest water table contribution as 34.7, 27.0 & 4.5 % for lettuce, barley & bean respectively. The reason for very low amount of groundwater use by bean was that initially, the soil moisture profile was not in equilibrium, i.e. the soil water draining was continuing.

Prathapar and Meyer (1992) reported that water table contributions were 16 and 29 % for two different treatments under irrigated conditions with same soils in Griffith Laboratory, Australia. The reason for the differing contribution was the initial soil moisture difference (17 & 23 %) in the top 10 cm of the soil. The maize crop was grown with a water table fluctuating between 60 to 130 cm for a period of 131 days.

A summary of the water table contribution under different crop water use experiments are presented in Table 2.1.

Table: 2.1 Summary of the percentage of the water table contribution to crop water use (experiments done by different investigators).

Authors	Crop	Water table depth (cm)	Water table salinity (dS m^{-1})	Water table contribution
Namken et al. (1969)	Cotton	91	Saline	54.0 %
		183	Saline	26.0 %
		274	Saline	17.0 %
Kruse et al. (1985)	Maize	60	0.66	58.6 %
		60	3.0	52.5 %
		60	6.0	55.3 %
		105	0.66	31.8 %
		105	3.0	25.1%
		105	6.0	31.1%
Kruse et al. (1986)	Alfalfa	60	0.66	76.0 %
		60	6.0	62.0 %
El-Hassy (1990)	Maize	60	3.0	43.3 %
Lal & Sharma (1974)	Wheat	126-266	Nonsaline	37.5 %
Stuff & Dale (1978)	Maize	125-200	Nonsaline	27.0 %
		60	Nonsaline	3.0 mm d^{-1}
		90	Nonsaline	2.3 mm d^{-1}
Wallender et al., 79	Cotton	212-266	Nonsaline	59.0-70.0 %
Stewart et al. (1980)	Clover	30	Nonsaline	——
		60	Nonsaline	——
		90	Nonsaline	——
McMullin & Read, 83	Barley	120	Nonsaline	50.0 %
Benz et al. (1984)	Alfalfa	155	Nonsaline	38.4%
Benz et al. (1985)	Maize	155	Nonsaline	63.0 %
	Alfalfa	155	Nonsaline	26.7 %

Table 2.1: (Continued)

Shih (1985)	Maize	30	Nonsaline	3.6 mm d^{-1}
Ragab and Amer, 86	Maize	25-55	Nonsaline	40.0 %
Tories (1987)	Wheat	50	Nonsaline	90.0 %
		100	Nonsaline	40.0-70.0 %
		150	Nonsaline	10.0-34.0 %
Meyer, 1987	Wheat	100 (Loamy soil)	Nonsaline	28.0-36.0 %
		100 (Clay soil)	Nonsaline	10.0-15.0 %
Hassan (1990)	Lettuce	60	Nonsaline	34.7 %
		90	Nonsaline	13.5 %
		120	Nonsaline	6.0 %
	Barley	60	Nonsaline	27.0 %
		90	Nonsaline	16.4 %
		120	Nonsaline	11.4 %
Prathapar & Meyer (1992)	Maize	60	Nonsaline	29.0 %
		130	Nonsaline	16.0 %

2.4.3 Conclusions

Most of the reported work was concerned with exploring nonsaline water tables; only a few papers dealt with saline water tables and the salinity was not very high. The water table experiments done by different investigators resulted in the following conclusions:

i) The water table can supplement the water needed for crop production, but not provide the total requirements, either saline or nonsaline.

ii) The proportion of the groundwater contribution varied widely and, the causes of such variations were expressed as general variations in the experimental conditions, i.e. the contribution varies as the water table depth, soil type, crop type, climate, and irrigation managements and water qualities. Thus, it is difficult to generalise the quantification of groundwater contribution.

iii) The lack of generating information regarding the absolute contribution of water table, especially, in case of a saline water table. The quantification of saline water table contribution was done in combination with surface water application, and as a result the actual effect of salinity and matric stress on capillary rise could not be reflected properly. Similarly, simply varying the water table depth can not explain the variations in contribution, unless the rooting depth is considered and this factor was not mentioned.

2.5 Water uptake by roots

Root water uptake is perhaps one of the most difficult components of the soil water balance to model. The sink term is used to represent water uptake by roots, as the volume of water extracted per unit time per unit bulk volume of soil, or in depth units, the rate of water extraction per unit depth. A usual method of taking root water extraction into account in the continuity equation for soil moisture flow is to incorporate a sink term in the latter; see, for example, Equation (2.11).

The water uptake by roots at various depths is required in order to compute water depletion in the root zone, soil moisture movement, etc. Water uptake by roots at different depths is governed by the rooting density distribution, hydraulic conductivities of the soil-root system and the availability of the soil moisture itself (Prasad, 1988), as well as pland demand for transpired water.

2.5.1 Water uptake modelling approach

All the available models for root uptake differ in the formulation of their sink term formulations. This means that, every model has accepted the Darcian flow equation for water movement through the soil, but water movement from soil to roots has been expressed in different ways. The method of formulating the sink term by all is to estimate the available water to plants. Two alternative approaches have been tried to model the sink term in quantitative physical terms:

a) the microscopic-scale approach which analyses the radial flow of water to individual roots, considered to be narrow tube sinks regularly spaced or clumping in the soil. Examples of such studies are: Philip (1957), Gardner (1960), De Willigen and Van Noordwijk (1987), Ehlers et al. (1991);

b) the macroscopic-scale approach which regards the root system in its entirety as a diffuse sink permeating the soil continuously, though not necessarily at uniform strength, through the root zone. Again the macroscopic approaches may be categorised into two groups based on i) consideration of the physics of water flow from the soil to root, and ii) simple empirical relation with the soil matric potential.

The first group of the macroscopic approaches applies the soil hydraulic conductivity to represent the soil to root flow. Examples of these models: Gardner and Ehlig (1962), Gardner (1964), Whistler et al. (1968), Nimah and Hanks (1973), Childs and Hanks (1975), Feddes et al. (1974), Hillel et al. (1976), Harkelrath et al. (1977), Rowse et al. (1978), Klepper (1991) and Wagnet et al. (1987). The second group incorporates an empirical stress coefficient based on matric or osmotic potentials. Such models are: Molz and Remson (1970), Feddes et al. (1976, 1978), Hoogland et al. (1980), Molz (1981), Belmans et al. (1983), Perrochet (1987), Van Genuchten (1987), Prasad (1988), Jarvis (1989), Ismail and Gowing (1990), El-Hassey (1991), Cardon & Letey (1992).

Among the models which exclusively dealt with saline water management are: Childs and Hanks (1975), Van Genuchten (1987), Ismail and Gowing (1990), El-Hassey (1991), and Cardon and Letey (1992).

Some of the sink term equations are as follows:

A. Microscopic approaches

The equation for water uptake by a single root proposed by Philip (1957) yields the following expression:

$$S = 2\pi K \frac{\Psi_m - \Psi_r}{\ln(\beta / \alpha)} \quad \text{..............................(2.5)}$$

where,

S = sink term, rate of water uptake per unit length of root ($LT^{-1}L^{-1}$),

Ψ_m = soil matric potential (L),

Ψ_r = water potential at root surface (L),

β = radius of effective cylinder of soil surrounding the root (L),

α = root radius (L), and

K = effective hydraulic conductivity of the soil (LT^{-1}).

The approach of the mathematical solution to a single root (Hillel, 1971), which is considered as a hollow cylinder of infinite length with uniform diameter and water extraction properties. In radial co-ordinates, the flow equation is expressed as :

$$\frac{\partial\theta}{\partial t} = \frac{1}{r}\frac{\partial}{\partial r}\left[rD(\theta)\frac{\partial\theta}{\partial r}\right] \qquad \text{....................(2.6)}$$

where,

r = radial distance from the axis of the root (L),
θ = volumetric moisture content ($L^3\ L^{-3}$),
D = hydraulic diffusivity ($L^2\ T^{-1}$), and
t = time (T).

Equation (2.5) is a flux equation which is simple to use in experimental work, whereas equation (2.6) is a second-order partial differential equation which can be used in modelling studies.

B. Macroscopic approaches (first group)

The macroscopic approach for root water uptake by Gardner and Ehlig (1962) used the following equation:

$$S = \frac{\Psi_t - \Psi_L}{R_{Soil} + R_{Roots}} \qquad \text{..........(2.7)}$$

where,

S = sink term caused by root extraction ($L^3\ T^{-1}$),
Ψ_t = total potential of soil water (L),
Ψ_L = leaf water potential (L), and
R = hydraulic resistance, when $R_{soil} = 1/BKl$, and $R_{roots} = R_u/l$

Root resistance is proportional to a specific resistance per unit length (R_u, in d cm $^{-1}$) and inversely related to the length of roots per unit volume of soil, l. Soil resistance is inversely proportional to unsaturated hydraulic conductivity, K and density of active roots, l. B is an empirical constant to represent a specific root activity factor.

Nimah and Hanks (1973) proposed an expansion of Darcy's equation between a point in the soil and the root surface to give the following equation:

$$S = \frac{[H_{Root} + RRES \times Z - \Psi_m]}{dx.dz}[RDF \times K(\theta)] \quad .....(2.8)$$

where,

H_{root} = root water potential at the soil surface where Z is defined as zero (L),

Ψ_m = soil matric potential (L),

$K(\theta)$ = soil hydraulic conductivity (LT^{-1}),

RDF = proportion of total active roots in depth increment dz,

Z = depth (L),

dx = distance between plant roots and the point in the soil where Ψ_m and Ψ_o are measured (L), and

RRES = head loss coefficient for longitudinal water flow in the root xylem (generally assumed to 1.05).

H_{root} is iteratively determined until its value is such that extraction equals potential transpiration, provided H_{root} is above a pre-established limiting lower value. After the lower limit on H_{root} is reached, extraction (as calculated by above equation) becomes less than potential transpiration and decreases as Ψ_m decreases. This reduced extraction proceeds until the Ψ_m equals H_{root}, whereupon extraction ceases.

Feddes et al. (1974) modified and field tested the model of Nimah and Hanks (1973), expressing the sink term as:

$$S = \left[\frac{\Psi_r - \Psi_m}{B_1}\right]K(\theta) \qquad .....(2.9)$$

where,

B_1 = coefficient to represent the geometry of the flow,

Ψ_r = water potential at the root surface (L),

Ψ_m = matric potential in the soil water (L), and

K = soil hydraulic conductivity (LT^{-1}).

Klepper (1991) expressed the water uptake by roots as:

$$U_i = V_i D_i K_i \left(\Psi_{Si} - \Psi_L + \Psi_{Zi} + \sum_{j=1}^{i} \Delta\Psi_{fi}\right) \qquad .... (2.10)$$

where,

U_i = inflow to roots (L^3T^{-1}),

V_i = volume of soil in the layer (L^3),

D_i = root density ($L\,L^{-3}$),

K_i = root-soil system permeability (LT^{-1}),
Ψ_{Si} = soil water potential in the layer (L),
Ψ_L = Leaf water potential (L),
Ψ_{Zi} = loss in potential due to elevation (L),
$\Delta\Psi_{fi}$ = loss in potential due to friction (L).

C. Macroscopic approaches (second group)

Molz & Remson (1973) expressed the combined equation for the continuity of flow and root water uptake as: the divergence of the Darcian velocity υ at a point in soils is equal to the negative of the volume rate of change of moisture and, in addition, moisture is removed directly from that point at a rate of S in such a way that is not included in the flow velocity. They also described the disadvantages of using the microscopic approach to simulate root extraction of water in relation to the difficulty of expressing boundary conditions. They developed a model assuming a static root extraction pattern. The sink term was a function of transpiration rate and a depth dependent function that follows the simple empirical rule of extraction that 40, 30, 20 & 10 percent of the total transpiration requirement is supplied by each successive quarter of the root zone. The one-dimensional combined flow equation is:

$$\frac{\partial\theta}{\partial t} = -\nabla\vartheta - S \qquad \text{.....................(2.11)}$$

$$= \frac{\partial}{\partial z}\left[K\frac{\partial}{\partial z}(\Psi_m + Z)\right] - S$$

and, $S = -\frac{1.6}{L^2}TZ + \frac{1.8}{L}T$

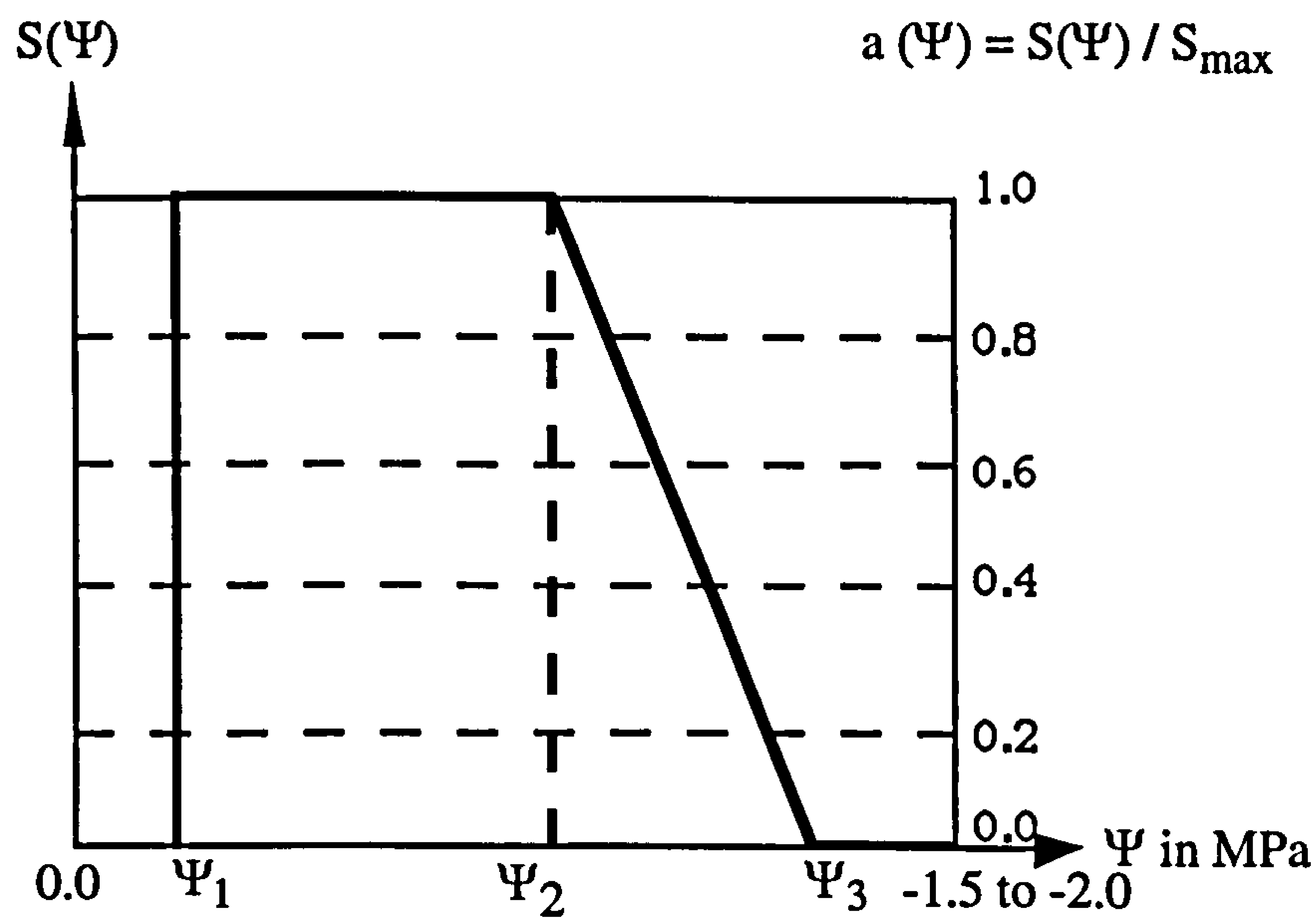
where,

θ = volumetric moisture content ($L^3\,L^{-3}$),
S = sink term ($L^3\,L^{-3}\,T^{-1}$)
L = vertical length of the root system (L),
Ψ_m = matric potential (L),
Z = vertical distance from the soil surface (L), and
T = transpiration rate per unit area of the soil surface and taken as average transpiration rate ($L\,T^{-1}$).
t = time (T).

Feddes et al. (1976) expressed the sink term by calculating a stress coefficient based on soil moisture content. Further, Feddes et al. (1978) modified the sink term to depend on soil moisture matric potential Ψ_m instead of the moisture content. The form of the equation is

$$S(\Psi) = S_{max}.\frac{\Psi - \Psi 3}{\Psi_2 - \Psi_3} \qquad \text{...............(2.12)}$$

The relation between $S(\Psi)$ and matric potential is as follows:



$S_{max} = E_{plant} / L_{er}$ and,

$E_{plant} = E - E_s$

where,

E = the maximum possible evapotranspiration (LT^{-1}),

E_s = maximum possible soil evaporation (LT^{-1}), and

L_{er} = effective rooting depth (L).

But, $L_{er} = L_{ar} - RNA$

when, L_{ar} = actual depth of root zone (L), and

RNA = correction depth (L), to account for non-active roots.

It is assumed that, under conditions drier than wilting point (Ψ_3) and wetter than a certain 'anaerobiosis point' (Ψ_1), water uptake by roots is zero. The value of Ψ_3 was taken -1.5 to -2.0 MPa of water. The stress coefficient based on soil matric potential ratio varies from 1.0 to 0.0 and the coefficient is assumed as 1.0 for some threshold value of matric potential (-0.04, 0.05 MPa , etc.) depending on the variation of climatic demand and 0.0 for -1.5 to -2.0 MPa. Hoogland et al. (1980) developed a similar model by integrating the root water uptake term from the soil surface to an increasing depth (z) less than or equal to the rooting depth until the integral became equal to the potential transpiration rate. If the integral over the rooting depth is less than the potential transpiration rate, water stress is considered to occur.

Saline water management models are discussed in Chapter 6.

2.5.2 Some views on model performance

Many analytical methods are available to determine the actual rate of soil moisture depletion in irrigated fields, with some considering it to be equal to the potential evapotranspiration, and others relating it to the availability of moisture in the root zone by linear, logarithmic, exponential or power functions.

Regardless of the approach, the depletion of moisture in many analytical models is computed on the basis of a root zone sealed at its lower boundary. Alternatively, irrigation is scheduled on the basis of a physical measurement of soil moisture content (or soil suction) in the root zone, and irrigation is applied when pre-defined limits are attained. While analytical models result in neglecting soil water flow from or to the sub-soil, the second method indirectly includes such a flow, but does not discriminate between evapotranspiration and moisture transfer at the lower boundary (Ghali & Svehlik, 1988).

Dynamic simulation models describe water uptake by root systems under field conditions as a function of soil depth and time. Many of these simulation approaches are based on Gardner's (1960) single root model. These simulation procedures follow the assumption that water uptake is proportional to a difference in water potential between the bulk soil and the root surface or the plant interior, to the hydraulic conductivity of the soil-plant system and to the "effectiveness" of competing roots in water uptake. The effectiveness factor accounts more or less empirically for the influence of various root system parameters on water uptake such as percentage of "active" roots absorbing water, root surface permeability, root length density determining the distance between neighbouring roots, or total root length and depth of the root system. Such models however, will not always reflect correctly the influence of root system characteristics on water uptake since these assumptions have rarely been tested under conditions. *In many instances, there is better agreement between simulated and measured total water use of plants than between predicted and observed water depletion by roots within individual layers of the soil profile (Alaerts et al. 1985).*

Choice of sink term might be difficult under real conditions and with validation data available. All sink terms require the choice of parameters adapted to the particular soil-plant combination. This almost implies some data-fitting, because independent data for similar conditions are rarely available, and the knowledge for extrapolating from different conditions is lacking. Too little seems to be known about plant transpiration and water extraction patterns to result in a general extraction term which is simple and economical to use and satisfying in concepts and results (Alaerts et al., 1985).

Most current models of the water uptake by plant roots are based on Richards' equation for the flow of water in unsaturated soils and, with the development of methods for measuring the hydraulic conductivity of soils, it became possible to calculate the resistance to flow in the soil portion of the flow path. The problem is how to couple the flow system of the plant to that of the soil. That problem, related with the functionall behaviour of the root membranes can be overcome by considering a distributed sink term moving downward through the soil profile (Gardner, 1991).

2.5.3 Water uptake and root development behaviour

Crop root systems are consist of two types of roots: i) downward-growing main roots and ii) lateral roots and their branches produced along the length of the main roots. These two types of roots may be functionally different. For the most part, the vertical main roots serve as conduits for water collected by branches, but the primary function of laterals, on average, is the absorption of plant resources from the soil (Klepper, 1991).

Molz and Peterson (1976) reported that water flow from roots to soil (negative direction) is very small compared to flow in the positive direction i.e. soil to roots. By contrast, Baker at al. (1992) reported that a measurable quantity lost from roots to dry soil can be found, referring to the work by McWilliam & Kramer (1968), Mooney et al. (1980), Baker & Van Bavel (1986,), Richard & Caldwell (1987) and Caldwell & Richards (1989).

The root depth increases steadily till mid season, and thereafter may be presumed to remain constant. It has generally been found that, during the growing stage, the ratio of consumptive use to the root zone depth is remarkably constant (Hansen et al., 1979). The rooting depth of well established perennials does not change significantly during one growing season and can be considered constant (Nimah and Hanks, 1973).

Reicosky et al. (1972) studied water uptake patterns of soybeans and indicated that water uptake was not necessarily related to root distribution and a small amount of roots in contact with the capillary fringe can supply most of the water required by the plant.

Water uptake rate increases with increase in root density depending on root age and soil water status (Sharma and Chaudhary, 1983). Borg and Grimes (1985) found that, irrespective of plant species, soil type, water regime or location, the increase in rooting depth with time follows a sigmoidal pattern. For simplicity the root development could be approximated by a straight line (Jensen et al, 1990). The root extraction pattern is assumed to be approximately 40, 30, 20 & 10%, respectively from the soil surface in the fourth-quartile divisions of the maximum rooting depth (Hansen et al., 1979).

The onset of soil drying stimulates root growth in more 'hospitable' layers. This new growth not only explores new soil volume but produces young roots with higher specific water uptake rates. Provided the balance between the availability of water from new explored soil volume and evaporative demand can be met, plant survival is assured (Meyer and Barrs, 1990). Holder and Brown (1980) found that water uptake of bean roots decreased as soil O_2 concentrations were lowered.

Under most field situations, a root system grows vertically downward at a rate of about 1 cm per day (Klepper, 1987). Physiologically, root tissues require O_2, a favourable temperature, delivery of materials for making new cells, and collection of ions & solutes to maintain cell osmotic values and turgor pressure for expansion (Klepper, 1990).

Van Bavel and Baker (1985) concluded from their experimental investigation that deeper roots can support functioning of smaller ones in the upper and driest part of the root zone. Vallderuten et al. (1975) reported that a relatively small proportion of the roots located in the vicinity of the water table is responsible for a large part of the water uptake.

Ehlers et al. (1991) concluded that, the potential water use by crops does not depend so much on rooting density but more on the maximum rooting depth on the root system.

2.5.4 Osmotic stress effect on water uptake

Osmotic stress in soil water causes the reduction of available water to plants and the degree of reductions vary with salt tolerance capability of plants, is a basic fact. The emerging concept of 'osmotic adjustment' is added to determine the plant uptake capacity from the salinity stress. Presumably, it is not yet clearly defined that what degree of salinity can attribute what degree of osmotic adjustment even for a particular plant species. However, the general concept of 'osmotic adjustment' is that plant can maintain their turgor potential by osmotic adjustment (lowering the osmotic potential due to solute accumulation).

Osmotic adjustment is a decrease in cell osmotic potential due to an increased number of solute molecules (Turner and Jones, 1980). Increased solute concentration by the mechanism of osmotic adjustment aids in the maintenance of cell turgor pressure when cell water potential decreases. Maintenance of root turgor pressure by osmotic adjustment may allow continued growth into more favourable root zones (Baker et al., 1992).

Mechanisms of salt tolerance and the ability of roots to take water up from a saline soil include: the ability to absorb soil water from rhizospheric soil solutions of decreasing osmotic potentials, and the expansion of a root system in order to utilize soil solution of lower concentrations from outside the rhizophere (Schleiff, 1983). He showed that, though barley roots extracted water from soil water osmotic potential up to -2.5 MPa, maize roots ceased water uptake at -0.9 MPa. He also mentioned that water uptake by barley roots was reduced a little at -0.9 MPa osmotic potential.

Taylor (1983) discussed soil salinity effects on root growth and concluded that i) increase in soil salinity decreases the rate of root elongation; ii) increase in salinity may upset the hormonal balance which affects root growth rate and membrane permeability to water and ions.

A study of the salinity effect on root water uptake showed that the critical (when water uptake start to cease) root water potential was -0.3 MPa for nine different crops including grass (Bresler and Hoffman, 1986).

Evlagon et al. (1990) reported that the length of the primary roots of maize seedlings was reduced by 54% after 4 days of growth in 0.1-strength Hoagland solution salinized with 100 mM NaCl and by 20% when 10 mM calcium was also added to the salinized root medium. Roots showed 69% osmotic adjustment in response to salinization, with or without extra calcium in the root medium.

Oosterhuis and Wullschleger (1987) observed that the osmotic adjustment of cotton roots was 0.18 MPa and osmotic adjustment of perennial grass was 0.20 MPa in tillers of water-stressed plants (Toft et al., 1987).

2.5.5 Climate & crop type influences on water uptake

Salazar et al. (1984) reported that the soil water reservoir available to the plant changes as the root system develops. Root depth varies with crop and variety, stage of growth and soil chemistry, structure, drainage and management. For example, too frequent irrigation may limit root development. The root system of a plant develops from seed depth at germination to a maximum depth when it reaches maximum vegetative development or until it encounters impermeable barriers or other obstacles to root development. Typical rooting depths for several commercial crops, divided into four group, is presented in Table 2.2:

Table 2.2: Typical rooting depth of some commercial crops.

Rooting depth & crop groups	Crop types
0.3 to 0.5 m I	Cabbage, celery, lettuce, onion, pineapple, sisal, potatoes, spinach, vegetables.
0.5 to 1.0 m II	Banana, beans, beets, carrots, clover, peppers, groundnuts, peas, soybeans, sugar beets.
1.0 to 1.5 m III	Barley, citrus, cucumber, sunflower, small grains, maize, melons, flax, sweet potato, wheat.
1.5 to 2.0 m IV	Alfalfa, cotton, deciduous orchard, grapes, sorghum, sugar cane.

For irrigation management purposes, the root zone may assumed to be developed linearly from planting depth at time of planting to typical maximum root depth at full cover. Salazar et al. (1984) also presented some data on the permissible soil water depletion fraction as a function of the above crop groups and evaporative demand for maximum yield conditions as follows (Table 2.3):

Table 2.3: Soil water depletion fraction for crop groups and maximum evapotranspiration.

Crop	Maximum Evapotranspiration (mm/day)								
Group	2	3	4	5	6	7	8	9	10
I	0.50	0.425	0.35	0.30	0.25	0.225	0.20	0.2	0.175
II	0.675	0.575	0.475	0.40	0.35	0.325	0.275	0.25	0.225
III	0.80	0.70	0.60	0.50	0.45	0.425	0.375	0.35	0.30
IV	0.875	0.80	0.70	0.60	0.55	0.50	0.45	0.425	0.40

2.5.6 Root development physics

Roots elongate as a result of turgor pressure in cells in the elongation zone. The physics of this process (Greacen and Oh 1972; Lockhart, 1965) is expressed by the formula:

$$\frac{1}{L}\frac{dL}{dt} = \Phi(P_t - Y - M_e) \qquad ....(2.13)$$

where,

L = root length (L)

t = time (T)

Φ = wall extensibility [$T^{-1}(MPa)^{-1}$]

P_t = turgor pressure (MPa)

Y = minimum turgor pressure for expansion (MPa)

M_e = external pressure of soil resistance (MPa).

Under most agricultural situations, radial root resistances are much are greater than perirhizal resistances which are those resistances associated with the soil in the rhizophere (draw down resistance) (Gardner, 1960) and with the interface between the root and soil (contact resistance) (Herkelrath et al. 1977; Klepper 1990).

The root length density in well-watered crops is high in surface soils and decreases with depth. With drying, there was a complete reversal of the root length density distribution with significant loss of root material in the top of the profile and gain in lower profile (Klepper, 1973). The soil water potential at which root elongation ceases varies with soil texture and bulk density but one can expect little root elongation in soils drier than about -0.8 MPa matric potential (Ehlers et al., 1980).

2.5.7 Conclusions

i) The quantity of root mass development varies exponentially down to the root zone and, therefore, water extraction of 40, 30, 20 & 10 % from the agronomic thumb rule under well-watered conditions is a reasonable approximation. But under stress (matric, or osmotic, or both) conditions, this approximation needs to be corrected by incorporating root depth-dependent stress coefficient.

ii) The maximum rooting lengths for many crops under well watered conditions are available in the literature, but the time needed for growing the rooting length is not available as such. Under matric stress conditions, roots have a habit of penetrating further down than usual to search for water. Under osmotic stress conditions, the maximum rooting length becomes shortened and thus an adjustment is needed by considering slower growth rate in respect to salinity.

iii) The maximum rooting depth for different commercial crops is classified as shallow, medium & deep rooted, but root growth rate could not classified as such because the crops biological life varies. Rather the evidence is that the root growth rate is faster at the beginning and then reduced with physiological maturity. The growth rate at full potential is usually between 2.0 and 1.0 cm per day.

v) The root water uptake model based on the macroscopic approach is becoming more popular because of the difficulty of the microscopic approach in matching the highly temporal & spatial variations in physical development and activity of roots in a simpler form with fair prediction of uptake. Such models have limitations, exemplified by poor performance in other situations except in conditions in which they have been developed. However, water uptake behaviour under prolong and high salinity stress above a saline water table without surface water application conditions, is not available. Till yet, the effect of salinity stress only has been tested along with surface water application conditions.

2.6 Crop water use and salinity

Crop water use is reduced when excessive dissolved salts occur in soils, apparently because plant bio-energy that would otherwise be used in biomass production is expended to extract water from the saline soil solution. Less frequently, growth may be reduced because of specific nutritional imbalances or ion toxicities when certain salt constituents are individually in excess (Rhoades, 1986). The relationship between osmotic stress and crop water use was first described by Wadleigh and Ayers (1945) they showed that the osmotic stress had an equivalent effect to matric stress on plant water uptake and biomass production of Red Kidney beans.

The limit of crop water use in terms of crop yield to salinity are now available as a general guide in the literature (Rhoades and Loveday, 1990). The salt tolerance limit of an individual crop can vary under different water management situations or one crop itself can develop more salt tolerance ability if it is repetitively grown in a salty environment. It also varies with stages of growth as well with climatic differences. However, the interaction between salinity and water use under some different situations is presented here.

De Malach et al. (1989) reported that strong interaction was found between the effects of salinity and temperature on onion germination. At 12^0 C, onion seeds germinated fully with water having an electrical conductivity (EC) of up to 30 dS m^{-1}. At 30^0 C, germination was arrested almost completely by an EC of 20 dS m^{-1}. Lower temperature and higher salinity both reduced the overall rate of onion germination.

Pearson and Bernstein (1959) investigated the effect of soil salinity at three stages of development on the growth of rice. The result indicated that salinity inhibited growth more severely at earlier stages than at later stages.

Bower et al. (1969) studied the effect of irrigation water salinity and leaching fractions on the root zone salt profile and yield of alfalfa, and found that yield was highly related to the average salinity of the root zone. For electrical conductivities of soil saturation extract of 5 and 11 dS m^{-1}, the yield decrease was 10 and 50% respectively.

Feigin et al. (1990) reported that salt greatly reduced the yield of lettuce tops and drymatter content though the reduction of drymatter content was not statistically significant. His experimental results showed that fresh yield and drymatter content of lettuce were reduced by 37% and 13% respectively when irrigation water salinity was 10-

11 dS m^{-1}. His result also showed that both fresh and dry root mass were increased by 38% and 50% respectively. The usual salt tolerance limit for lettuce is 1.3 dS m^{-1} (threshold value) with a reduction gradient of 13% per dS m^{-1}.

Moustafa et al. (1975) gave some results for the effect of saline water table (15 dS m^{-1}) on cotton yield. They showed that the highest yield was obtained from the treatment where the depth of saline water table was 160 cm, which was considered 100%. The yields of the other treatments were 89.1%, 81.9%, 70.4% and 46.5% for 130, 100, 70 and 40 cm depth of saline water table, respectively.

Pennington (1986) grew alfalfa in soil columns with irrigation by saline water of 4.0 and 8.0 dS m^{-1}. The irrigation doses were 1.1, 1.0, 0.75, 0.5 and 0.25 of measured actual evapotranspiration. He found that the lowest root zone osmotic potentials attained at the end of cropping period were -1.9, -2.0, -1.8, -2.6 & -2.4 MPa and -1.8, -2.2, -2.8, -3.1 & -4.5 MPa correspond to the leaching fractions of 9, 9, 6, 5 & 5 % and 23, 25, 18, 15 & 17 % respectively.

Bresler (1972) reported that, for most crop plants, growth reduction is controlled mainly by the total salt concentration of the soil solution and is largely independent of the different salt constituents in the solution.

Increased water use by grass has been associated with higher mowing heights. Mitchell and Kerr (1966) reported that a 37% decline in evapotranspiration between ryegrass (*Lolium perenne L.*) mowed at 50 and 25 mm under well-watered condition.

2.7 Water use efficiency

The term efficiency is generally understood to be a measure of the output obtainable from a given input. Irrigation and water-use efficiency can be defined in various ways, depending on the nature of the inputs and outputs considered. The criterion of plant water-use efficiency is the amount of dry matter produced per unit volume of water taken up by the plants from the soil. The economic criterion of efficiency is the financial return in relation to the investment in the water supply. Irrigation efficiency is generally defined as the net amount of water added to the root zone divided by the amount of water taken from some source. The difference between the net amount withdrawn from the source represents the seepage and evaporative losses incurred in conveyance to the crop, as well as the losses due to deep percolation below the root zone within the field and to runoff from the field (Hillel, 1987).

The relationship between salinity and water use efficiency depends upon the definition chosen for water use efficiency. Two definitions are commonly used, viz.: yield per unit of water evapotranspired and yield per unit of applied water. The rationale for the definition 'yield per unit evapotranspiration' is that liquid water which is not converted to vapour through evapotranspiration remains available for use. Only the water lost through evapotranspiration is consumed. This approach considers water quantity and ignores the quality considerations. Several investigations showed that the relationship between yield and evapotranspiration (ET) was identical as salinity led to reduced ET. The deficiency in using the ratio of yield to ET as definition is that large quantities of water may be necessary under nonuniform irrigation and with saline water to achieve maximum ET (Letey, 1993). Water-use efficiency is therefore a trade-off between productivity and financial return.

Chapter 3

THEORY

The purpose of this Chapter is to summarize the theory that has been used in the present investigation. Only one-dimensional unsaturated water flow along with solute transport has been employed.

3.1 Water movement through unsaturated zone

3.1.1 Unsaturated zone

The unsaturated zone, also known as the vadose zone or partially saturated zone, is limited at the top by the soil surface and at the bottom by the capillary fringe of the groundwater table. This zone includes the portion of the soil profile with a water content smaller than the soil prosity. In other words, it includes soil with negative water pressure less than the air or entry pressure. Temporary water saturation related to a perched water table or surface ponding conditions is usually considered within the unsaturated zone as well (Nielsen et al., 1986). The term 'unsaturated zone' is commonly associated with hydrological processes like infiltration, evaporation, groundwater recharge, soil moisture storage, soil erosion, and biological processes, especially these related to root growth (Costa, 1991).

3.1.2 Soil water equilibrium

The concept of 'total potential' has been used to analyse the equilibrium and transport of water in soils and plants. The 'total potential' of soil water is the amount of useful work that must be done per unit quantity of pure water to transfer reversibly and isothermally an infinitesimal quantity of water from a pool of pure water at a specified elevation at standard atmospheric pressure to the soil water at the point under consideration (Marshall and Holmes, 1988). The components of the total potential Ψ_t of soil water are the matric potential Ψ_m, the gravitational potential Ψ_z, osmotic potential Ψ_o and the pneumatic potential Ψ_n arising from the external gas pressure. Thus,

$$\Psi_t = \Psi_m + \Psi_z + \Psi_o + \Psi_n \qquad (3.1)$$

where,

Ψ_t = total potential (L),

Ψ_m = matric potential (L),

Ψ_z = gravitational potential (L),

Ψ_o = osmotic potential (L), and

Ψ_n = pneumatic potential (L), arising from changes in external gas pressure.

As the pneumatic potential in soil does not differ from the atmospheric pressure, $\Psi_n = 0$ and the total potential becomes:

$$\Psi t = \Psi_m + \Psi_z + \Psi_o \qquad ....(3.2)$$

The main advantage of the total potential concept is that it provides a unified measure by which the state of water in the soil or in the plant can be evaluated within the soil-plant-atmosphere continuum (Hillel, 1982).

Gravitational potential: It is independent of the chemical and pressure status of soil water, and dependent only on relative elevation. If the soil surface is chosen as the reference level, the gravitation for all points below the surface is negative with respect to that reference level. The gravitational potential per unit weight at a height (Z) below the surface is -Z.

Osmotic potential: salts in soil water affect its thermodynamic properties and lower its potential energy by lowering the vapour pressure of the solution. The osmotic potential does not affect liquid mass in soil significantly but plays an important role in the uptake of water by plant and in the diffusion of vapour.

Matric potential: It arises from the attraction of a matrix (soil, cellulose, protein, etc.) for water and the attraction of water molecules for each other. It refers to a negative pressure potential as found in unsaturated soil.

To describe the liquid water movement through soils, only the hydraulic potential (H) is considered, which is:

$$H = (\Psi_m + \Psi_z) \qquad ... (3.3)$$

Since, the matric potential represent the negative pressure, is used with negative sign. Assumed, soil surface is the datum for gravitational component, and as the water table is below the considered datum Ψ_Z will become -Z. Therefore, the equation (3.4) can be re-written as:

$$H = -(\Psi_m + Z) \qquad \text{... (3.4)}$$

To obtain the effect of osmotic potential, the presence of a membrane, which permeable to water but not to solutes, is necessary. The presence of root-membrane in the soil solution, which acts as a semi-permeable membrane, causes a need to consider the osmotic potential effect to describe the water movement from soil to roots. Hence, all the components of the equation (3.3) are needed to define the soil water status within the plants' root zone.

3.1.3 Available soil water to plants

Water will move from soil to roots, if soil water potential outside the root is greater than the water potential inside the roots, i.e. root xylem. Transpiration establishes a hydraulic gradient in the xylem which results in water flow from the roots to leaves. The difference between leaf water potential and soil water potential is an estimate of the driving force for water movement from soil to foliage. Note that, the driving force affects flow of liquid water to the evaporating surfaces, but it does not affect transpiration except when leaf water stress causes stomatal closure (Fiscus and Kaufmann, 1990)).

A representative midday transpiration rate in a mature plant produces leaf water potential (Ψ_L) around -1.5 MPa at a transpiration rate of 2.3 x 10^{-4} m^{-2} s^{-1} (Campbell and Turner, 1990). They also mentioned that variation in leaf water potential due to changes in soil water potential are generally very small compared to variations due to transpiration rate. However, the available potential for water movement from soil to roots is the difference between the leaf water potential and the soil water potential.

Again, within the plant cells, the water potential is the sum of osmotic potential of the symplasm and the turgor in the cell. The presence of dissolved solutes lowers the vapour pressure of the solvent (water), raises the boiling point, and lowers the water potential of aqueous solutions.

3.1.4 Soil water flow equation

The formulation of unsaturated soil water flow is based on Darcy's equation and it relates flux density q to the hydraulic head and it is represented by :

$$q = -K\frac{\delta H}{\delta Z} \qquad \text{........(3.5)}$$

where,

q = flux density ($L^3\,L^{-2}\,T^{-1}$),

H = hydraulic potential (L),

K = hydraulic conductivity (LT^{-1}), and

z = depth (L)

Darcy's equation is coupled with the conservation of mass principles to derive a continuity equation. The mathematical statement of the law of conservation as applied to water flow through soil can be written as:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta q}{\delta z} \qquad \text{......(3.6)}$$

where,

θ = volumetric moisture content ($L^3\,L^{-3}$), and

t = time (T).

Combining the equations (3.5) and (3.6), yields:

$$\frac{\delta\theta}{\delta t} = \frac{\delta}{\delta z}\left[K\frac{\delta H}{\delta z}\right] \qquad \text{......(3.7)}$$

Substituting the equation (3.4) in (3.7), results in :

$$\frac{\delta\theta}{\delta t} = -\frac{\delta}{\delta z}\left[K\left(\frac{\delta\Psi_m}{\delta z}+1\right)\right] \qquad \text{.....(3.8)}$$

The solution of the equation (3.8) can be found by knowing the soil water retention (Ψ–θ relationship) and hydraulic (K- θ relationship) properties for particular soil. Note that, unlike saturated flow, K does not act as a proportionally constant in the unsaturated flow as it is highly dependent on moisture content.

3.1.5 Soil water retention characteristic

The soil moisture retention characteristic (MRC) is the relationship between the soil moisture content θ and the soil water suction Ψ. MRC can be determined in the laboratory by different methods for a wide range of 0 to 10^7 cm suction. The purpose is to obtain a moisture retention equation (MRE) to estimate soil moisture content for use in the field.

The MRC differs between wetting and drying cycles of the soil. When soil wets from air-dryness or dries from saturation, the characteristics are called wetting or drying curves and this effect is termed hysteresis. The consideration of hysterisis is needed during irrigation or precipitation cycle with greater intervals or soil drying. The present experiments are concerned only with drying or desorption conditions.

There are many approaches available for obtaining MRE in the literature. That of Hassan(1990) has been chosen for the present purpose. This approach employed three-line segments divided between low (0 to -25 cm), medium (-26 to -3165 cm) and higher (-3165 to -10^7 cm) matric potential ranges, with matric potential Ψ_m presented in p^F ($p^F = \log_{10}(-\Psi_m)$ defined by Schofield, 1935) scale and θ in linear scale. The form of the equation is:

$$\Psi_m = e^{a_1(b_1-\theta)} \quad \text{for } \theta_1 \le \theta \le \theta_s \quad \text{.........................(3.9)}$$

$$\Psi_m = e^{a_2(b_2-\theta)} \quad \text{for } \theta_2 \le \theta \le \theta_1 \quad \text{.........................(3.10)}$$

$$\Psi_m = e^{a_3(b_3-\theta)} \quad \text{for } \theta_3 \le \theta \le \theta_2 \quad \text{.........................(3.11)}$$

where, θ_s is the saturation moisture content.

3.1.6 Hydraulic conductivity

The hydraulic conductivity was determined by 'Soil water depletion' and 'Drainage flux' methods (see section 4.2.5) , and estimated by the Gardner (1958) equation. Hassan (1990) obtained good agreement between the hydraulic conductivity obtained from soil water diffusivity (laboratory) and soil water depletion method (crop experiments), and so the soil water depletion method will be taken as basis of our experimental purpose.

3.1.6.1 Soil water depletion method

Hydraulic conductivity was determined from the upward flux and the hydraulic gradient data obtained from the lysimetric crop water use experiment. There was no intervention of downward flux (drainage), i.e. the equilibrium soil moisture profile was attained from the beginning. Equation (3.5) in conjunction with equation (3.4) can be expressed as:

$$q = C_r + M = K(\Psi_m)\left[\frac{d\Psi_m}{dz} + 1\right] \quad ......(3.12)$$

where,

q = upward flux (LT^{-1}),

C_r = capillary rise from water table (LT^{-1}),

M = soil moisture extraction (LT^{-1}), and

K = hydraulic conductivity (LT^{-1}).

If C, M and $\Psi(z)$ are measured, then $K(\Psi)$ can be calculated. The flux in equation (3.12) is not the same for all depths but is equal to the sum of capillary rise from water table and the amount of moisture extracted from below the depth concerned (Hassan, 1990).

3.1.6.2 Drainage flux method

The concept of Green et al. (1986) was used for this method and their form of the equation:

$$\frac{\delta}{\delta t}\int_0^{z_1} \theta(z,t)\, dz = -K(\theta)\left.\frac{\delta H(z,t)}{\delta z}\right|_{z_1} \quad ....(3.13)$$

which is basically one form of equation (3.8).

where,

$\theta(z,t)$ = soil water content ($L^3 L^{-3}$),

$H(z,t)$ = hydraulic potential (L),

$K(\theta)$ = hydraulic conductivity (LT^{-1}),

z_l = vertical distance from soil surface down to the water table (L), and
t = time (T).

The initial condition for equation (3.13) is the soil moisture profile at the moment infiltration at the soil surface ceases. No evaporation is allowed during drainage.

3.1.6.3. Empirical method

From the hydraulic conductivity measurements, data-pairs (Ψ, θ) and (K, θ) are available. An empirical equation can be fitted to the data to represent the hydraulic conductivity.

The hydraulic conductivity was estimated by Gardner's (1958) equation, which as:

$$K(\Psi_m) = \frac{a}{b + \Psi_m^{\ n}} \qquad ....(3.14)$$

where,

$K(\Psi)$ = hydraulic conductivity (LT^{-1}),
Ψ_m = matric potential (L), and
a, b & n = fitted parameters (dimensionless).
K_{sat} = saturated hydraulic conductivity is equal to 'a / b'.

3.1.7 Estimating upward flux

The upward flux can be estimated by combining equations (3.14) and (3.16) as:

$$q = -\left[\frac{d\Psi_m}{dz} + 1\right]\left(\frac{a}{b + \Psi_m^{\ n}}\right) \qquad .....(3.15)$$

Numerical solution of the water flow equation in the unsaturated zone gives the vertical flow between nodes for each time step. Flow at the lower node represents the capillary rise from water table.

The upward flow at each incremental increase in depth can be calculated from equation (3.15) as follows:

$$q_i^j = -\left[K_i^j\left(\overline{\Psi_{m_i}^{\ j}}\right) \frac{\Psi_{m_i}^{\ j} - \Psi_{m_{i+1}}^{\ j}}{Z_{i+1}^j - Z_i^j} + 1 \right] \quad ....(3.16))$$

where $$K_i^j\left(\overline{\Psi_{m_i}^{\ j}}\right) = \frac{a}{b + \left(\overline{\Psi_{m_i}^{\ j}}\right)^n}$$

where the 'i' subscripts refers to the depth and the 'j' refers to time. The value of hydraulic conductivity is computed at each incremental increase in depth using equation (3.14).

The water table depth is constant so that bottom boundary is static. Therefore, $\Psi_m = 0$ and $\theta = \theta_s$ at the water table where θ_s = saturation water content. The flow at the bottom boundary could be up or down depending on the location of the plane of zero-flux. The initial condition of the soil surface at equilibrium is that Ψ_m is equal to the height above the water table, ($\Psi_m = -Z$ when $t = 0$).

3.2 Solute transport

3.2.1 Convective transport

Ions are carried by moving water. This transport, related exclusively to water movement, is called convective, viscous, or mass flow. The dissolved ions tend to move with the water whenever soil water is dynamic, such as during infiltration, redistribution or evaporation.

The flow of soil water carries with it a convective flux of solutes, J_c, proportional to their concentration C and therefore;

$$J_C = qC = C\left(-K\frac{dH}{dx}\right) \quad .....(3.17)$$

where,

q = flux density ($L^3 L^{-2} T^{-1}$),
C = concentration of solute in solution (ML^{-3}),
J_c = flux of solute ($ML^{-2}T^{-1}$),
K = hydraulic conductivity (LT^{-1}),

H = hydraulic potential (L), and
x = distance (L).

3.2.2 Diffusive transport

In the solution phase, particles move at random by thermal motion. The flow rate of solute particles is proportional to the concentration gradient (Fick's first law):

$$J_d = -D_p \frac{dC}{dx} \qquad \text{......(3.18)}$$

Where,
J_d = diffusive flux ($ML^{-2}T^{-1}$),
D_p = molecular diffusion coefficient in porous media (L^2T^{-1}), and
C = concentration of solute (ML^{-3}), and
x = distance (L).

3.2.3 Dispersive transport

Dispersive transport can be described by the equation:

$$J_h = -D_m \frac{dC}{dx} \qquad \text{.....(3.19)}$$

where,
D_m is the mechanical dispersion coefficient (L^2T^{-1}) and is assumed to be a function of pore velocity v_p, given by :

$$D_m = \lambda\left(V_p\right)^{n_1} \qquad \text{.......(3.20)}$$

where λ is the dispersivity and n_1 is an empirical constant, roughly equal to 1 (Bear, 1972). The mechanical dispersion coefficient includes the effect of solute spreading due to the nonuniform distribution of the microscopic velocity within pores and the intricate geometry of microscopic streamlines.

The coefficient of hydrodynamic dispersion (D_c) has two components: one velocity dependent, D_m, and the other velocity independent, which is the D_p (Nielsen et al., 1981). The expression for hydrodynamic dispersion (D_c) is

$$D_c = D_p + D_m \qquad \text{......(3.21)}$$

The total flow of solute is the sum of the dispersive and convective components and combining the equations (3.17), (3.18), (3.19) & (3.21) becomes:

$$J_{c-d} = -D_c \frac{dC}{dx} + C\left(-K\frac{dH}{dx}\right) \qquad \text{......(3.22)}$$

where, J_{c-d} is the convective-dispersive transport of solute ($ML^{-2}T^{-1}$).

3.2.4 Soluble salt determination from soil water extracts

The term 'soluble salts' refers to the major dissolved inorganic solutes. Soil salinity is described and characterized in terms of the concentrations of soluble salts. The management of saline soils is evaluated from measurements of such concentrations.

Soluble salts can be determined or estimated from measurements made i) on soil samples, ii) on solution collected insitu, iii) in soil using salinity sensors of any kinds. Collection of solution is more convenient for salt analysis but is limited to relatively wet soil conditions. Soil sample extracts give relative comparison only (Rhoades, 1982).

Szabolcs (1985) stated that, although study of the salt regime supplies important data on the dynamics of salt affected soil, it does not present in itself any further information. Therefore, the next step is to compare the salt contents of the soil at given times and to express them in salt balances. He considered the following data to be needed to establish the salt balance:

a) Total amount of soluble salts at the beginning and at the end of the observation.
b) The increase of soluble salt contents during the observation. and
c) The decrease of soluble salt contents during the observation.

The literature suggests that there are two main approaches dealing with determination of salinity. The first approach is a simple salt balance. The second approach uses mathematical models of the dynamics of soil moisture and salt movement (Ismail , 1990).

3.2.5 Salt balance

The simplified salt balance was used for this experiment. The assumptions and conditions of the salt balance equation prevailing in this lysimeter experiment are as follows:

a) There is no surface water application either by rain or irrigation,
b) The drainage is nil because the soil profile was in equilibrium before starting the experiment,
c) The mass of salt removed by the crop is negligible, and
d) The mass of salt precipitated in the soil after evapotranspiration was completely dissolved in the soil solution.

The salt balance equation for this experiment is written as follows:

$$V_{sm} C_{sm} + V_{cr} C_{cr} = \Delta S_{rz} \qquad \text{..........(3.23)}$$

where,

V_{sm} = volume of soil moisture evaporated from the root zone (L^3),
C_{sm} = concentration of solute in soil moisture (ML^{-3}),
V_{cr} = volume of upward flow into the root zone (L^3),
C_{cr} = concentration of solute in the upward flow solution (ML^{-3}), and
ΔS_{rz} = change solute content in the root zone (M).

Osmotic suction was calculated according to the widely used equation derived by USDA (1954) as follows:

$$\Psi_o = -\,0.036 \times EC \qquad \text{....... (3.24)}$$

where,

EC = electrical conductivity of the solution (dS m^{-1}),
Ψ_o = osmotic potential (MPa).

3.3 Water uptake by roots

The usual macroscopic approach of quantifying water uptake by roots is represented by a volumetric sink term S, which is simply added to the continuity equation (3.8) for soil water flow, as:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta q}{\delta z} - S \qquad .....(3.25)$$

where,

θ = volumetric water content ($L^3\ L^{-3}$),

t = time (T),

q = soil water flux ($L^3\ L^{-2}\ T^{-1}$),

S = water uptake by roots (sink term) taken positive from soil to roots (T^{-1}), and

z = vertical distance (downward directed) from soil surface (L).

Equation (3.25) may directly be applicable in nonsaline situations, elaborating S in terms of root distribution function. The equation implies that the hydraulic conductivity in the soil medium is equivalent to the hydraulic conductivity from soil to roots. But this concept cannot be acceptable in saline conditions. The hydraulic conductivity through the soil may be unaffected or affected due to solute presence (depends on soil constituents), but the conductivity from soil to roots (which acts as a semi-permeable membrane) will be proportionately restricted by the degree of salinity irrespective of soil constituents. Therefore, the term S needs to incorporate an additional reduction factor to account for salinity decreasing the actual water uptake. This reduction factor or rate coefficient (R_o) can be determined by taking the concept of available plant uptake potential as described in section (3.1.2). Including the reduction factor in the equation (3.25), yields:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta q}{\delta z} - S.R_o \qquad .....(3.26)$$

where,

R_o is the reduction factor in water uptake due to osmotic potential in the root zone soil water.

The formulation of R_o in the present experimental context is described in Chapter 6.

Part 2: LYSIMETER EXPERIMENTS

Chapter: 4

MATERIALS AND METHODS

4.1 Materials

4.1.1 Lysimeters

A lysimeter is an instrument for measuring the actual water use by plants. The purpose of a lysimeter is to maintain a controlled representative environment for the measurement of water states or process. The lysimeter typically confines the side and base of the soil and water reservoir leaving the surface as representative as possible of the undisturbed surroundings.

The present lysimeters are a PVC (polyvinyl chloride) container having a diameter of 106 cm and 145 cm in height. They were filled with soil up to 140 cm height except the bottom 10 cm layer of fine gravel mixed with coarse sand. The texture of the lysimeter soil was sandy silt loam (Hassan, 1990) with a composition of silt (59.3%), sand (37.2%) and clay (3.5%). The average bulk density of the soil was 1.54 ± 0.005 g cm^{-3}.

Each of the three lysimeters was instrumented with a Mariotte siphon, tensiometers, soil solution extraction device and gypsum blocks (Fig. 4.1).

4.1.2 Tensiometers

The measurement of soil moisture tension can allow the determination of moisture contents. Tensiometers are commonly used to measure soil water suctions up to approximately 0.8 atmosphere. Each of the tensiometers consisted of a porous ceramic cup, with an air tight connecting tube leading to the pressure measuring device, a mercury manometer. Tensiometers were inserted 5 & 15 cm below the soil surface and then at 15 cm intervals to the water table. Tensiometers were filled with de-aired distilled water and periodically refilled to purge air. Measurements of suctions were made daily at 9.00 a.m.

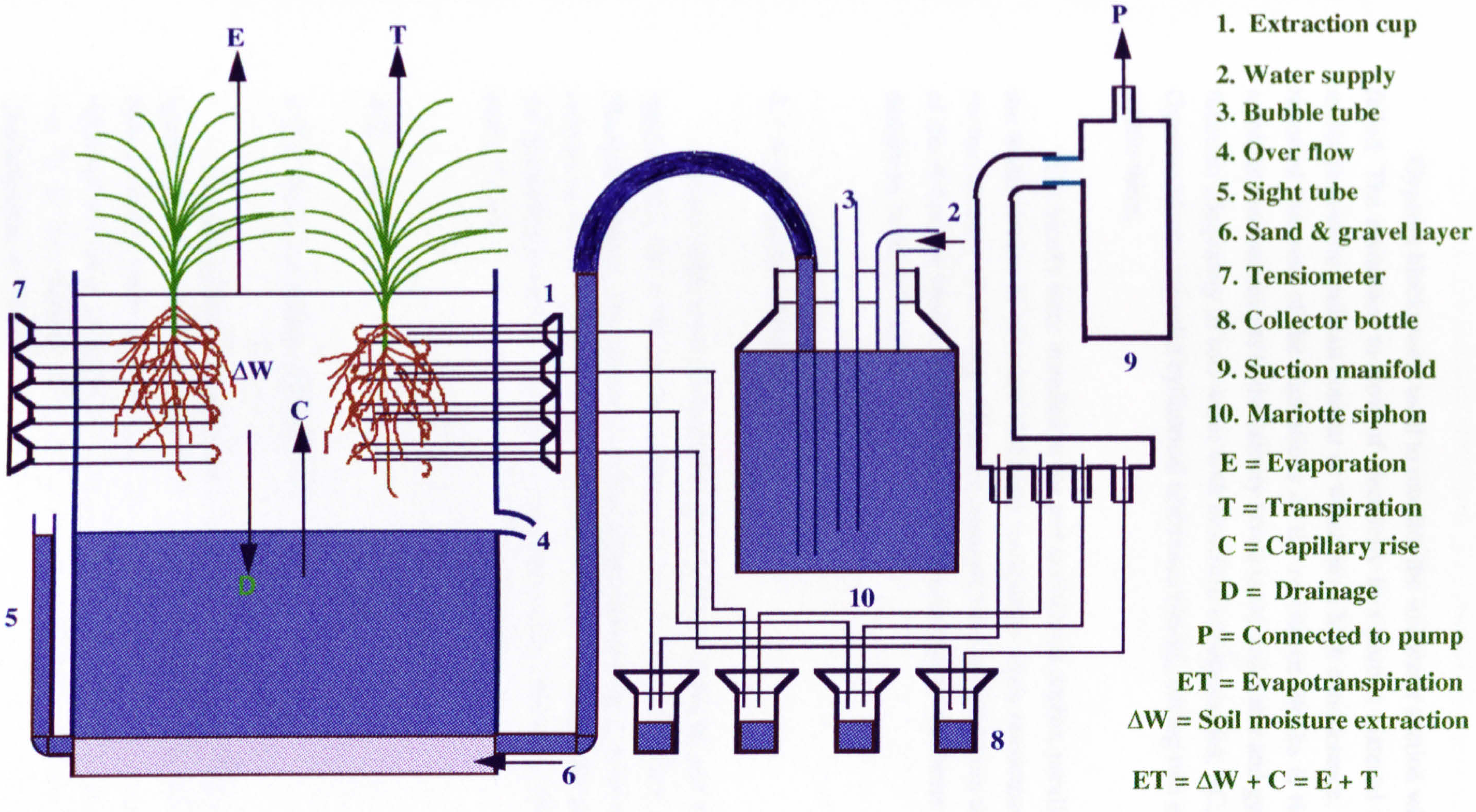


FIG. 4.1: SCHEMATIC OF LYSIMETRIC SOIL -WATER-PLANT SYSTEM THAT OPERATES ON THE MARIOTTE SIPHON PRINCIPLE.

4.1.3 Gypsum blocks

Gypsum blocks were used to monitor the soil water suction when the tensiometers failed. The resistance to flow of electricity in a porous material is a function of the moisture content and salt content as water has a high conductance. Salazar et al.(1984) reported that one of the limitations of the resistance blocks is the sensitivity of the resistance measurements to the salinity levels in the soil water and gypsum blocks usually function adequately in soil-water with electrical conductivities, EC, up to 2000 dS m^{-1}. Gypsum blocks are solid cylindrical resistance blocks, having two electrodes embedded within them.

The blocks were installed in the soil at different depths, parallel to the positions of the tensiometers, to measure electrical resistances when tensiometers failed to record suction because of air entry. Moisture contents were inferred from the calibration curves of the resistance blocks determined in the laboratory on the same soil with same bulk density as in the lysimeters.

4.1.4 Mariotte siphons

Water tables were controlled at three levels, viz. 60, 90 and 120 cm, for lettuce & ryegrass'92, and at 90 cm for ryegrass'93 below the soil surface in the lysimeters by Mariotte siphons. The siphons supplied water through the bottom of the lysimeters soil column as shown in Fig. 4.1. The siphon reservoirs measured the water that moved into the lysimeters to replace that used by the plants in the lysimeters. The records were taken every 5 days.

4.2 Methods

4.2.1 Soil moisture characteristic

The soil moisture characteristic equations of Hassan (1990) were used because his lysimeters with same soil have been utilized for this investigation. In the year 1993, before starting the crop experiment, the top 15 cm soils from each of the lysimeters were replaced with new but same textured soils by maintaining the same bulk density (1.5 g cm^{-3}) as was before. After harvesting the ryegrass'93 crops, the soil moisture characteristic of that soil was determined to check the water retention properties if any change happened. Twelve undisturbed but resampled soil cores (four for each of the three lysimeters) in brass containers, 40 mm in diameter and 25 mm in height, were used. The resampling was done from the collected undisturbed soil cores having 75 mm in diameter and 50 mm in height. The soil moisture characteristics were determined in the laboratory following the desorption process. Haines method was used for 0 to 100 cm suction range and 100 to 5,000 cm suction range by pressure plate apparatus. For every range interval considered in between 100 to 5,000 cm suction range including zero suction, moisture content contents were determined by gravimetric method. Finally, suctions were related to respective volumetric moisture contents and plotted as p^F against volumetric moisture content.

4.2.2 Gypsum block calibration

Gypsum blocks were calibrated in a constant temperature room at 20 ^{o}C. Disturbed core (10 cm in diameter and 15 cm in height) samples having uniform density of 1.55 g cm^{-3} were used. Two blocks were installed in a single core at 5 & 10 cm deep. Evaporation loss along with electrical resistance was recorded and resistance as a function of volumetric content was plotted. Temperatures measured for soil solution during cropping periods was assumed to be the same temperature within the gypsum blocks. Hence temperature correction was made in determining the moisture content from the calibration curve.

4.2.3 Water use experiments.

A perennial medium-rooted and a seasonal shallow-rooted crop, ryegrass and lettuce, respectively were selected for this investigation to assess the potentials of growing a crop only with saline water sub-irrigation. The lysimetric experiments on crop water use with different combinations of water table depths and salinities were conducted from 1991 to 1993 in Moorbank glasshouse, University of Newcastle upon Tyne. The monthly mean temperatures during the growing seasons in the glasshouse are shown in Table 4.1. The weekly temperature profiles in the different cropping seasons are also shown in Fig. 4.2. The salient features of the different experiments are shown in Table 4.2.

At the beginning of the each experiment, the soils were saturated with saline water of the respective electrical conductivity (shown in Table 4.2) and the soil solutions were supplied from the bottom of the lysimeters. Then the saturated soils were left covered for two days and then drained for two days. The process of alternate saturating and draining

Table 4.1: Temperatures (°C) in the glasshouse during different cropping seasons.

Month	Lettuce'91		Ryegrass'92		Ryegrass'93	
	Tmax	Tmin	Tmax	Tmin	Tmax	Tmin
Mar.	—	—	19.8	8.0	—	—
Apr.	—	—	24.8	8.8	24.5	9.8
May	—	—	29.0	11.4	28.5	10.2
Jun.	—	—	32.5	13.0	31.6	11.9
Jul.	—	—	31.7	13.1	30.5	10.9
Aug.	29.2	13.9	31.5	12.2	29.6	11.3
Sept.	22.7	10.6	26.0	10.7	20.7	11.1
Oct.	16.3	10.1	19.7	8.4	18.4	9.4
Nov.	—	—	19.5	7.5	16.5	8.8

was continued until the soil profiles attained the desired salinity. The soil profile salinities were checked from the collected soil solution as well as from soil samples at planting. The saturated soils were allowed to drain for a specified periods of time (Table 4.2) to achieve the equilibrium soil moisture profiles before planting. The soil was loosened on the planting dates, to a depth of 20 cm. On the same day seeds were sown maintaining a row-to-row distance of 17.5 cm for lettuce, 3.0 cm for ryegrass'92 and 2.5 cm for ryegrass'93. Before seeding, some soil samples were collected and analysed to determine the amounts of essential nutrient for each of the crops, to guide fertilizer application.

For lettuce, 15 days after sowing, N, P & K were applied at the rate of 125 kg N ha^{-1}, 200 kg P_2O_5 ha^{-1} and 125 kg K_2O ha^{-1}, respectively. The plants were thinned keeping the plant to plant distance of about 15 cm, 25 days after sowing. There were 32 plants in each lysimeter. The lettuce plants were uprooted after 90 days from sowing, using a hand shovel. Fresh top weight, root weight and root penetration depth were recorded for each plant. Plants were dried after harvesting at 60°C for 48 hours to determine the dry matter content.

For ryegrass'92, it took about 14 days to complete emergence and the lysimeters were left covered the first 7 days until the germination started. 32 days after sowing, N,

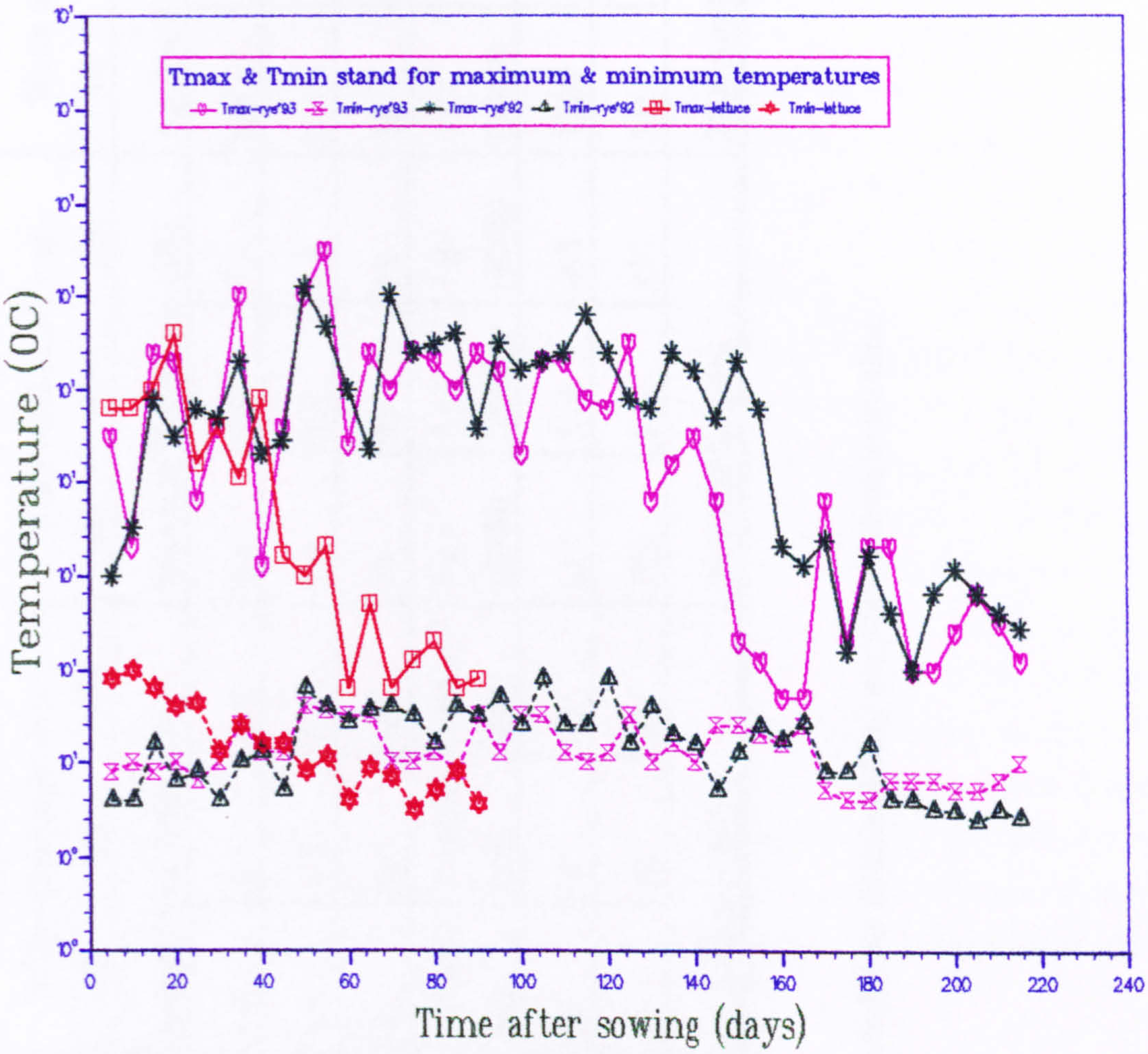


Fig. 4.2: Temperature profile in different cropping seasons

Description of components	Lettuce 1991 Water table depth (cm)			Ryegrass 1992 Water table depth (cm)			Ryegrass 1993 Water table depth (cm)		
	60	90	120	60	90	120	90	90	90
Time periods of crop grown	August to October, 1991			March to November, 1992			April to November, 1993		
Water table salinity (dS m^{-1})	4.5	4.5	4.5	9.4	9.4	9.4	0.4	7.5	15.0
Presaturation soil water salinity (dS m^{-1})	4.5	4.5	4.5	9.4	9.4	9.4	0.4	7.5	15.0
Draining period (days)	90	90	90	75	75	75	25	25	25
Seeding rate, g m^{-2} or (No. of seeds)	3 seeds / hole	3 seeds / hole	3 seeds / hole	3 g / (1000)	3 g / (1000)	3 g / (1000)	3 g / (1350)	3 g / (1350)	3 g / (1350)
Established plant population (No. m^{-2})	32	32	32	614	519	490	1229	1067	432
Cropping periods (days)	90	90	90	223	223	223	216	157	213
Planting dates	1st August, 1991			26th March, 1992			27th April, 1993		

Table 4.2: Salient features of experimental conditions distinguishing from one season to another.

P & K were applied at the rate of 120 kg N ha^{-1}, 40 kg P_2O_5 ha^{-1} and 30 kg K_2O ha^{-1} respectively and 100 kg N ha^{-1} was applied after each cut of grass. 1.1 mm of water was applied with every fertilizer application. The first and second cuts were taken after 62 and 83 days from sowing and the subsequent cuts were at intervals of 28 days.

Before starting the ryegrass'93 experiment, the soil in the lysimeters was leached thoroughly with water to eliminate as much salt as possible. For resalinization of the soil profiles, the process of saturation and draining was continued for a month. It took about 10, 12 & 15 days to complete germination for 0.4, 7.5 & 15.0 dS m^{-1} treatments, respectively, and the lysimeters were left covered for the first 5 days until germination started. 12 days after sowing, N was applied at the rate of 120 kg N ha^{-1} and 100 kg N ha^{-1} was applied after each cut. K was applied at the rate of 60 kg/ha K_2O for 0.4 & 7.5 dS m^{-1} treatments and 80 kg ha^{-1} for the other treatment. The rate of P was 20 kg ha^{-1} for all. 2.2 mm of water was applied at the first fertilization and then 1.1 mm at the following fertilizer applications. The variation in K_2O appilcation was due to the variation of the availability in the soil in different lysimters. The first cut was after 45 days and the subsequent cuts were spaced at intervals of 2, 4 and 8 weeks for 0.4, 7.5 and 15.0 dS m^{-1} treatments respectively. The reason for choosing the different intervals of cut was to allow the maximum possible vegetative developments in the salinity treatments. The detailed crop height during different cuts of both ryegrass crops are shown in Table 4.3. During the entire period of all experiments, the soil solution of the same electrical conductivity was supplied through the Mariotte bottles to maintain the constant water table and replace the capillary water taken up during evapotranspiration.

4.2.4 Salinity measurement

The soil solution was extracted from each depth by a vacuum pump through the ceramic cups. The extraction devices, i.e. the ceramic cups, were installed at the same depths as the tensiometers. Solution was extracted at 0 and 30 days after sowing, and subsequently on the day following each cut. Soil samples were taken from the top 5 cm to determine the electrical conductivity when and where the vacuum pump was not capable of extracting solution as the soil became drier. Around 1 cm^3 of solution was collected each time from each soil depth and the sample solution was diluted to 10 cm^3 with distilled water to determine the electrical conductivity.

Table 4.3: Average plant height (cm) during different cutting times of the ryegrass crops.

No. of cuts	Ryegrass'92 Water table depth (cm)			Ryegrass'93 Water table salinity (dS m^{-1})		
	60	90	120	0.4	7.5	15.0
1st	22.5	18.5	15.5	25.0	16.0	10.0
2nd	24.5	17.5	13.5	27.5	21.0	12.5
3rd	29.0	22.5	21.0	32.5	21.5	15.5
4th	35.0	22.5	22.5	30.0	22.5	13.5
5th	27.5	24.5	22.5	28.5	17.5	—
6th	25.0	20.0	22.5	28.0	—	—
7th	20.0	17.5	21.0	27.5	—	—
8th	—	—	—	25.0	—	—
9th	—	—	—	26.0	—	—
10th	—	—	—	17.5	—	—
11th	—	—	—	14.0	—	—
12th	—	—	—	14.0	—	—

After harvesting each crop, triplicate soil samples were collected down to the water table to measure the salt content and the moisture distribution gravimetrically. An amount of 150 g oven-dry soil was used for each determination of salt content . Each sample was placed in a bottle and distilled water added maintaining the soil:water ratio of 1:1. Then each bottle containing soil solution was shaken vigorously by hand for one minute four times at 30-minute intervals. Solution from the supernatant was used to determine electrical conductivity, and hence salt content.

4.2.5 Hydraulic conductivity

4.2.5.1 Drainage flux method

This method of determining unsaturated hydraulic conductivity was based on an analysis of transient soil-water content and hydraulic-head profiles during vertical drainage. Hydraulic conductivity was determined using the method described by Green et al.(1986) using equation (3.13). The saline water table was raised to the soil surface from the base of the lysimeters, so that the soil became saturated without air entrapment.

Then draining was started and the soil surface was left covered to prevent evaporation. Tensiometers were read every 24 hours at all depths. Hydraulic heads (H) from these tensiometric data were plotted against time, and a smooth curve was drawn through the points. From this curve of H versus t, H values were plotted for respective tensiometer depths (z) and another smooth curve was drawn. The hydraulic head gradient, $\delta H/\delta z$ was determined from the smooth curve of H versus z. The matric potential (Ψ_m) was determined from the H versus z curve and $\Psi_m(z)$ values were recorded.

Soil water content profiles were not measured directly. $\Psi_m(z)$ data points were converted to water content values $\theta(z)$ using the soil moisture characteristic equations determined in the laboratory. $\theta(z)$ values were plotted against time (t), and a eye-fitted smooth curve was drawn for $\theta(z)$ versus t. Using the water content profile for a given time, the integral $\int\theta(z,t)dz$ was estimated by a trapezoidal approximation.

A smooth curve was eye-fitted through the data of $\int\theta(z,t)dz$ versus time and the derivatives $\delta[\int\theta(z,t)dz]/\delta t$ at different times were calculated. The time derivatives are the fluxes at fixed positions and times.

Unsaturated hydraulic conductivity values were then calculated by dividing the fluxes calculated above with the hydraulic gradients at the same positions and time.

4.2.5.2 Soil-water depletion method

Hydraulic conductivity of the soil in the lysimeters was calculated from the soil-water depletion measurements in the lysimeters following the procedure of Hassan (1990). In this method, hydraulic conductivity was determined from the flux and the hydraulic gradient data when there is loss of water from evapotranspiration. Equation 3.12 was used to calculate the hydraulic conductivity. Hassan (1990) noted that, for short dry periods, with continuous flow upwards, one may use arithmetic average values of matric potential (Ψ_m) and upward flux (q) to calculated hydraulic conductivity for other calculations.

To apply the method, it was required to know the amount of moisture depleted from the soil and the capillary rise from water table. Moisture extraction was estimated from tensiometer readings. Matric potentials were converted to volumetric moisture contents using the soil moisture characteristics equations. Depths of moisture depleted in each layer were calculated for periods of five days and the water table contribution for the same period was measured.

The total vertical flow at all depths for each particular period was calculated as the sum of capillary rise from the water table and the amount of moisture extracted from below the depth concerned. The upward flow, q, with respect to depth per day was calculated by dividing the total vertical flow at that depth by the time period. Average values of q and H were calculated for successive layers. From this, the potential gradient was determined and the hydraulic conductivity calculated assuming Darcy's law to hold.

6.2.6 Root sampling and separation.

After final harvesting of the ryegrass crops, the roots were sampled with a root sampler 7.5 cm in diameter and 15.0 cm deep. Triplicate samples were taken from each lysimeter down to the water table. Each core of the root sample was soaked overnight with a solution of water and tetra-sodium pyrophosphate (hydrated) at the rate of 3 g per litre of water. Then the roots were dispersed in water by manual shaking on a sieve of 2 mm wire-mesh and thus separated from soil. Finally, the samples were oven-dried at 60 ^{o}C for 48 hours to obtain the dry mass.

Chapter 5

RESULTS AND DISCUSSION

5.1 Soil moisture extraction

Soil water status resulted from the combined effect of soil moisture depletion and capillary rise from the water table.

The moisture contents at different depths of soil profile were determined from the measured matric potentials of soils. The time and depth dependent moisture content profiles are presented in Figs. 5.1.1(a),(b) & (c); 5.1.2(a),(b) & (c) and 5.1.3(a),(b) & (c) for lettuce, ryegrass'92 and ryegrass'93 respectively. The upper layer (0~15 cm) showed rapid moisture depletion while deeper layers (below 15 cm) had a slower change because of upward flux from the lower soil depths and the water table. Moisture depletion from all the soil layers of ryegrass'92 treatments stopped after 167 days from the sowing date while, in the ryegrass'93 treatments, the soil moisture depletion stopped after 150 days. The change of moisture in the lower layers was caused by the upward flux to the upper layer to satisfy the crop water demand. The Figures also show that there was good agreement between the measured values of the moisture content at harvest by tensiometers or gypsum blocks and gravimetric method. Fig. 5.1.3(d) also shows that the matric stress in salinity treatments was much smaller than the nonsaline treatment. The comparison of matric suction trends for variation in water table depths are shown in Figs. 5.1.1(d) and 5.1.2(d).

Cumulative moisture extraction patterns in Figs. 5.1.4(a),(b) & (c); 5.1.5(a),(b) & (c) and 5.1.6(a),(b) & (c) show that soil moisture was extracted from almost the entire soil profile above the water tables. The moisture extraction zone in all treatments extended below the maximum rooting depth. From the very beginning of the cropping period, moisture extraction occurred from the whole of the soil profile 15 cm above the water table. Figs. 5.1.4(d), 7.2.5(d) and 5.1.6(d) compare the moisture extraction patterns from different water-table treatments. Fig. 5.1.5(d) shows that there was no difference in the maximum possible soil moisture depletion in the same salinity treatments though there was a difference in water table depths, while Fig. 5.1.6(d) shows that there was a big difference in the soil moisture depletion between the nonsaline and saline treatments.

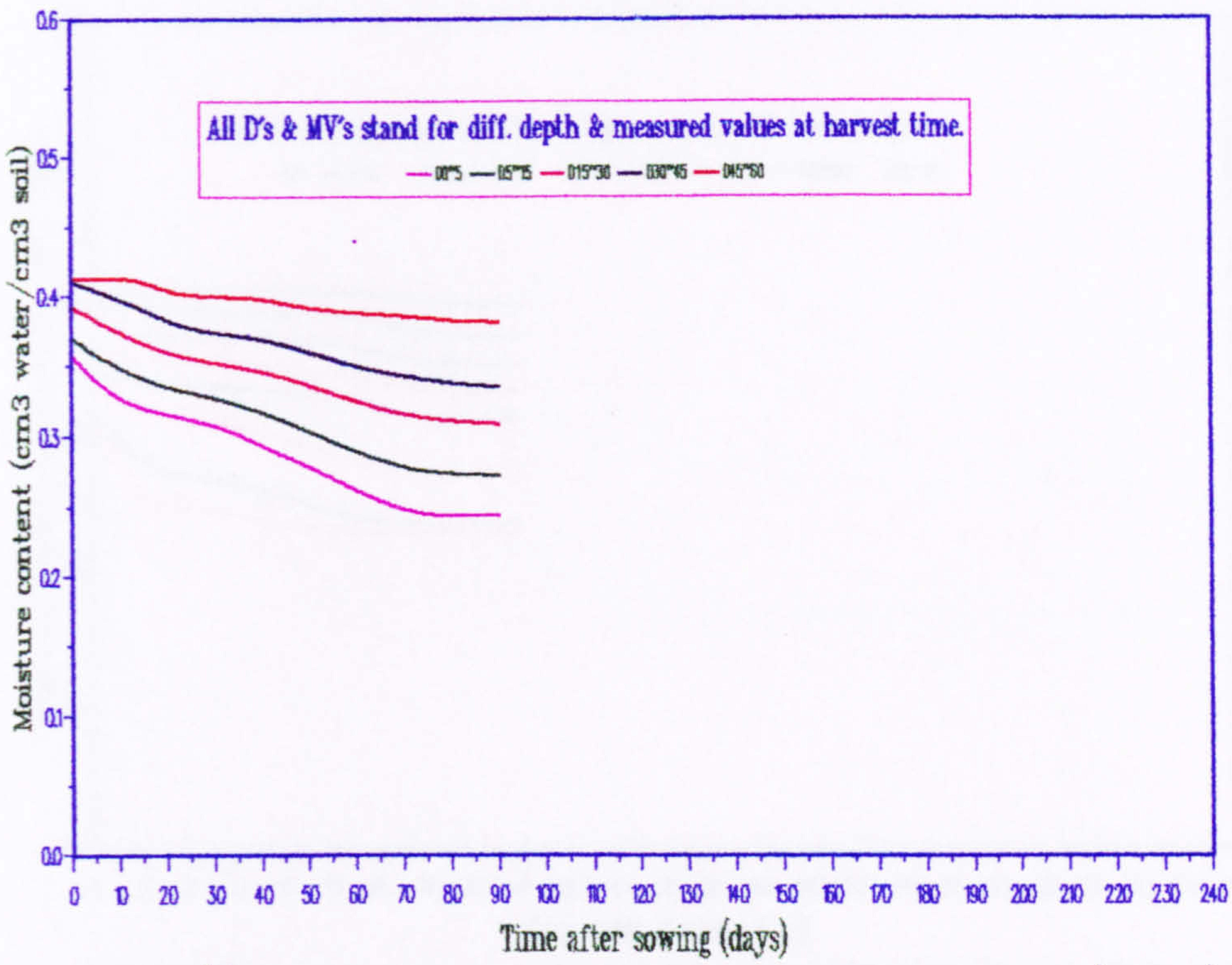


Fig. 5.1.1(a): Soil moisture content Vs time for WT-60 lysimeter (lettuce)

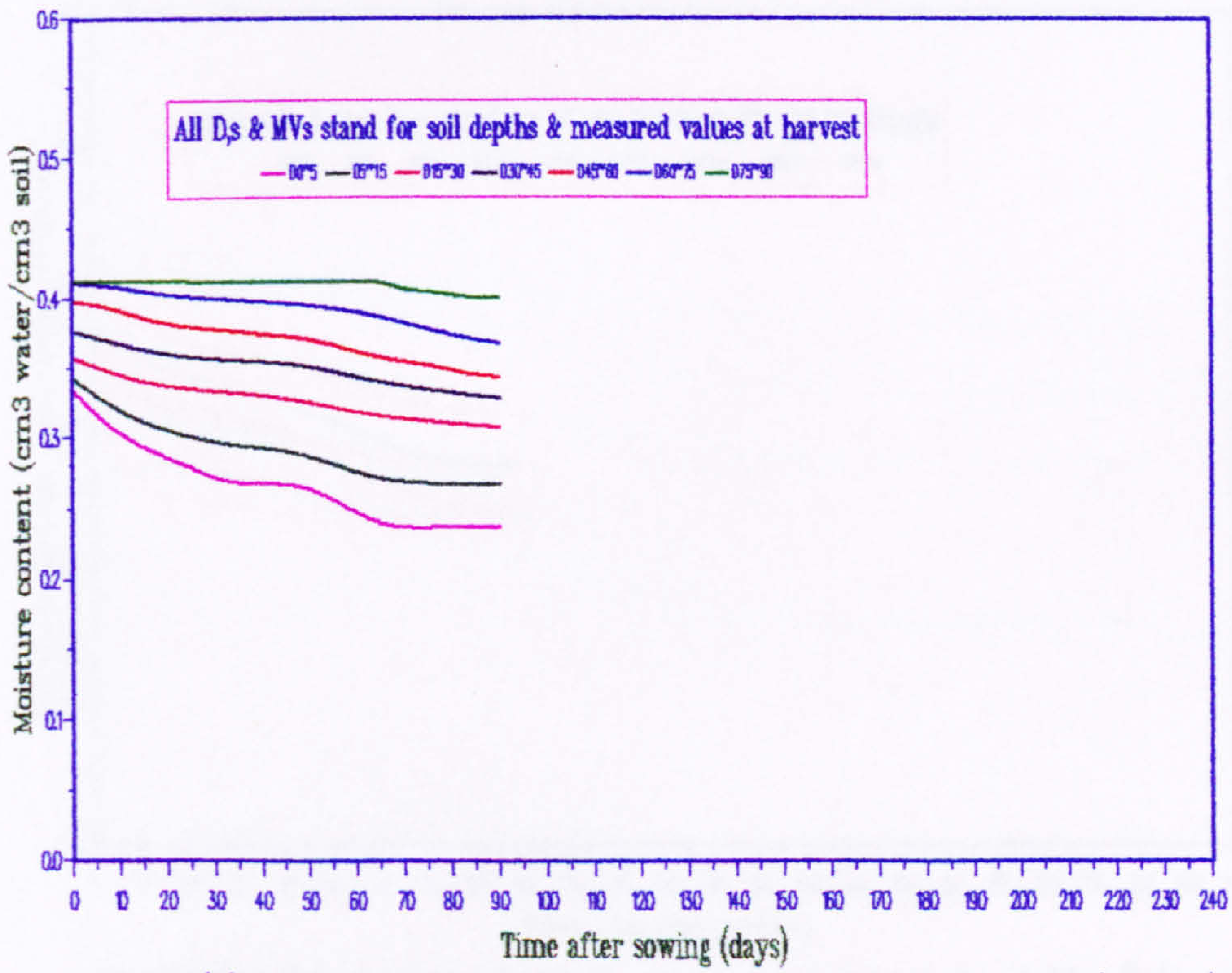


Fig. 5.1.1(b): Soil moisture content Vs time for WT-90 lysimeter (lettuce)

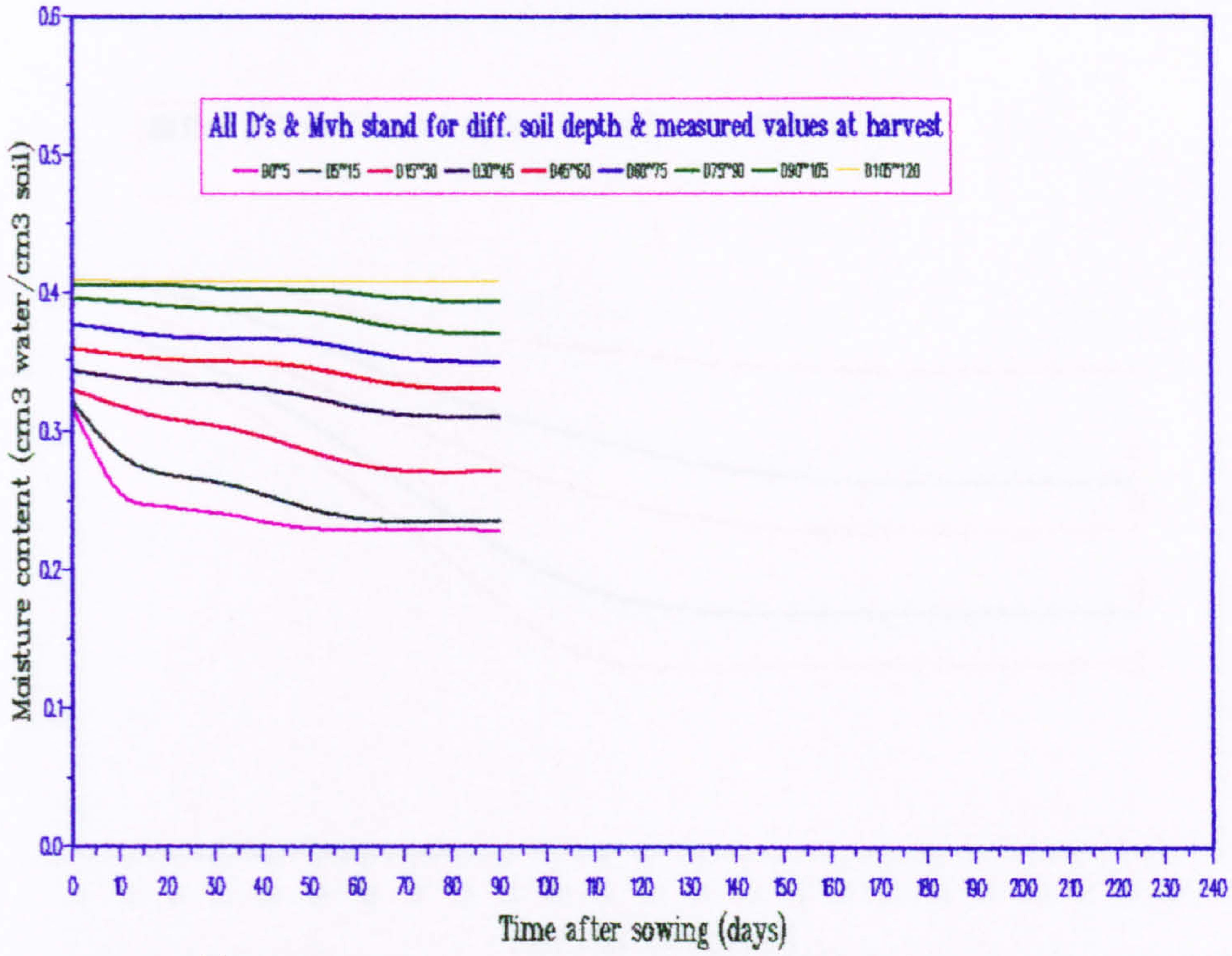


Fig. 5.1.1(c): Soil moisture content Vs time for WT-120 lysimeter (lettuce)

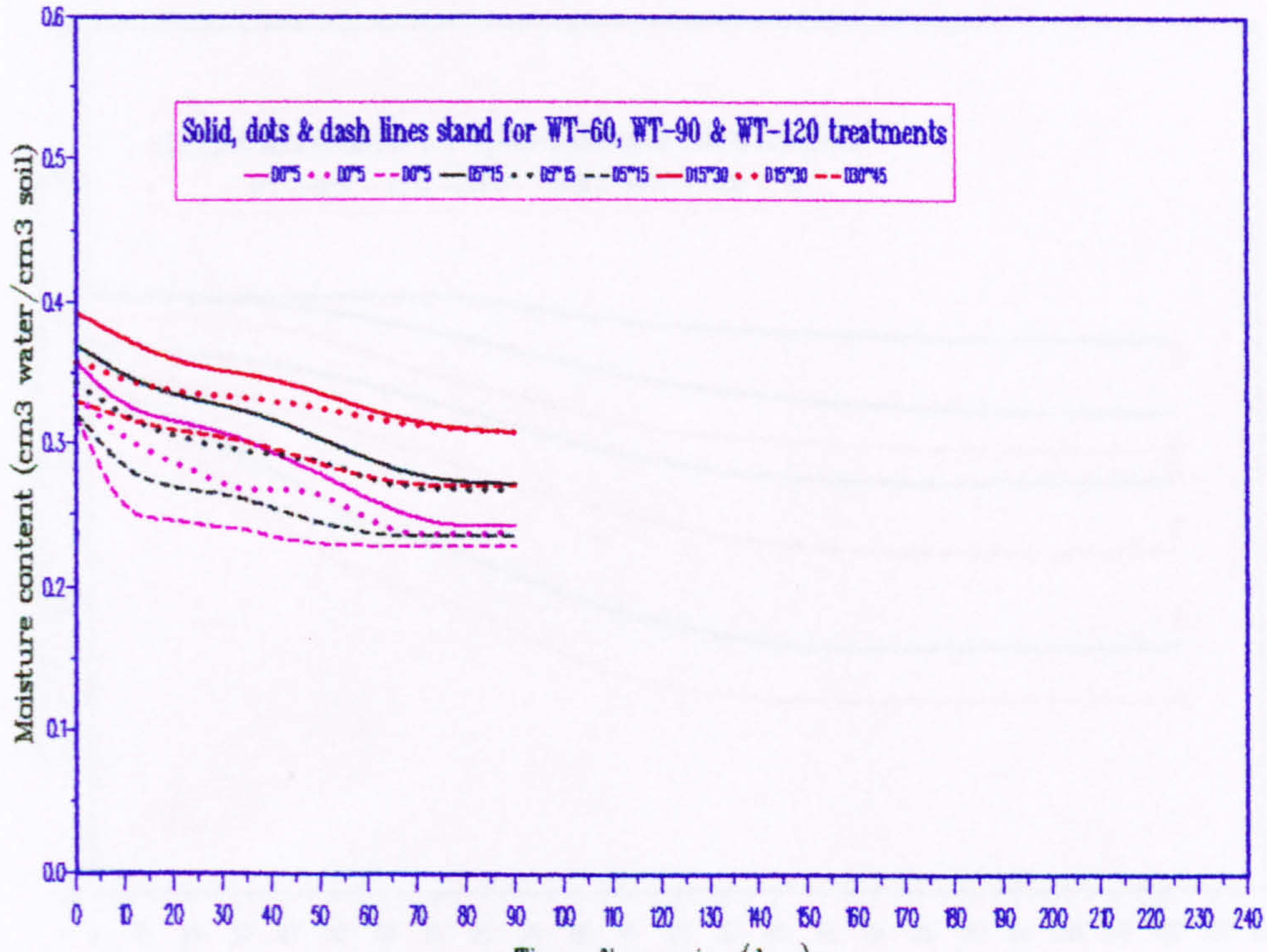


Fig. 5.1.1(d): Comparison of moisture content among water tables (lettuce)

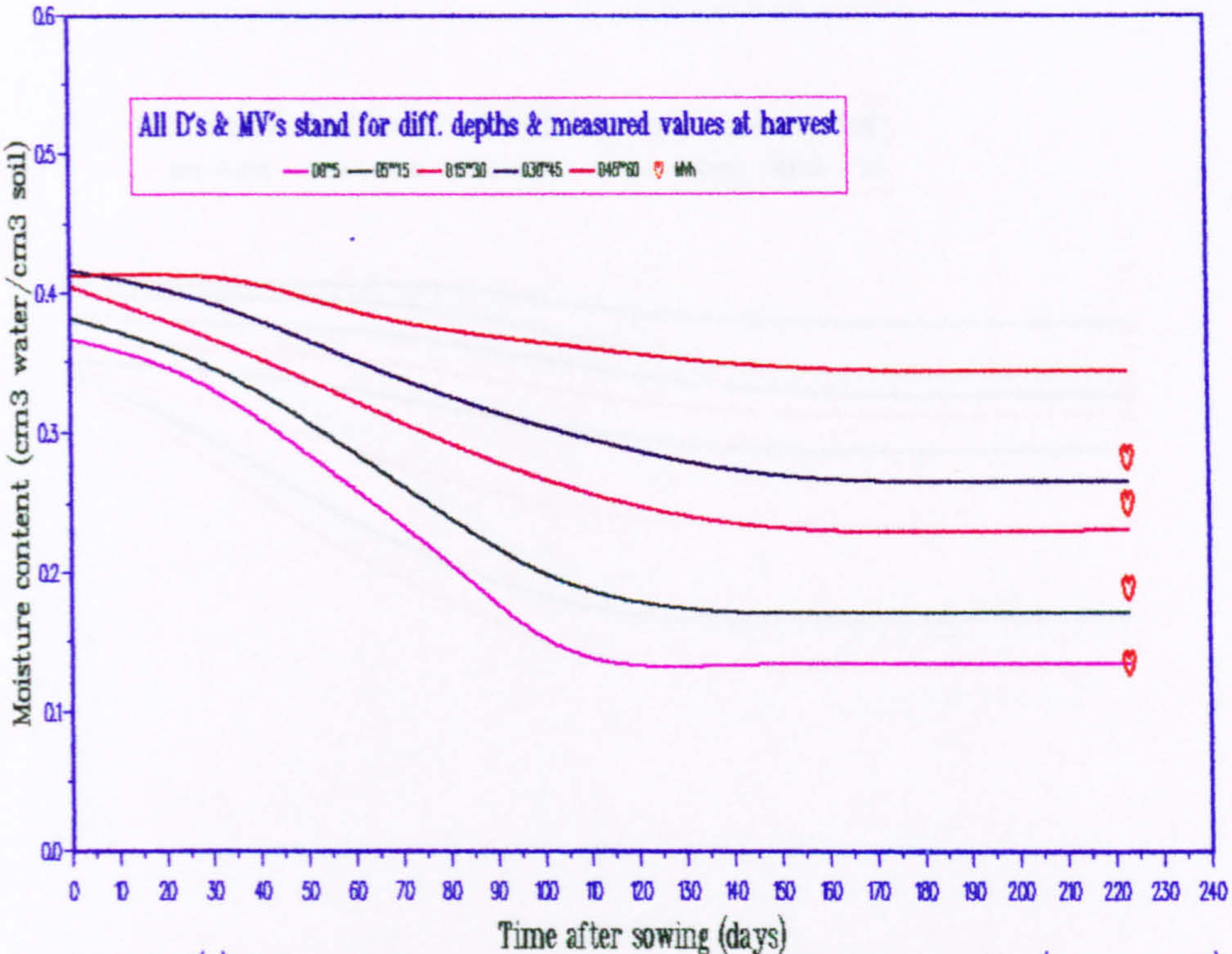


Fig. 5.1.2(a): Soil moisture content Vs time for WT-60 lysimeter (ryegrass'92)

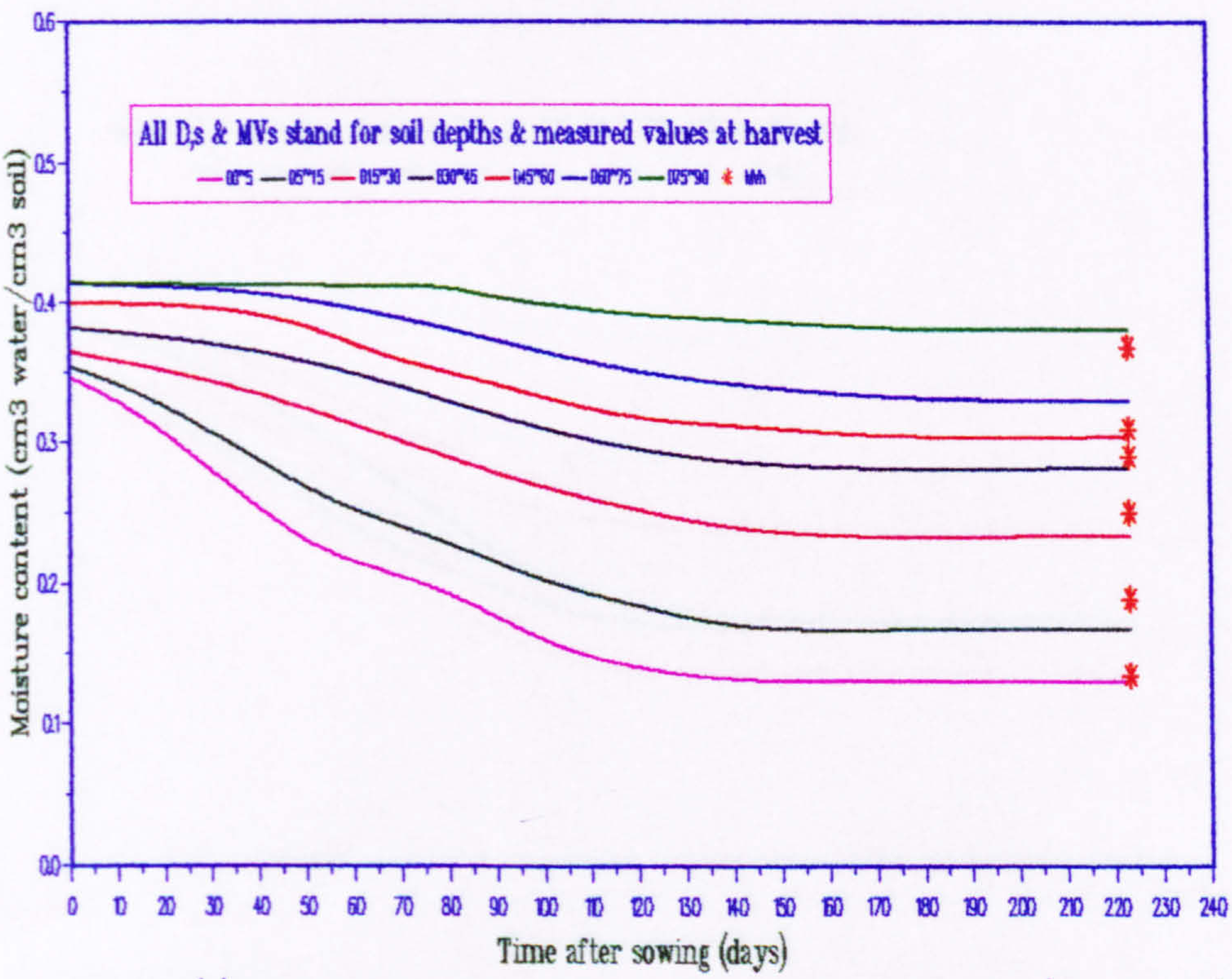


Fig. 5.1.2(b): Soil moisture content Vs time for WT-90 lysimeter (ryegrass'92)

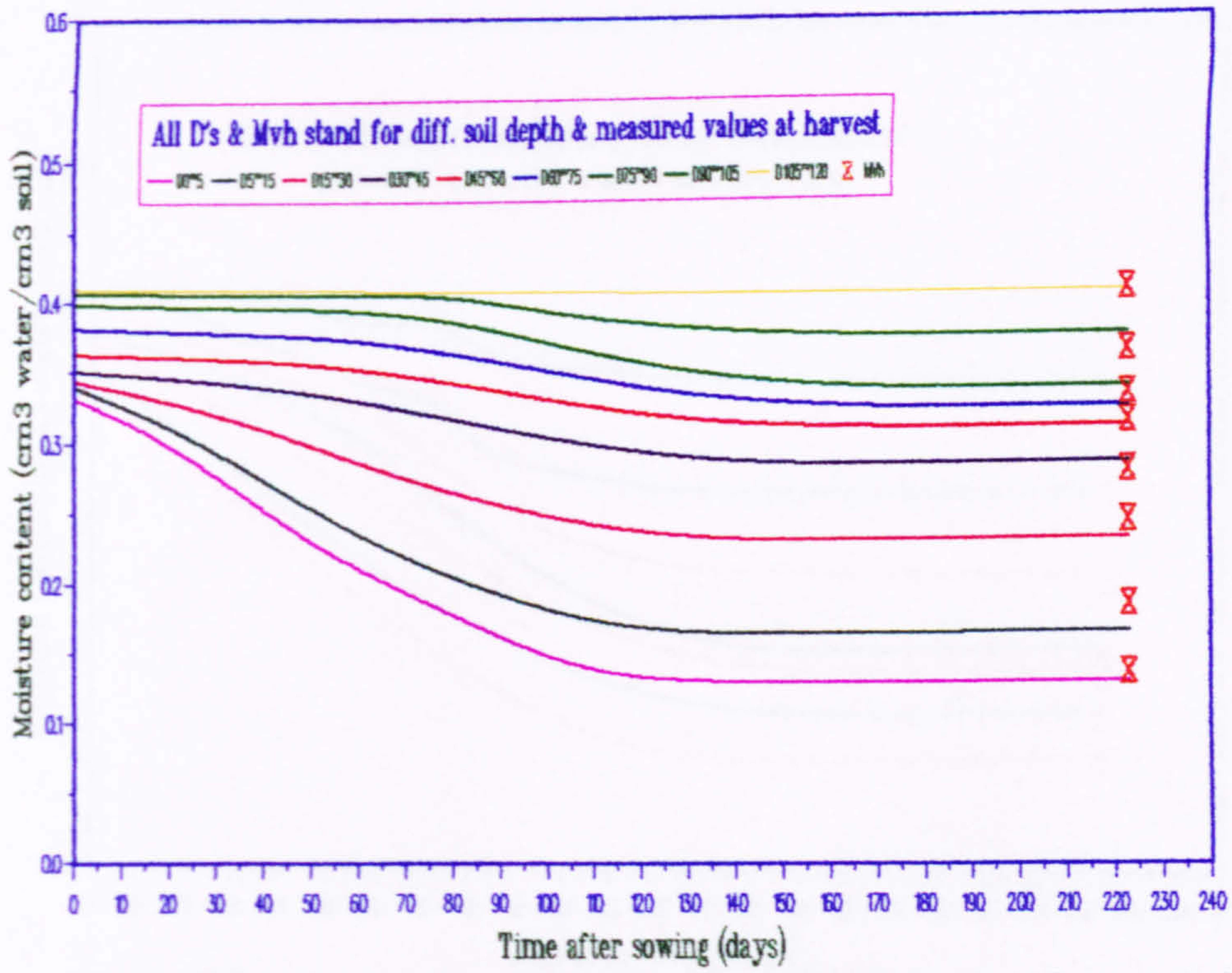


Fig. 5.1.2(c): Soil moisture content Vs time for WT-120 lysimeter (ryegrass'92)

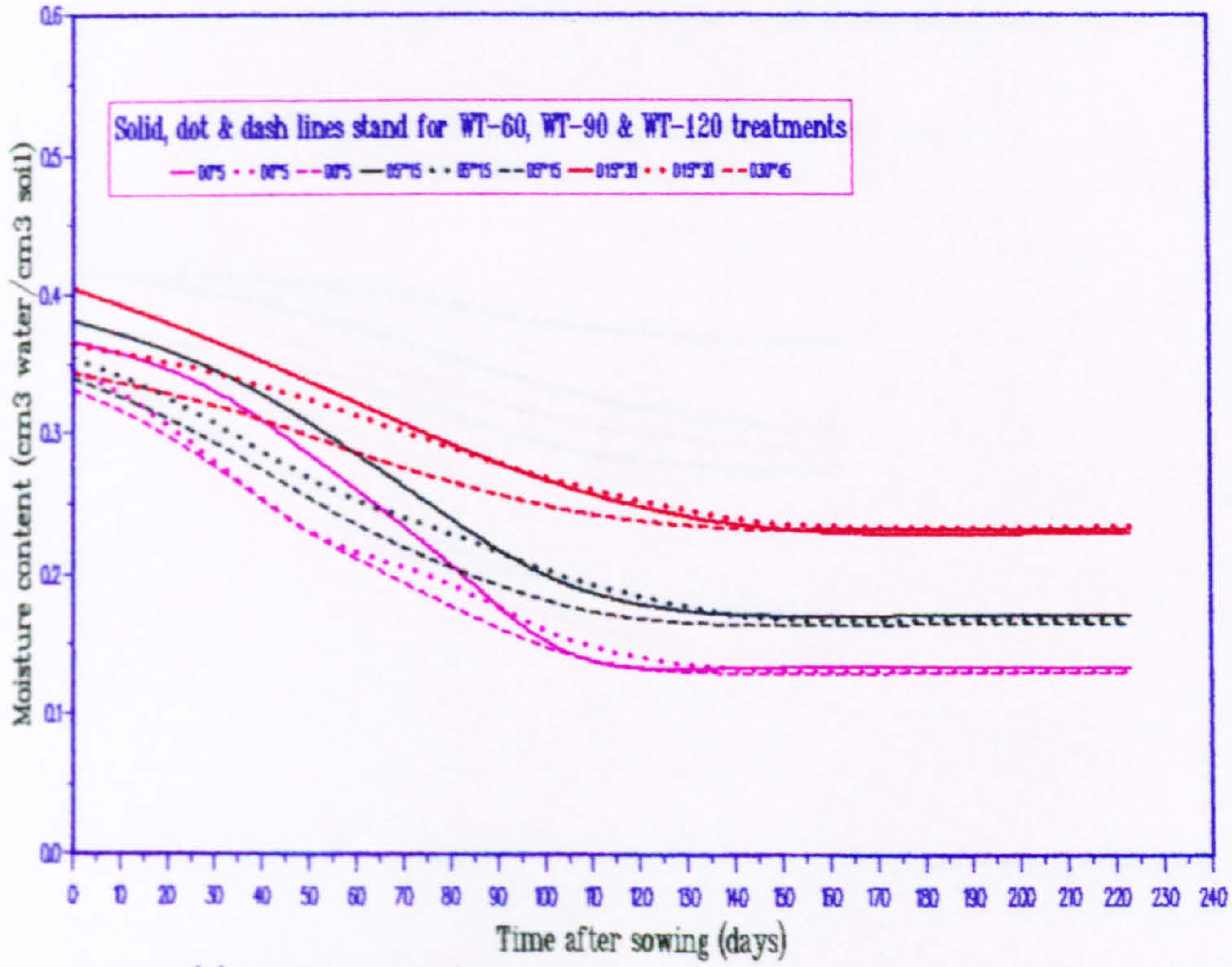


Fig. 5.1.2(d): Comparison of moisture content among water tables (ryegrass'92)

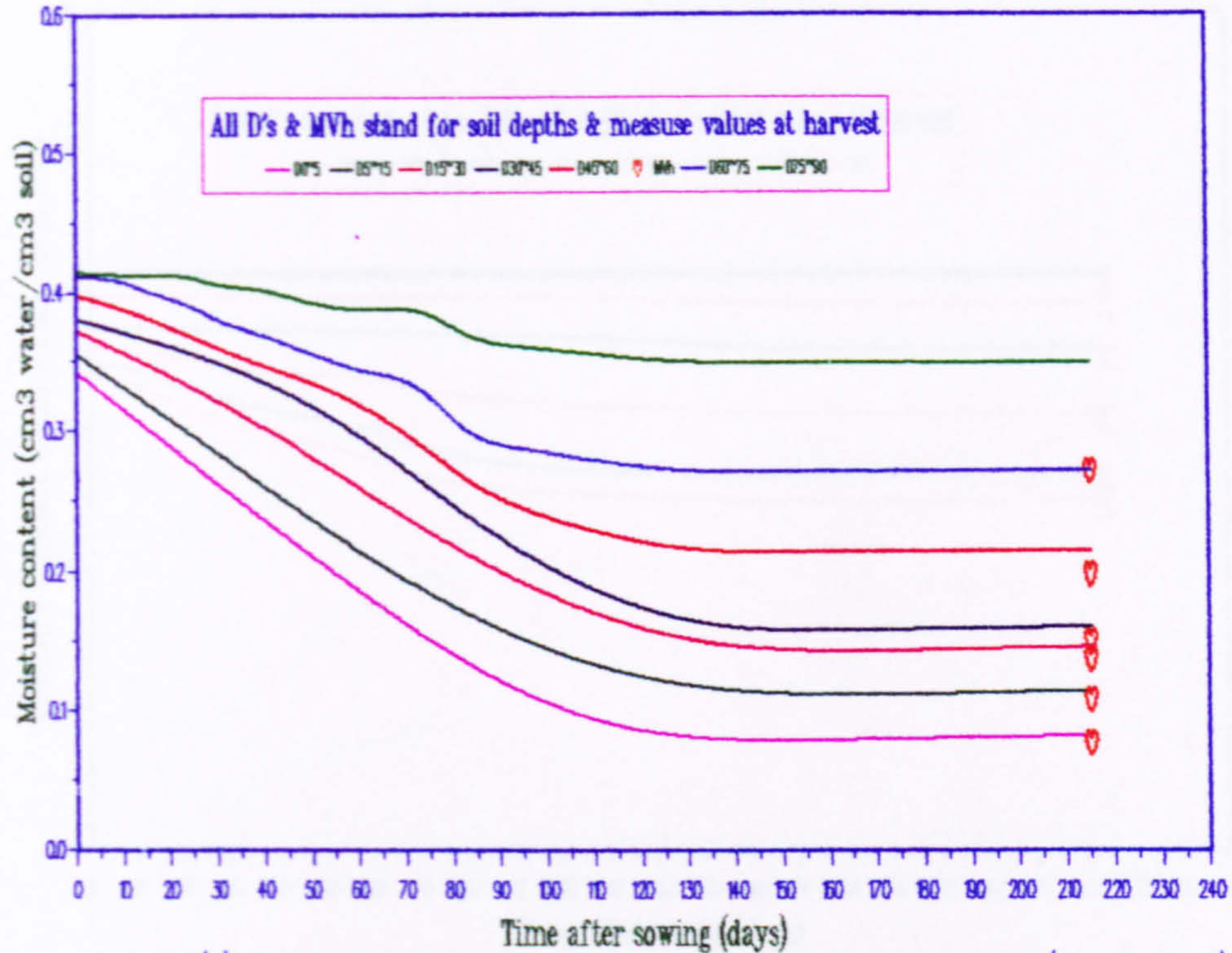


Fig. 5.1.3(a): Soil moisture content Vs time for WT-0.4 lysimeter (ryegrass'93)

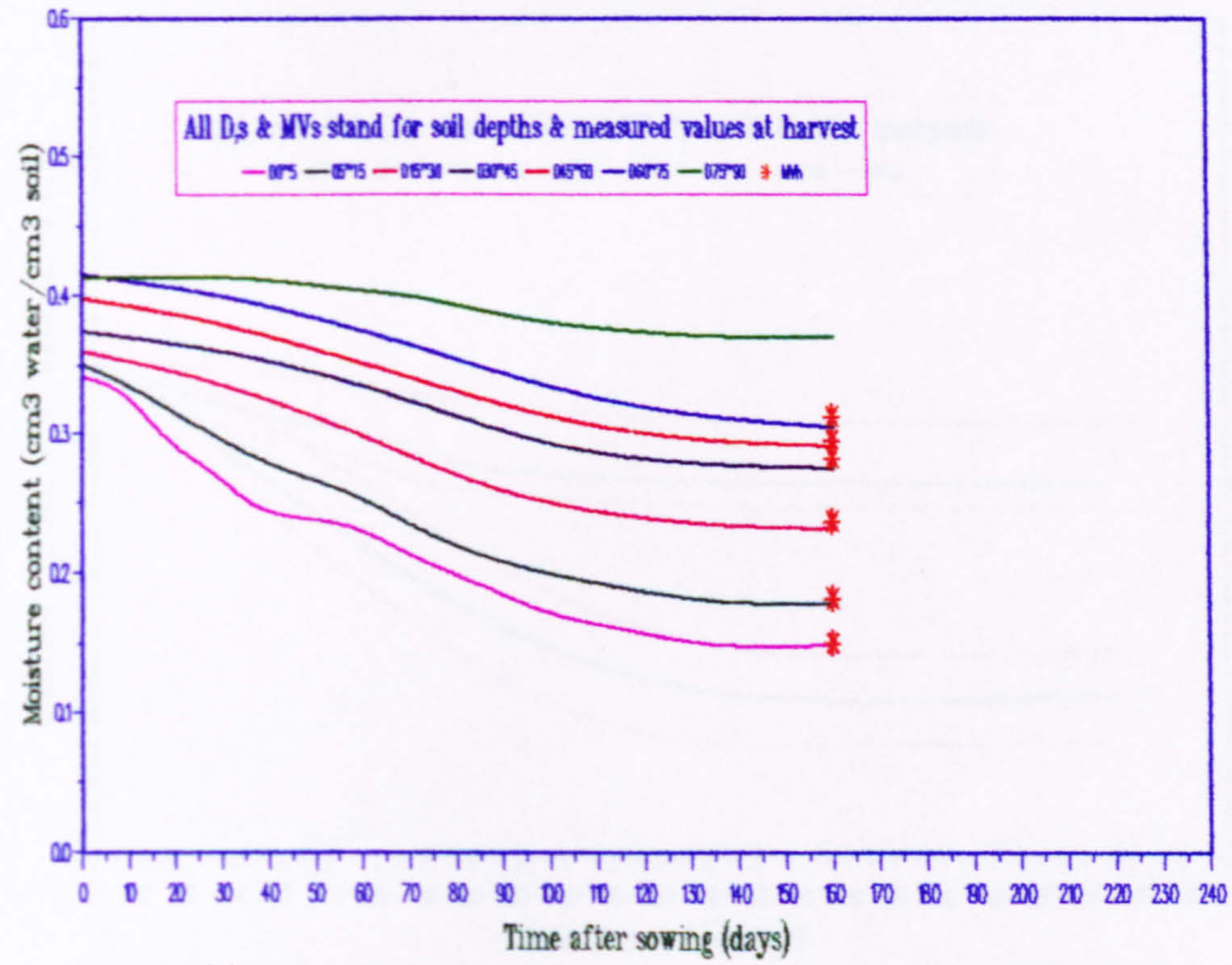


Fig. 5.1.3(b): Soil moisture content Vs time for WT-7.5 lysimeter (ryegrass'93)

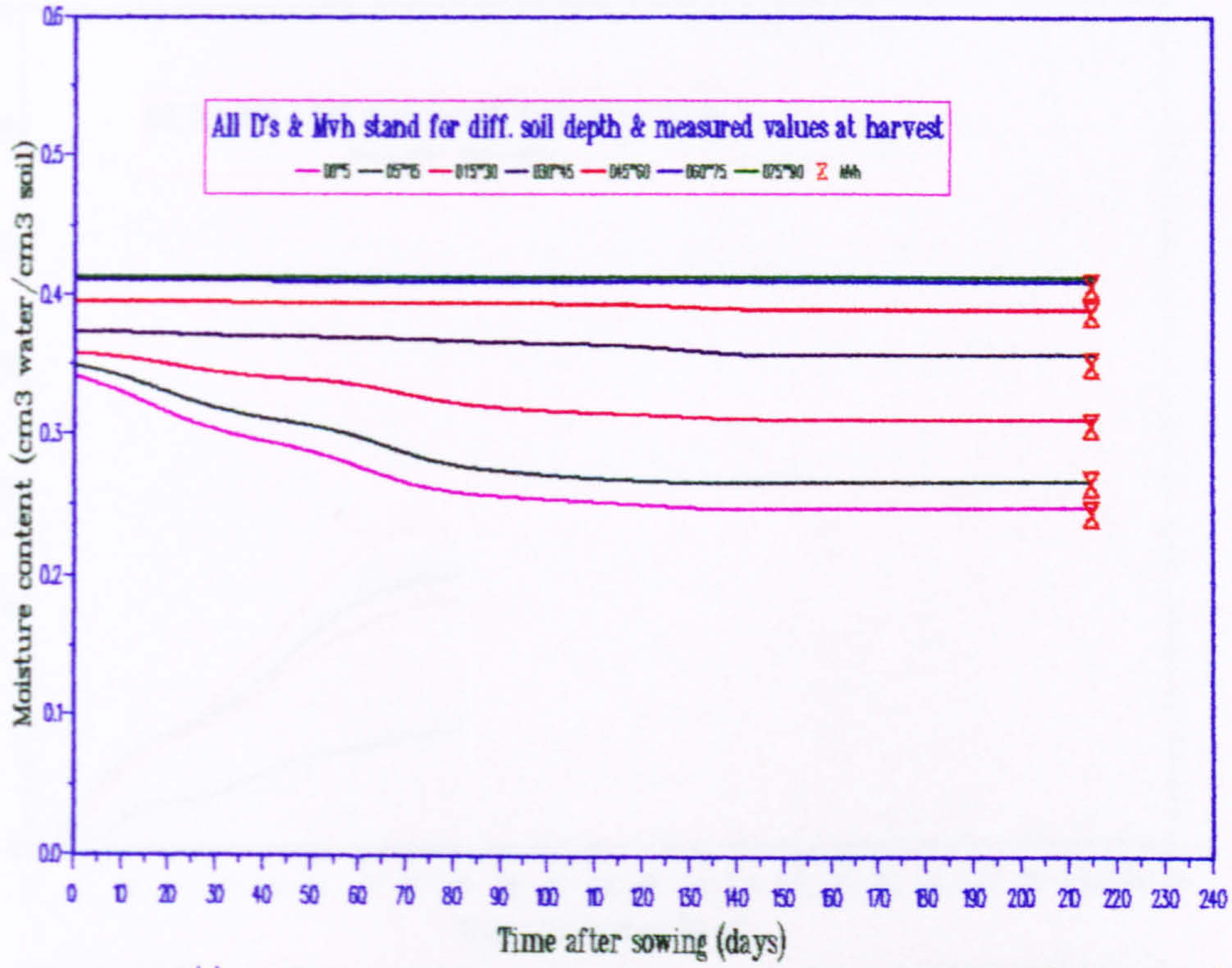


Fig. 5.1.3(c): Soil moisture content Vs time for WT-15.0 lysimeter (ryegrass'92)

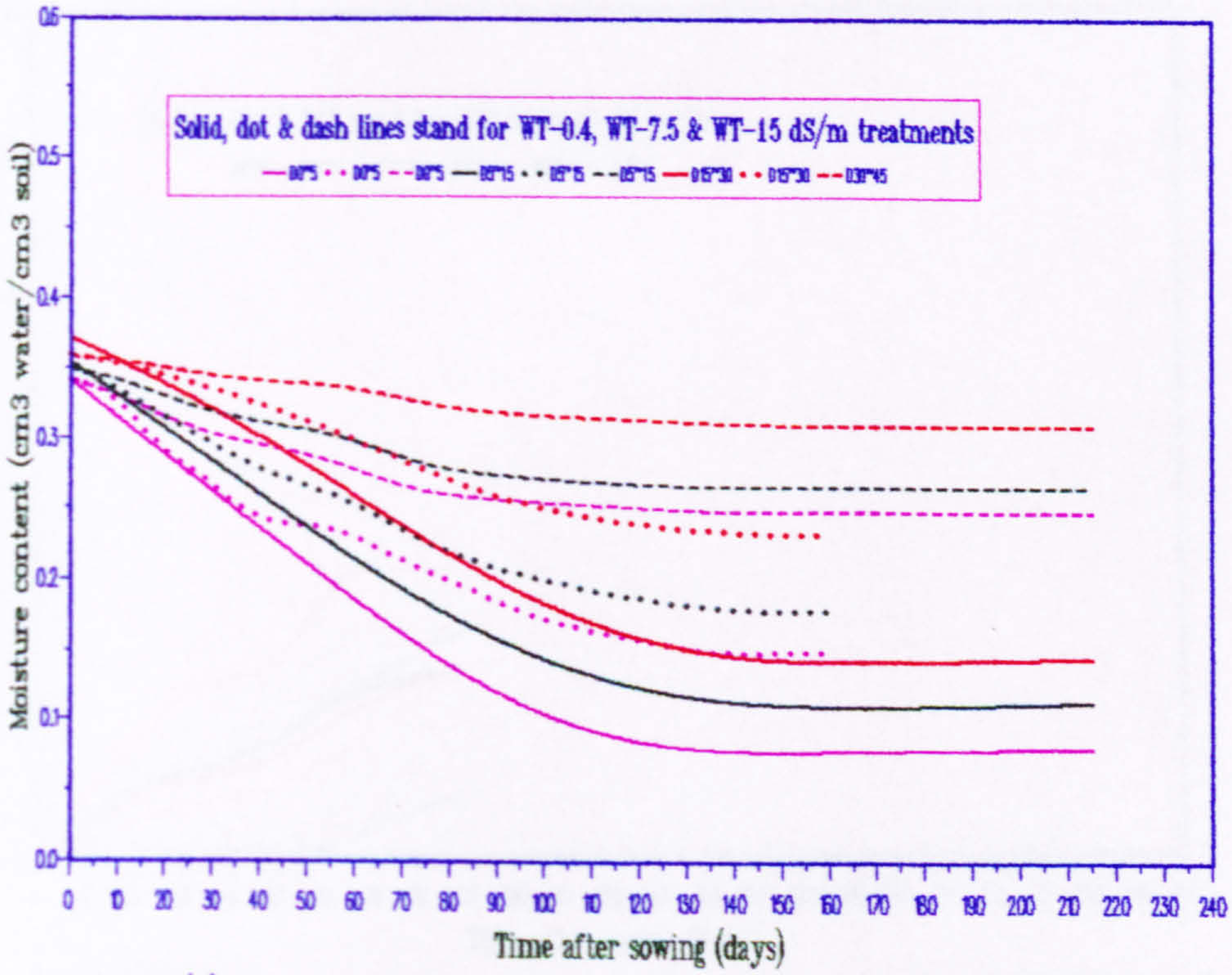


Fig. 5.1.3(d): Comparison of moisture content among salinity treatments (rye'93)

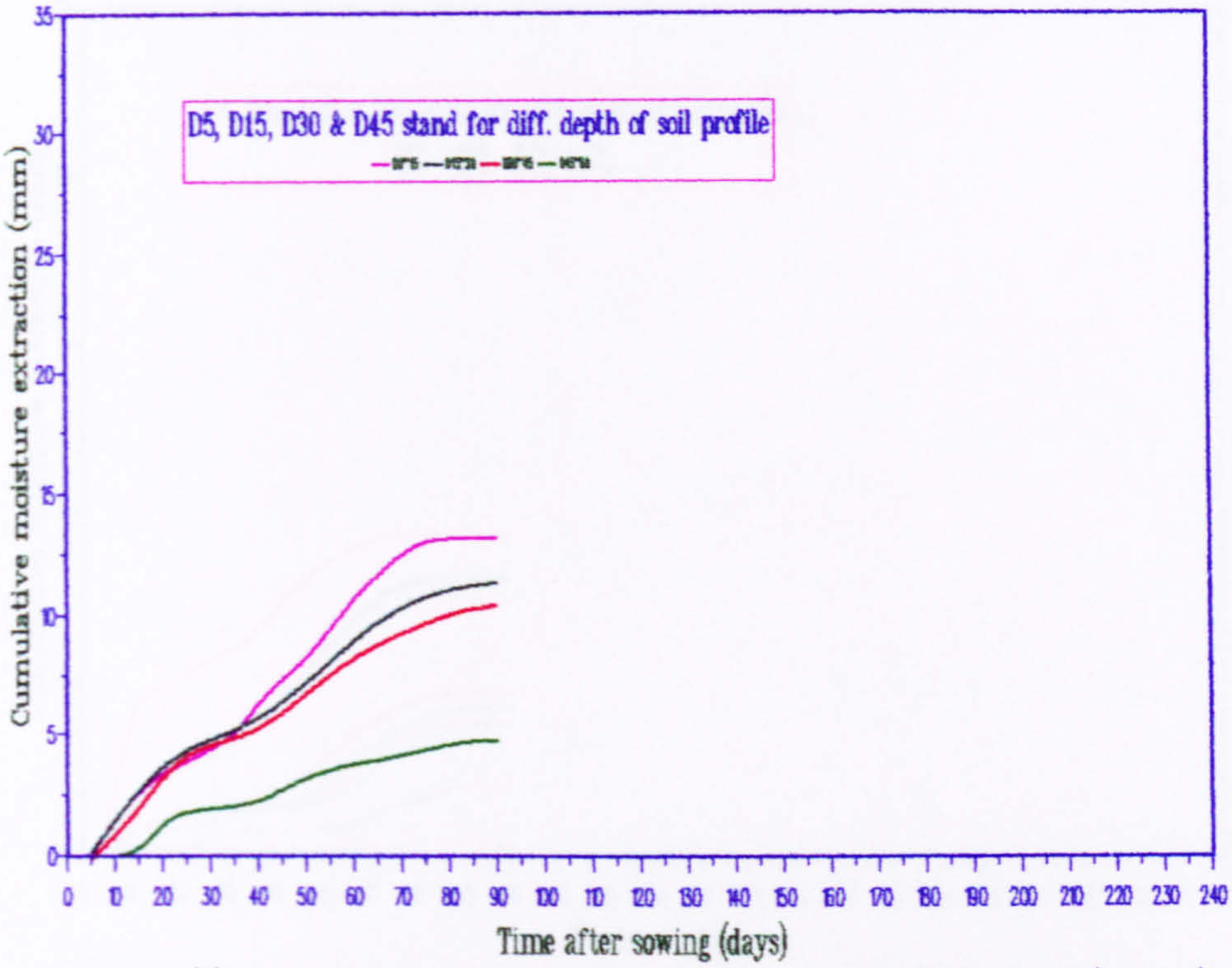


Fig. 5.1.4(a): Cumulative moisture extraction versus time for WT-60 (lettuce)

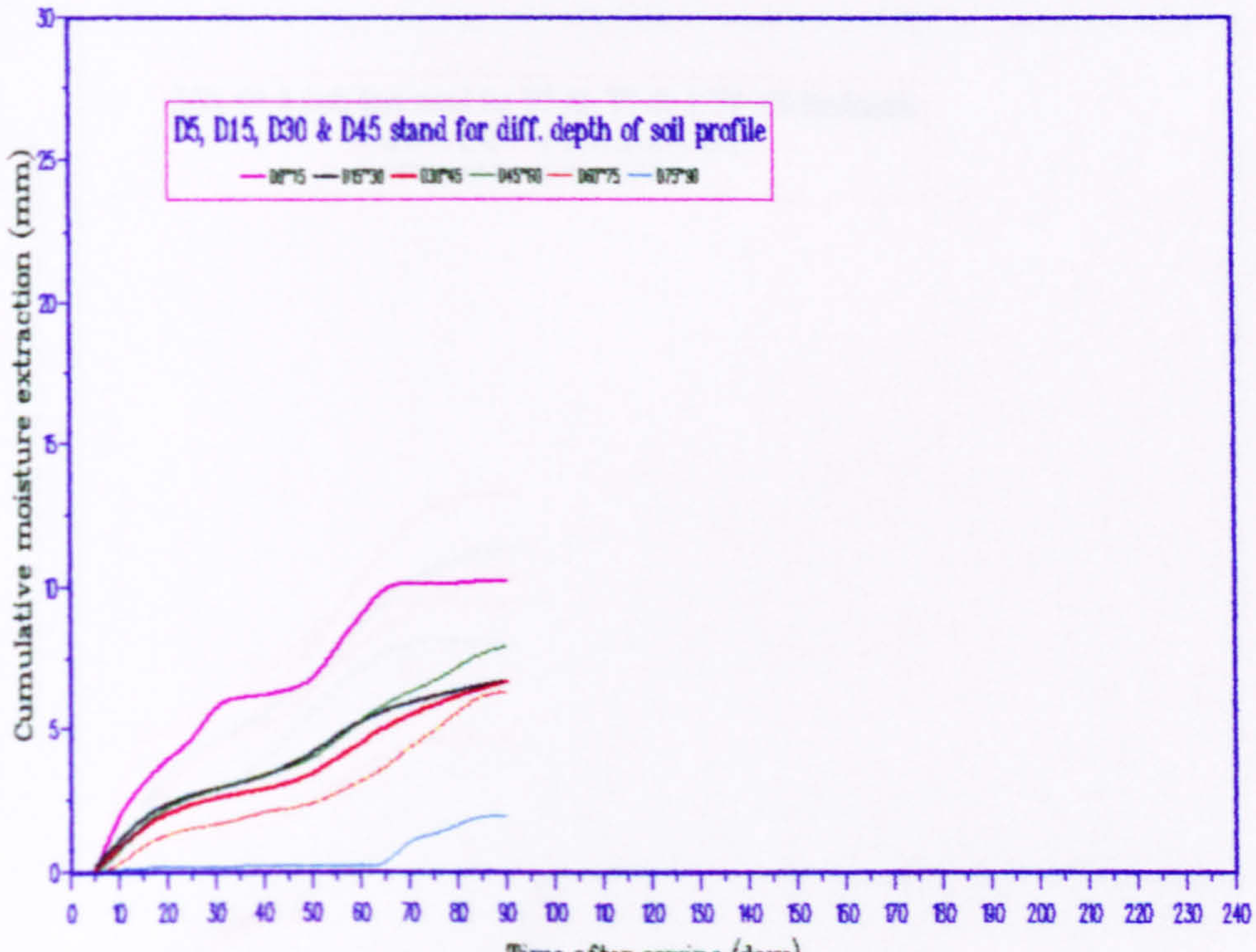


Fig.5.1.4(b): Cumulative moisture extraction versus time for WT-90 (lettuce)

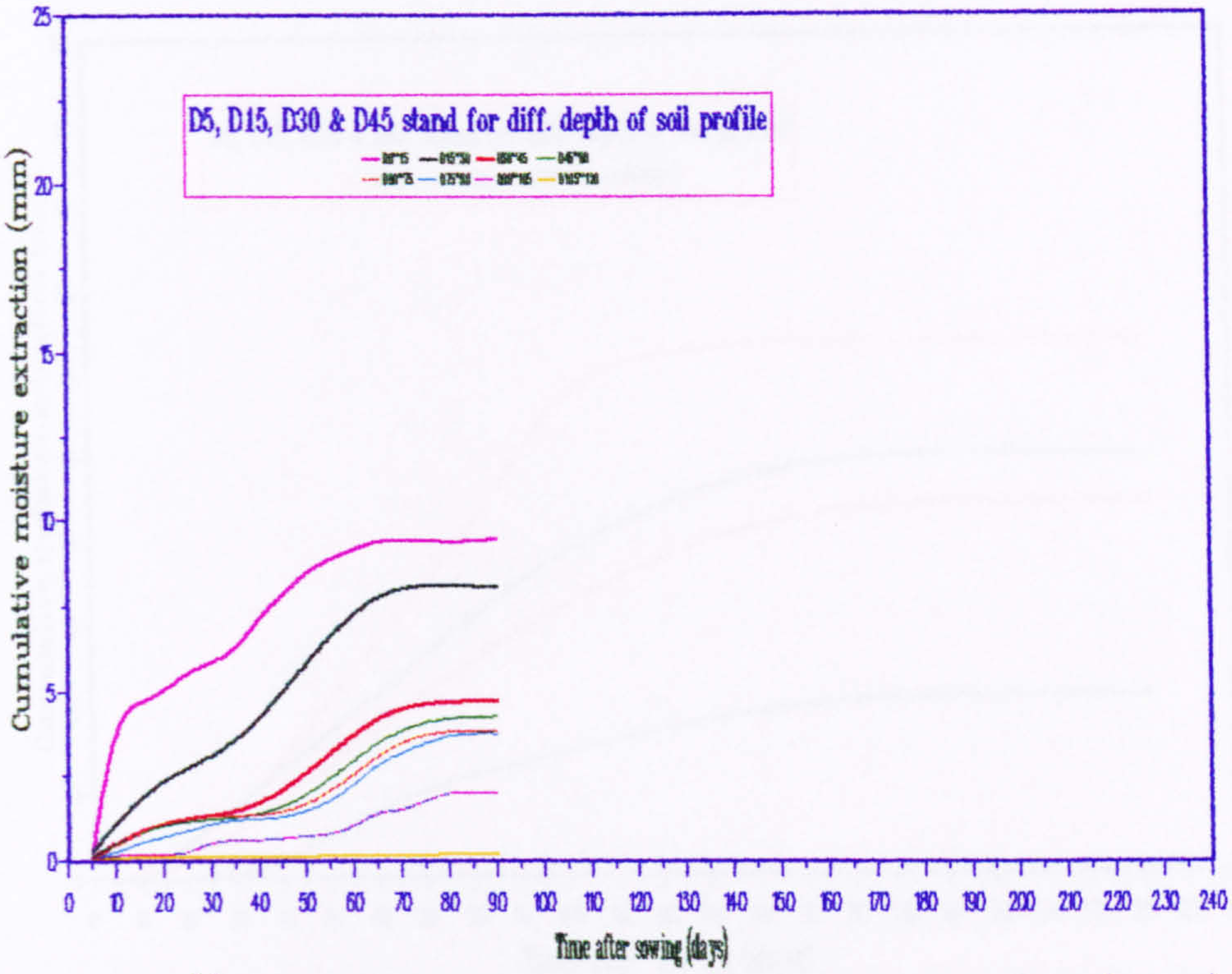


Fig. 5.1.4(c): Cumulative moisture extraction versus time for WT-120 (lettuce)

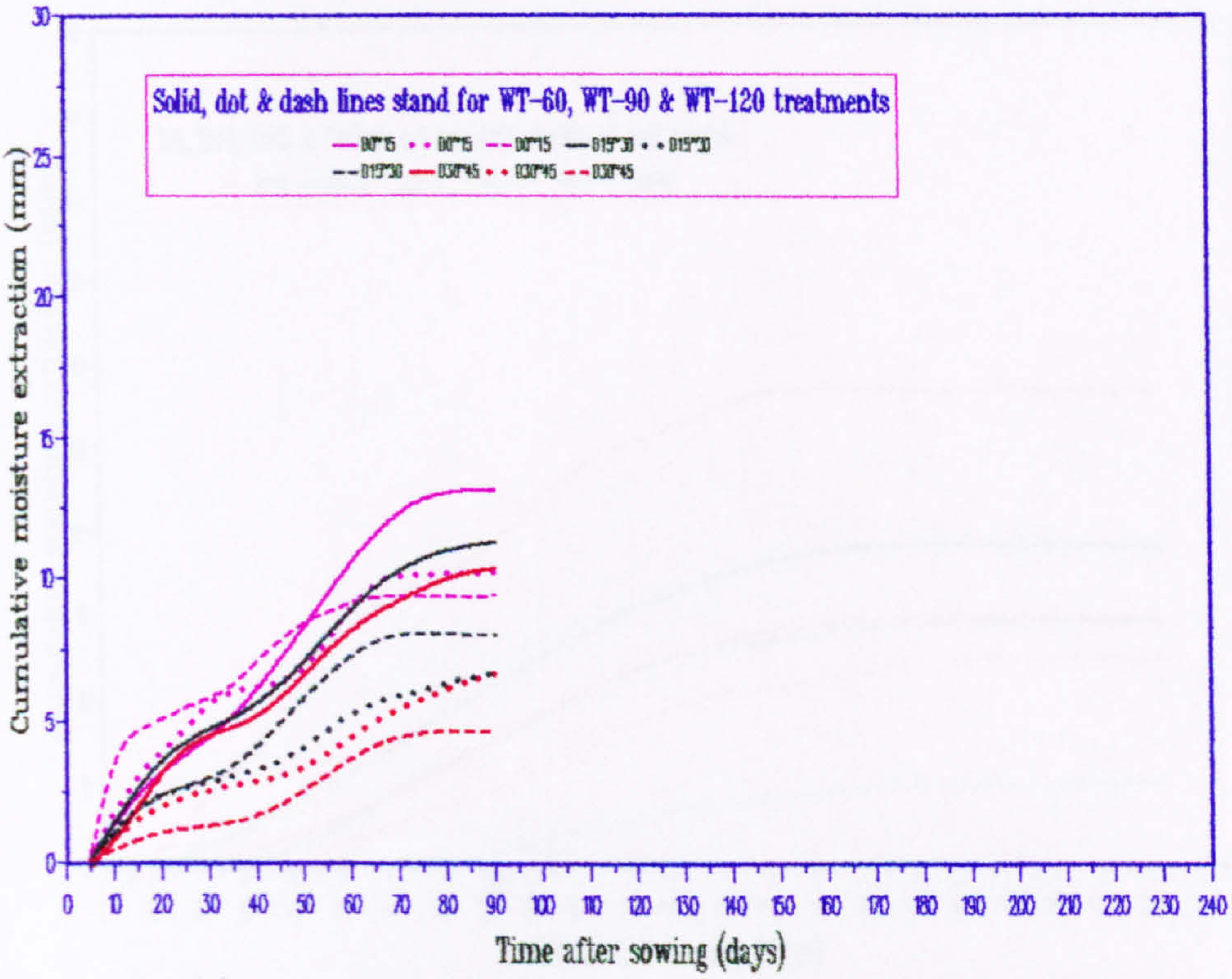


Fig. 5.1.4(d): Comparison of moisture extraction among water tables (lettuce)

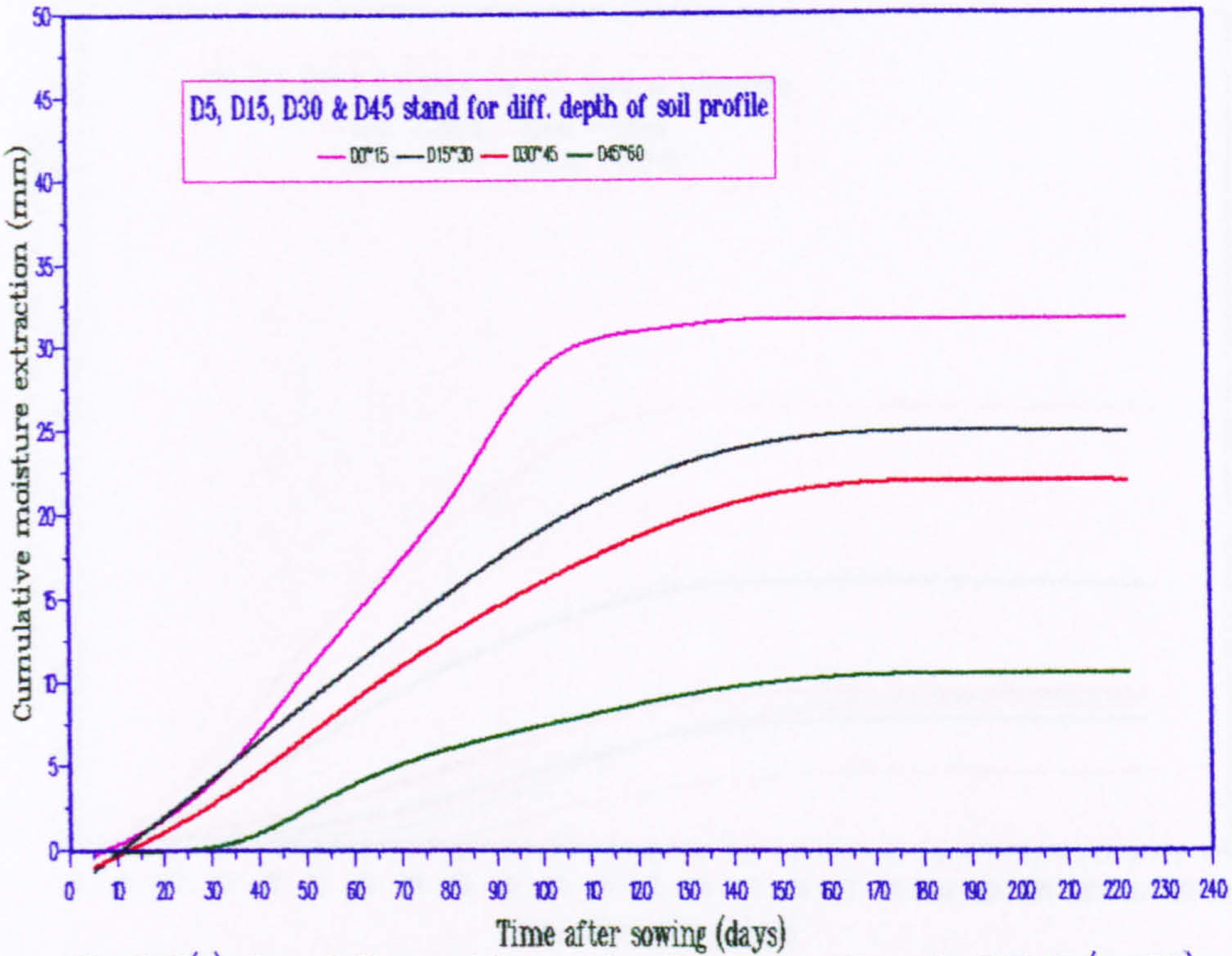


Fig. 5.1.5(a): Cumulative moisture extraction versus time for WT-60 (rye'92)

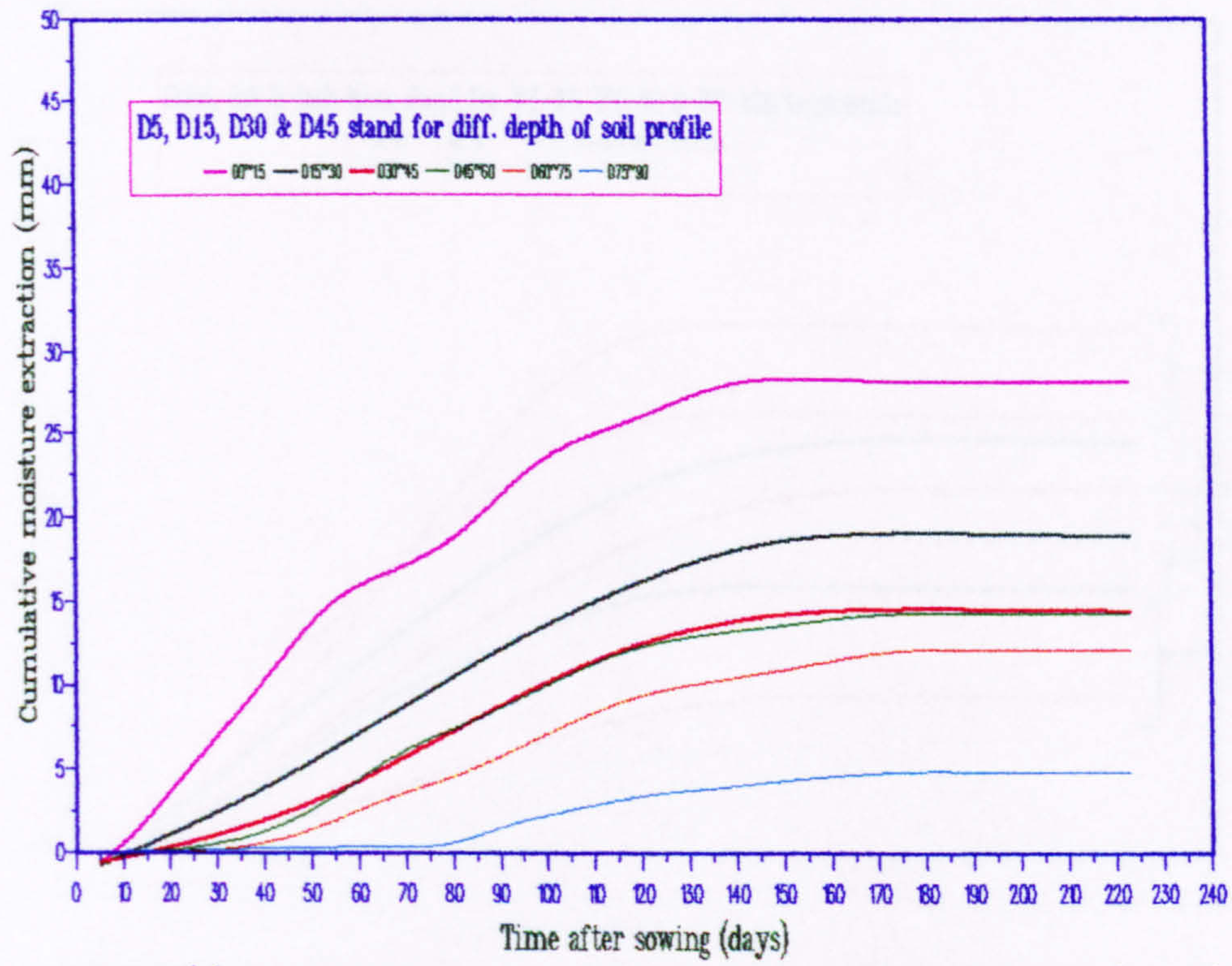


Fig.5.1.5(b): Cumulative moisture extraction versus time for WT-90 (rye'92)

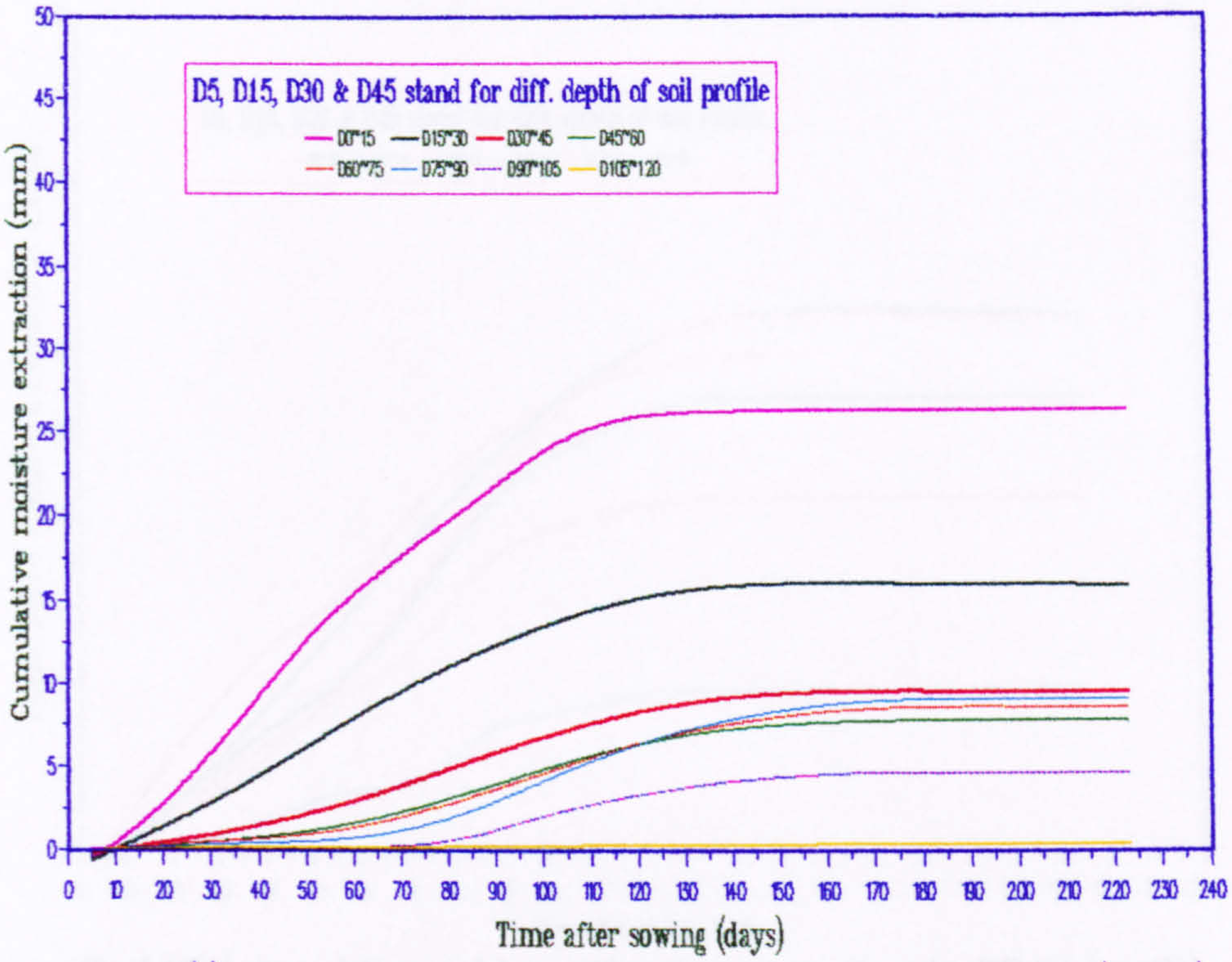


Fig. 5.1.5(c): Cumulative moisture extraction versus time for WT-120 (rye'92)

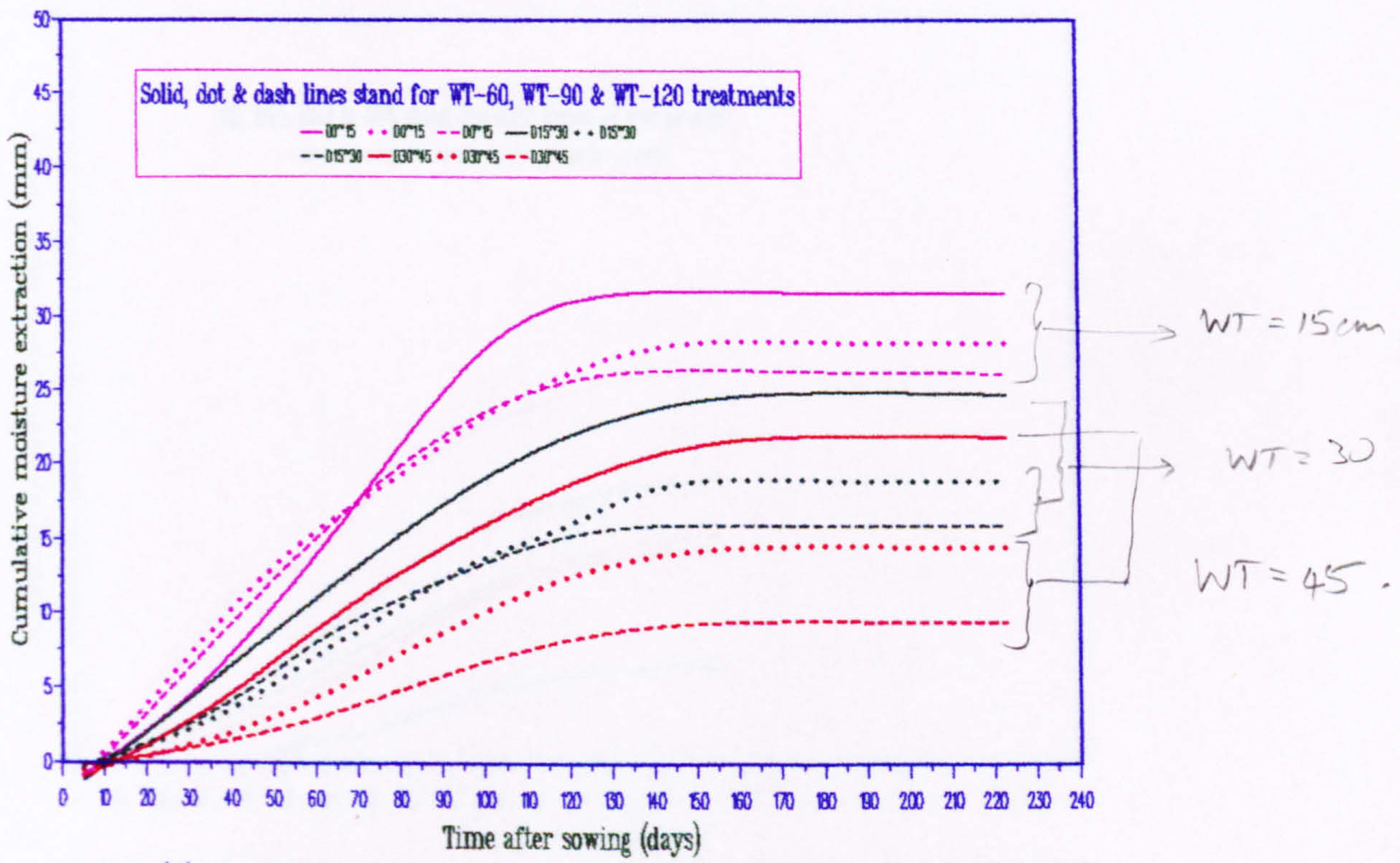


Fig.5.1.5(d): Comparison of moisture extraction among water tables (rye'92)

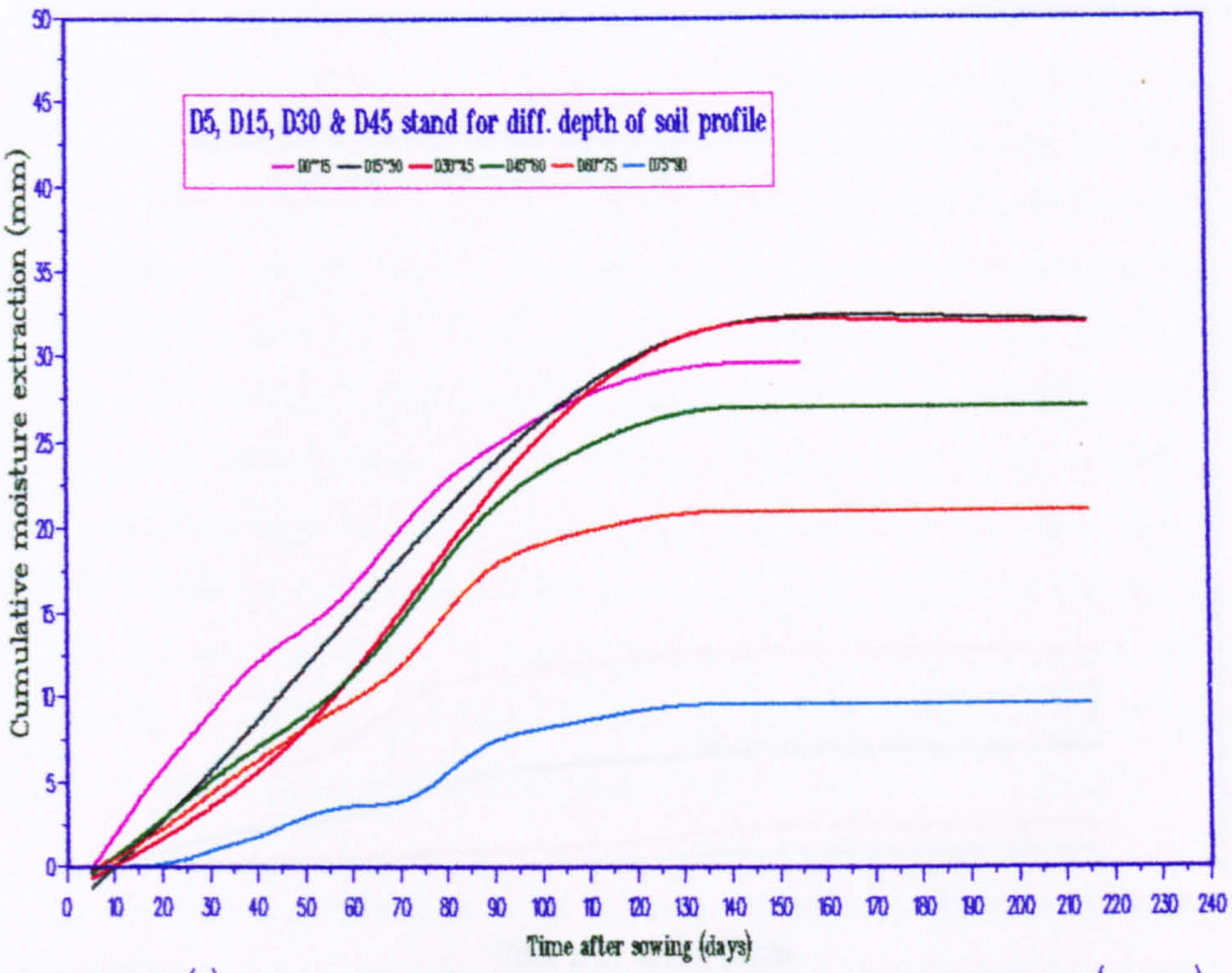


Fig. 5.1.6(a): Cumulative moisture extraction versus time for WT-0.4 (rye'93)

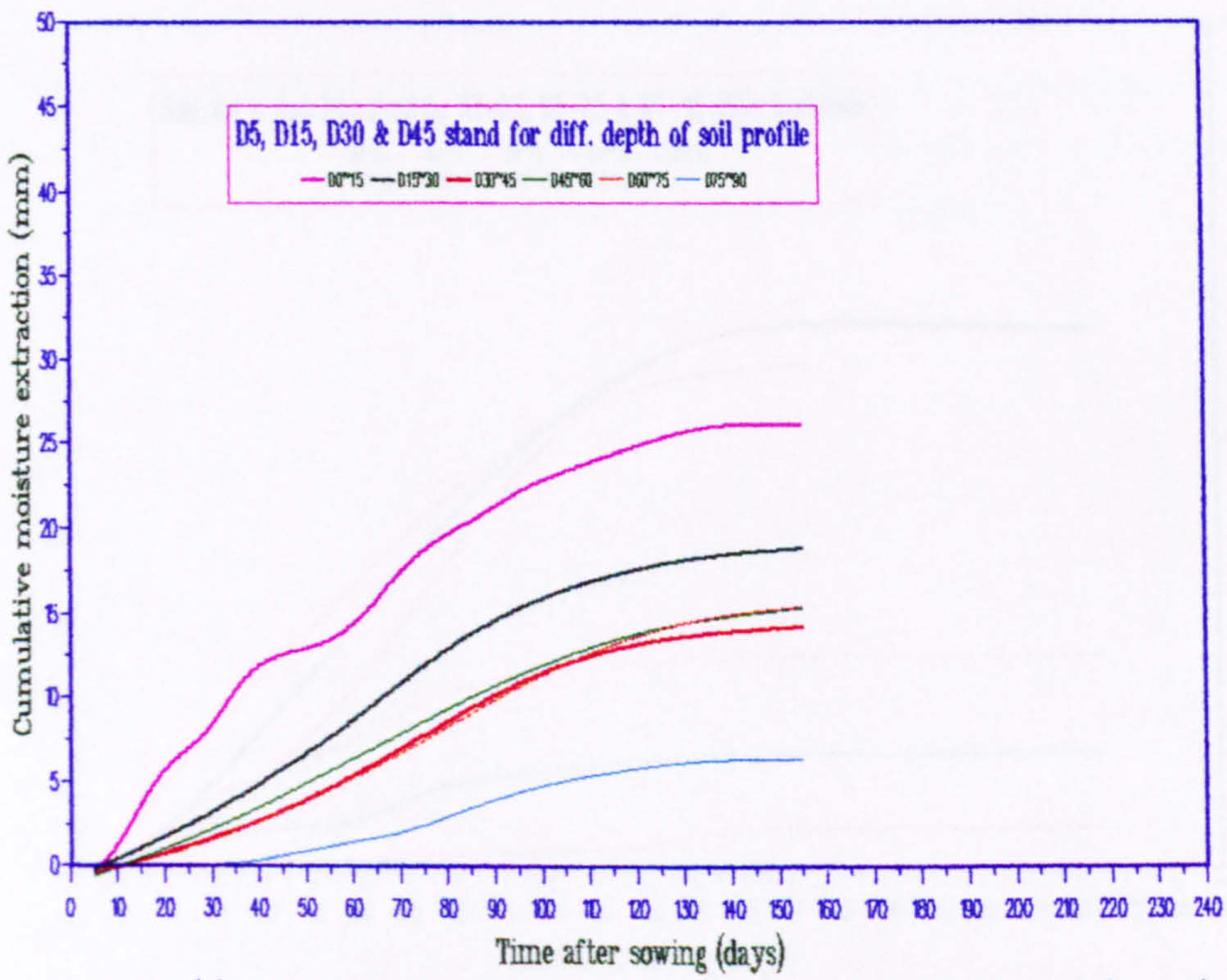


Fig. 5.1.6(b): Cumulative moisture extraction versus time for WT-7.5 (rye'93)

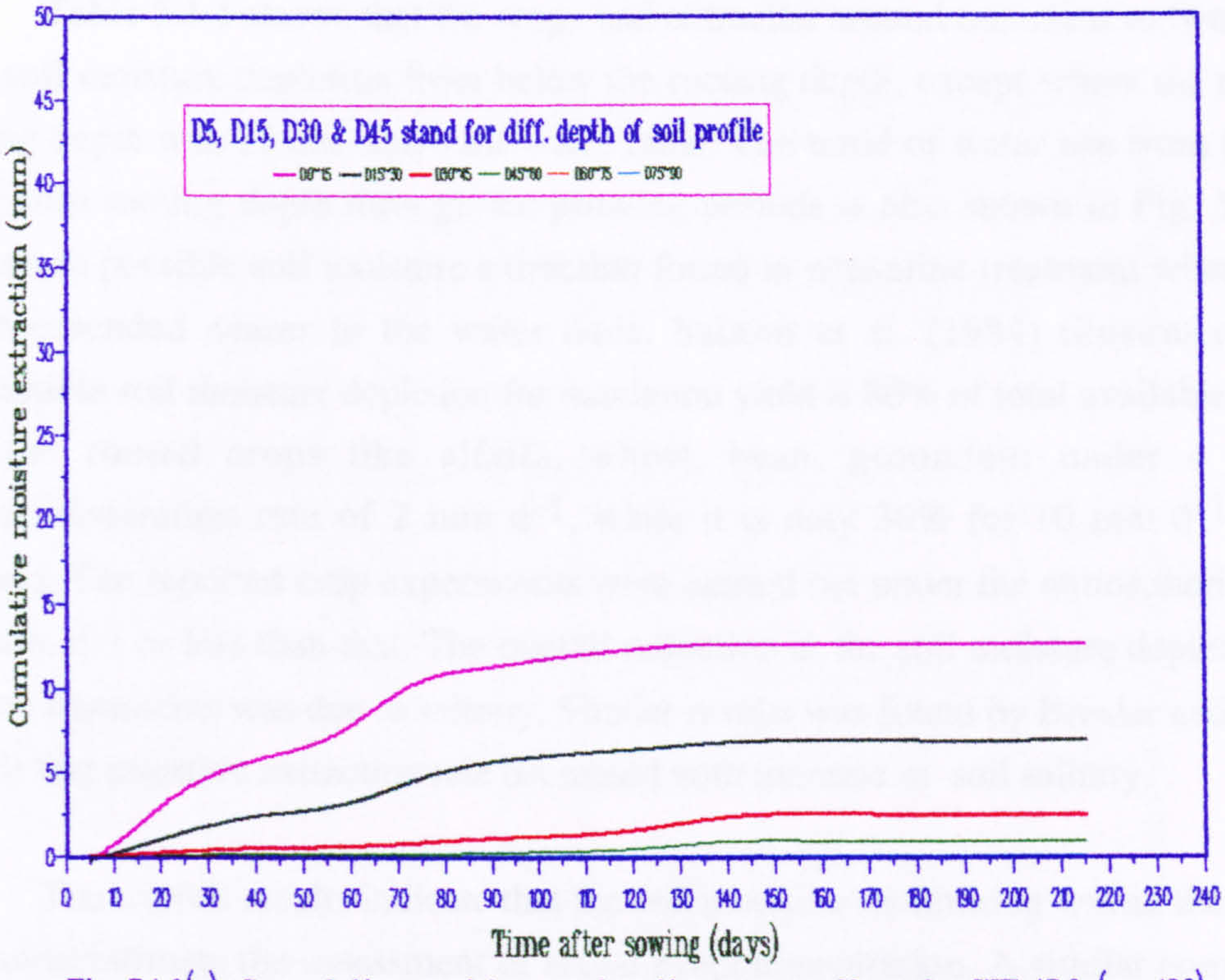


Fig. 5.1.6(c): Cumulative moisture extraction versus time for WT-15.0 (rye'93)

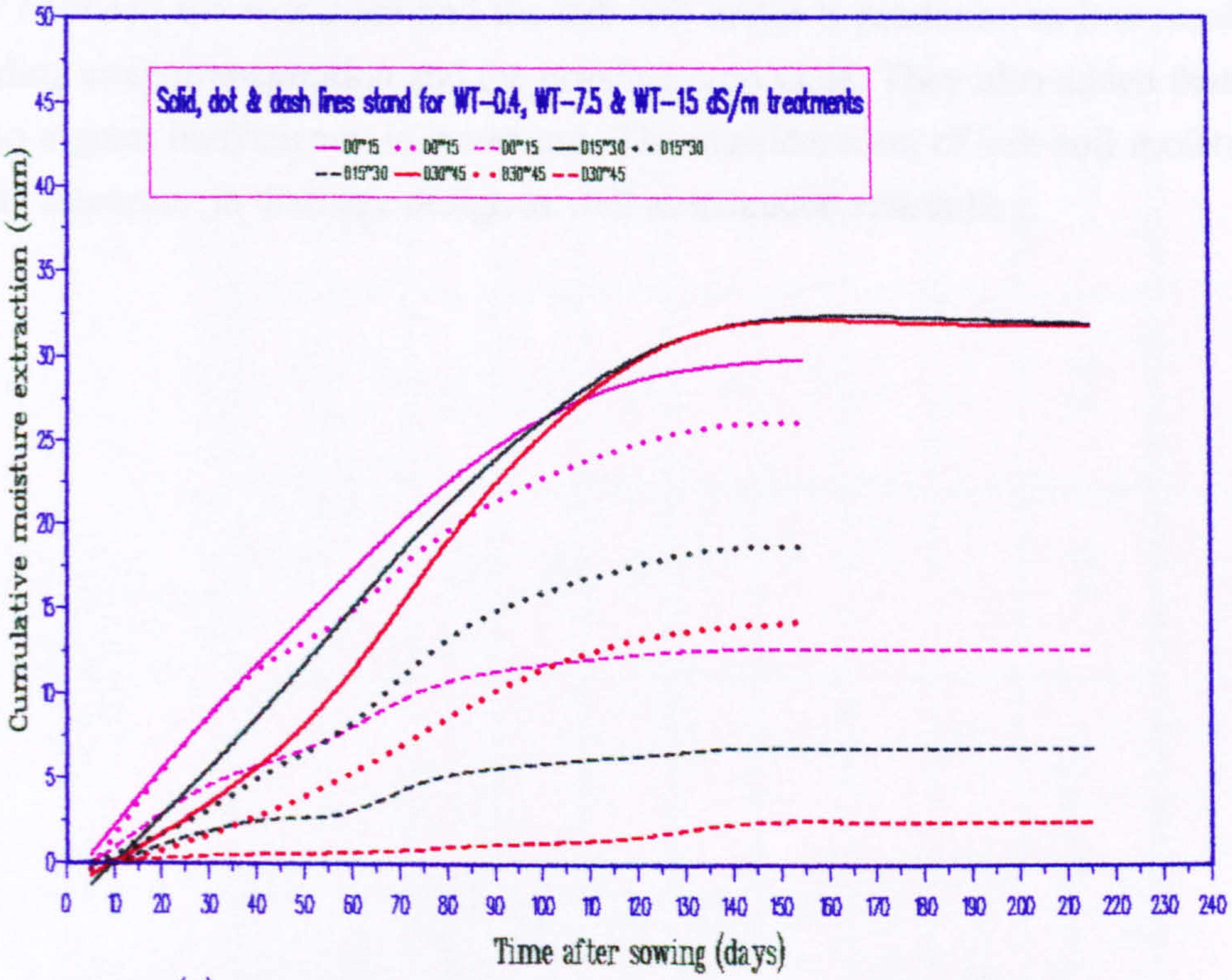


Fig5.1.6(d): Comparison of moisture extraction among salinity treatments

Table 5.1.1 shows that the crops had extracted around one-third to two-thirds of total soil moisture depletion from below the rooting depth, except where the maximum rooting depth was 15 cm above the water table. The trend of water use from below the maximum rooting depth through the growing periods is also shown in Fig. 5.1.7. The maximum possible soil moisture extraction found in nonsaline treatment where rooting depth extended nearer to the water table. Salazar et al. (1984) illustrated that the permissible soil moisture depletion for maximum yield is 80% of total available water for medium rooted crops like alfalfa, wheat, bean, groundnut under a potential evapotranspiration rate of 2 mm d^{-1}, while it is only 30% for 10 mm d^{-1} climatic demand. The reported crop experiments were carried out under the atmospheric demand of 2 mm d^{-1} or less than that. The overall reduction in the soil moisture depletion in the salinity treatments was due to salinity. Similar results was found by Bresler and Hoffman (1986) that moisture extraction rate decreased with increase in soil salinity.

The overall results indicate that the soil moisture monitoring within the root zone will underestimate the assessment of actual evapotranspiration. A similar comment was made by Ghali and Svehlik (1988) who reported, based on numerical simulations, that scheduling irrigation on the basis of analytical models which ignored the transfer of soil water between the root zone and the sub-soil under it produces an inaccurate picture regarding evapotranspiration and the possible crop yield. They also added that this may lead to a great inefficiency in water use. The consideration of sub-soil moisture use is equally important in drainage design as well as irrigation scheduling.

Depth of soil profile (cm)	Types of crops / Cropping year								
	Lettuce 1991	Lettuce 1991	Lettuce 1991	Ryegrass 1992	Ryegrass 1992	Ryegrass 1992	Ryegrass 1993	Ryegrass 1993	Ryegrass 1993
	Water table depth (cm) / Water table salinity (dS m^{-1})								
	60 / 4.5	90 / 4.5	120 / 4.5	60 / 9.4	90 / 9.4	120 / 9.4	90 / 0.4	90 / 7.5	90 / 15.0
0-5	12	10	8	13	11	12	8	10	20
5-15	21	16	18	23	19	20	15	18	35
15-30	29	17	22	28	20	19	21	20	30
30-45	26	17	13	24	17	13	20	15	11
45-60	12	20	12	12	15	9	17	15	4
60-75	—	16	11	—	13	10	13	16	0
75-90	—	4	10	—	5	11	6	6	0
90-105	—	—	6	—	—	6	—	—	—
105-120	—	—	0	—	—	0	—	—	—
Rooting depth (cm)	7.0	9.0	11.0	60.0	45.0	45.0	75.0	45.0	30.0

Table 5.1.1: Percentage of soil moisture extraction from different depths of soil profile for different treatments.

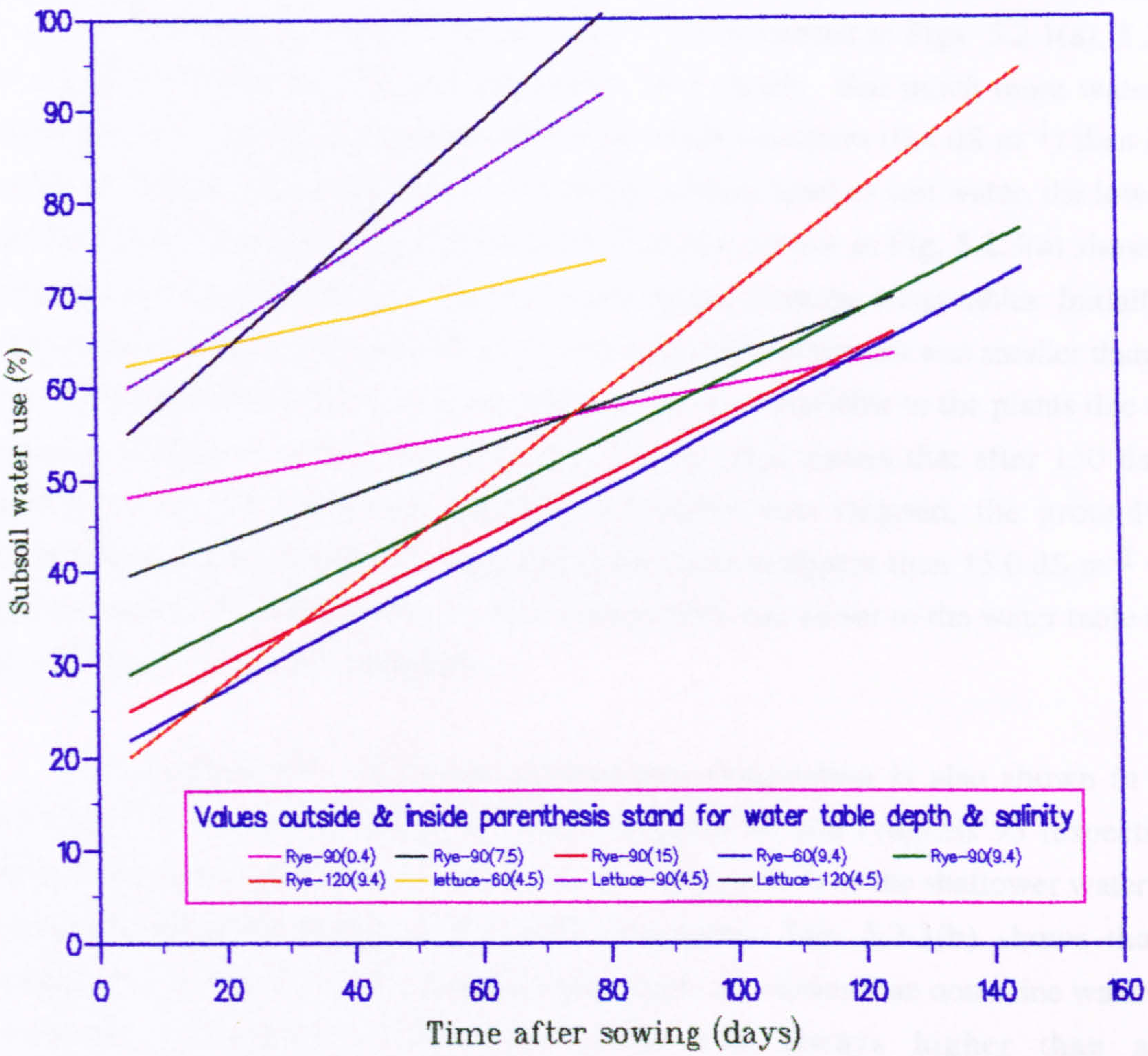


Fig. 5.1.7: Percentage of sub-soil water use for different treatments.

5.2 Groundwater contribution and total water use

Total water use was the sum of soil moisture use, the groundwater contribution and a little irrigation during fertilizer applications. Cumulative soil water depletion and groundwater contribution as a function of time for different salinity levels and water table depths with lettuce, ryegrass'92 & ryegrass'93 are presented in Figs. 5.2.1(a), 5.2.2(a) & 5.2.3(a) respectively. Fig. 5.2.3(a) shows very clearly that much more water was extracted from the soil in the non-saline water table treatment (0.4 dS m^{-1}) than in the saline water table treatments and the higher the salinity level of soil water, the lower the soil moisture extraction. The groundwater extraction pattern in Fig. 5.2.3(a) shows that there was not a big difference in the total water uptake from the water tables. Initially, the groundwater contribution from the non-saline water table treatment was smaller than from the saline water tables because more soil moisture was available to the plants due to the absence of salts or with a little salts. Fig. 5.2.3(a) also shows that after 150 days of sowing to harvest when soil moisture extraction was stopped, the groundwater contribution was higher in the non-saline water table treatment than 15.0 dS m^{-1} water table treatment. The reason may be the rooting depth was closer to the water table in 0.6 dS m^{-1} than 15.0 dS m^{-1} treatment.

The cumulative total water use and pan evaporation is also shown in Figs. 5.2.1(b), 5.2.2(b) and 5.2.3(b) for lettuce, ryegrass'92 and ryegrass'93 respectively. More water was used by the crops over the growing periods in the shallower water table treatments than the deeper water table treatments. Fig. 5.2.3(b) shows that the consumptive use in the saline water table treatments was lower than nonsaline water table treatment. Evaporation from open water was always higher than actual evapotranspiration, because of matric and osmotic stresses to the plants. The ratio of actual evapotranspiration to free water evaporation gradually decreased as the total stress in the root zone increased.

Detailed water use by lettuce and for different cuts of ryegrass crops are presented in Tables 5.2.1 and 5.2.2 and 5.2.3 respectively. The ratio C/E is the proportion of total water use contributed by groundwater.

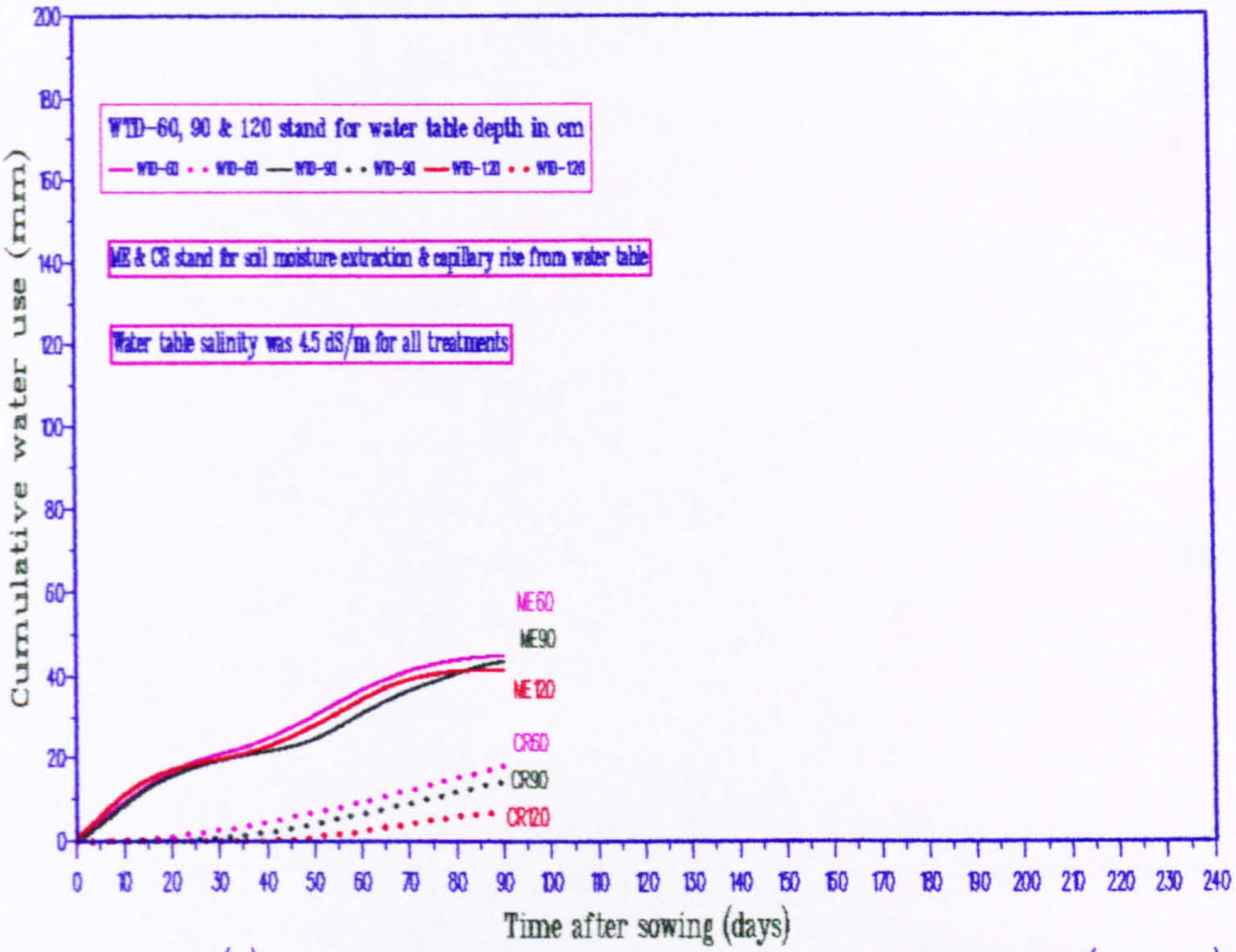


Fig. 5.2.1(a): Cu. water use from soil profile & water table (lettuce)

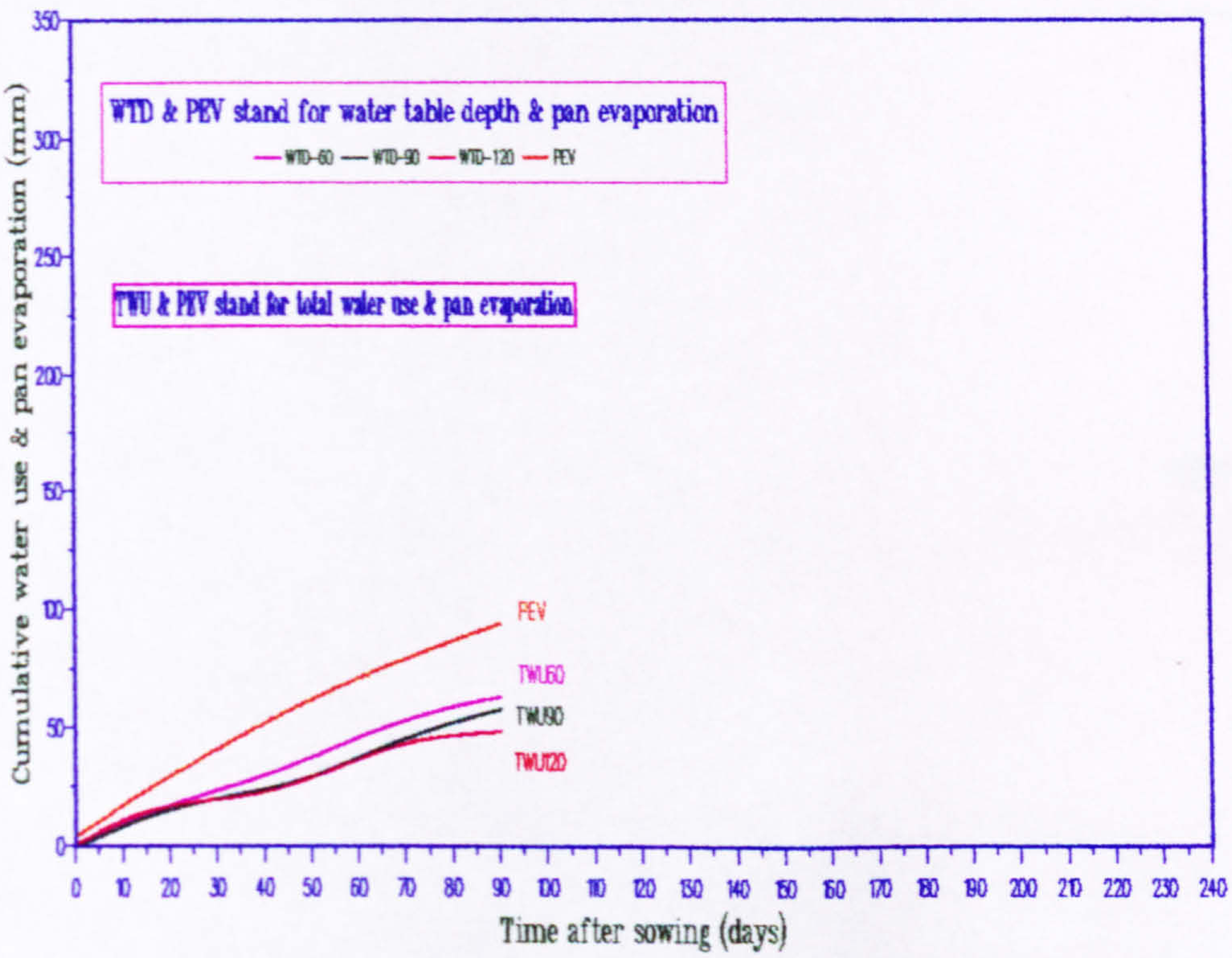


Fig. 5.2.1(b): Cu. total water use & pan evaporation (lettuce)

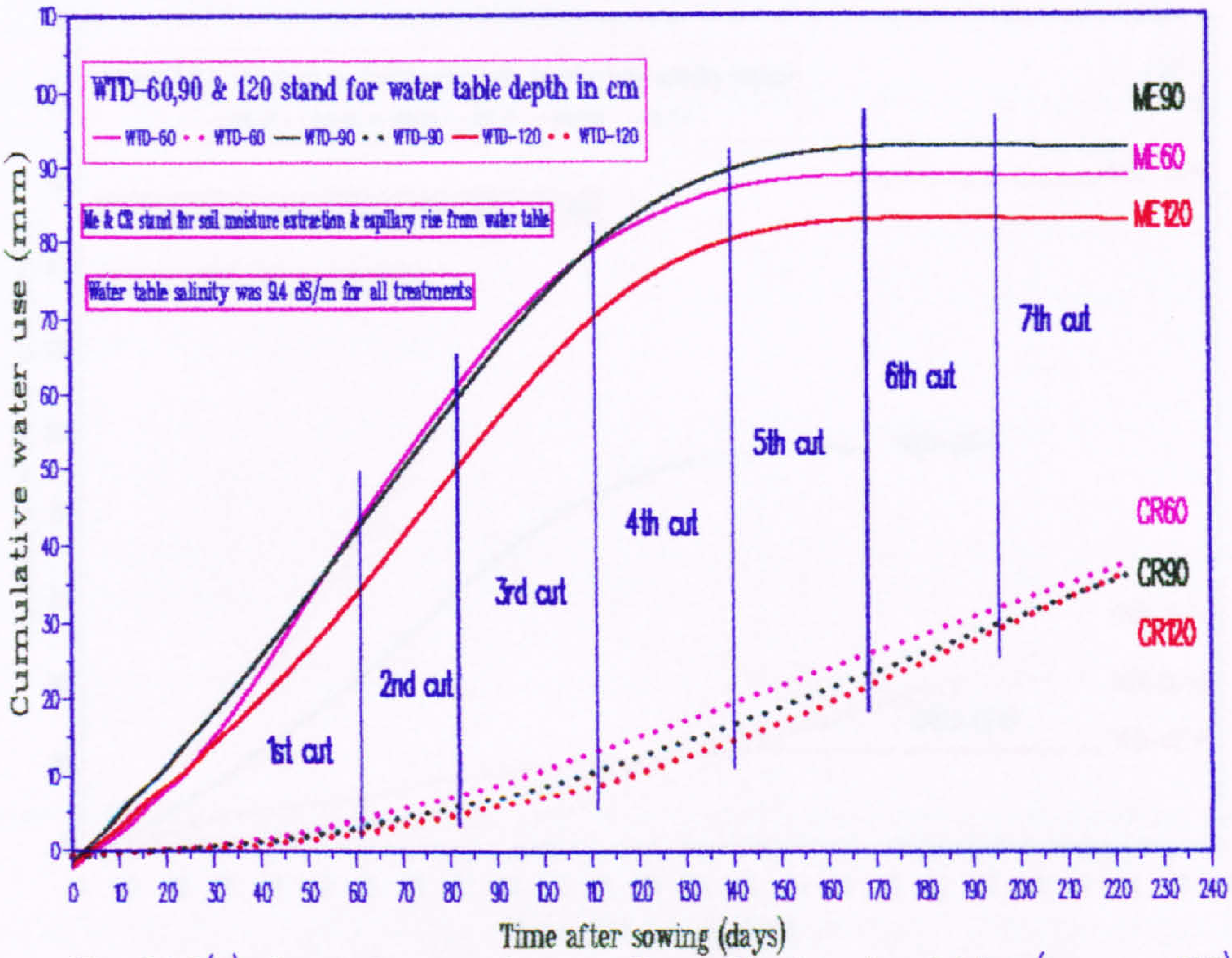


Fig. 5.2.2(a): Cu. water use from soil profile & water tables (ryegrass'92)

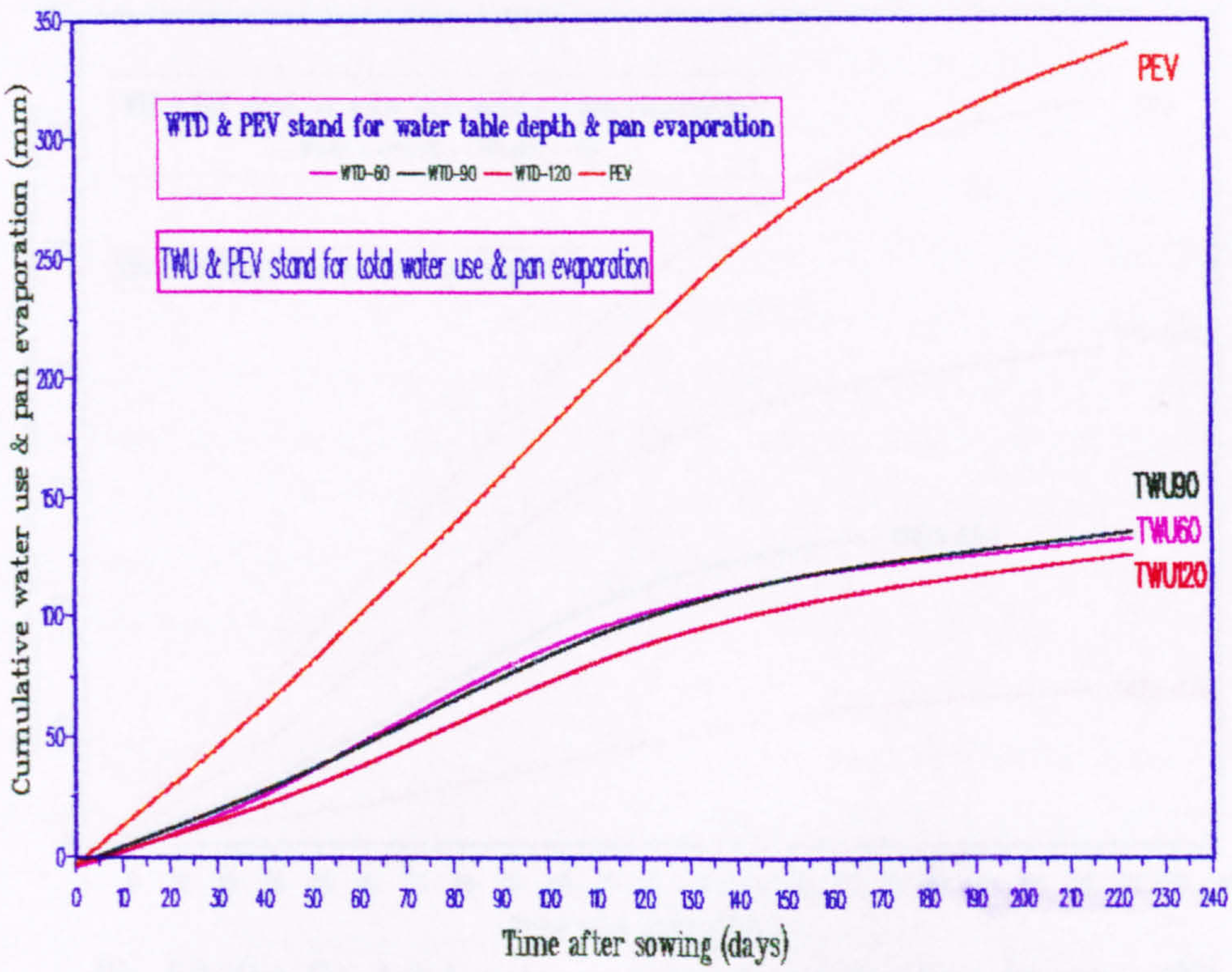


Fig. 5.2.2(b): Cu. total water use & pan evaporation (ryegrass'92)

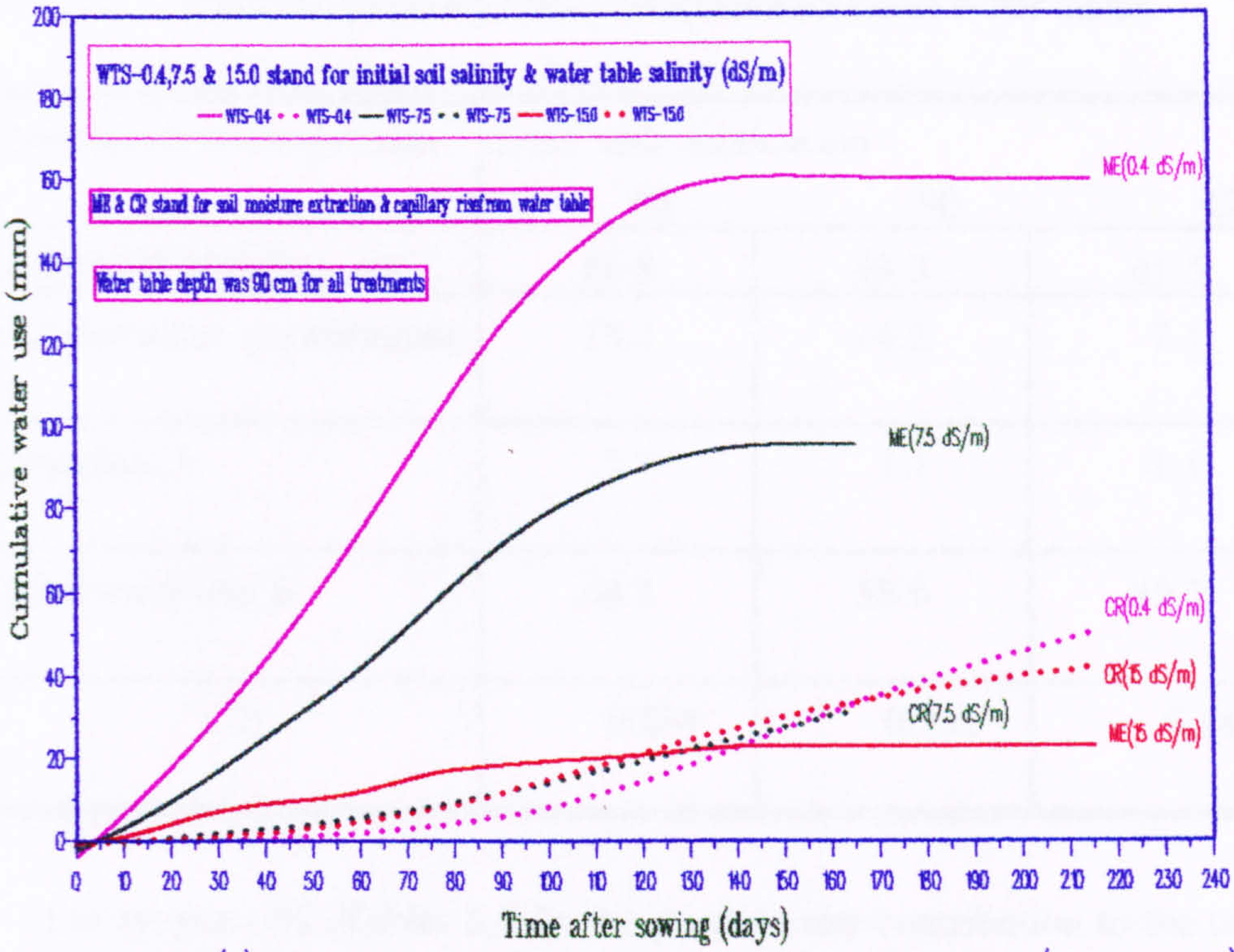


Fig. 5.2.3(a): Cu. water use from soil profile & water table (ryegrass'93)

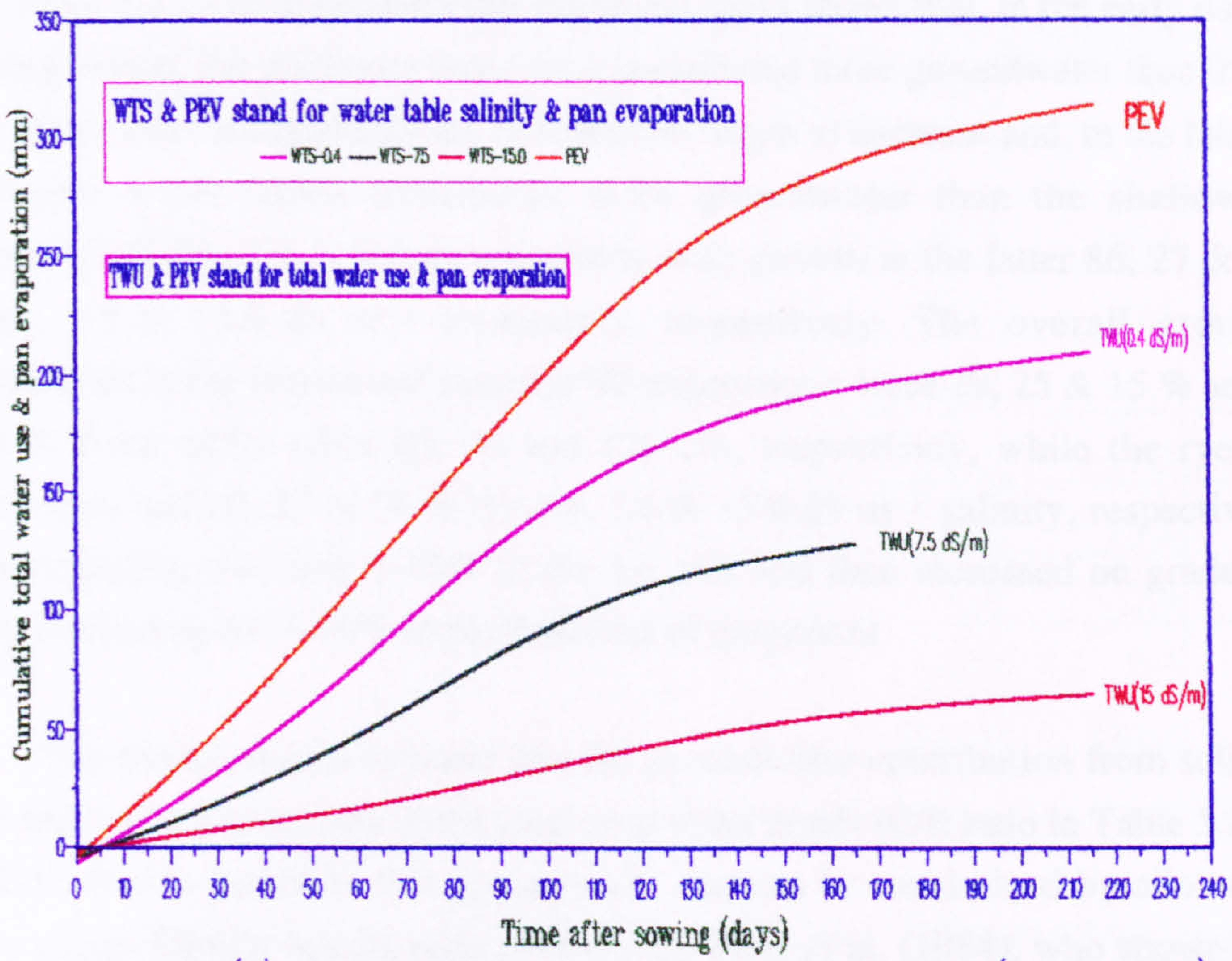


Fig. 5.2.3(b): Cu. total water use & pan evaporation (ryegrass'93)

Table 5.2.1: Water use (mm) by lettuce from different saline water tables.

Description of components	water table depth in cm 60	90	120
Soil water depletion, Δw	44.8	43.3	41.5
Groundwater contribution, C	18.2	14.2	7.1
Irrigation, I	1.1	1.1	1.1
Total water use, E	64.1	58.6	49.7
C/E	0.284	0.242	0.143

For ryegrass'92 (Tables 5.2.2), the groundwater contribution to the total water use gradually increased and the growth in the latter 75, 46 and 49 days out of cropping periods of 216, 223 and 213 days, respectively, was accomplished with groundwater only. Table 5.2.2 (same groundwater salinities) again shows that, in the early days of the cropping period, the shallower water table contributed more groundwater than the deeper water tables and then gradually the contribution began to decrease and, in the later stages, the deeper water tables contributed more groundwater than the shallower. The ryegrass'93 (Table 5.2.3) behaved similarly with growth in the latter 86, 27 & 84 days for 0.4, 7.5 & 15.0 dS m^{-1} treatments, respectively. The overall groundwater contributions in the lettuce and ryegrass'92 experiments were 29, 25 & 15 % and 28, 27 & 29 % from water table 60, 90 and 120 cm, respectively, while the ryegrass'93 experiments had 22, 23 & 59 % for 0.4, 7.5 & 15.0 dS m^{-1} salinity, respectively. But the contribution was only 2-18% in the 1st cuts and then increased on gradually and finally reached up to 76-90% in the final cuts of ryegrasses.

The overall results indicate that the groundwater contribution from saline water tables had met a greater part of the total crop water needs (C/E ratio in Table 5.2.1, 5.2.2 & 5.2.3). It also indicates that groundwater use can be maximized by allowing more matric stress. Similar results were obtained by Benz et al. (1984), who showed that the water table through sub-irrigation provided a sizeable contribution to actual evapotranspiration, which increased as the level of surface irrigation decreased. They also noted that sub-irrigation provided 38.4% of total ET from the low (0.3 ET) irrigation level, compared to only 0.6% of total ET for the high (1.3 ET) irrigation level.

Description of components		Soil water depletion, dW (mm) Water table depth (cm)			Groundwater contribution, C (mm) Water table depth (cm)		
Cut	Days	60	90	120	60	90	120
First	62	43.8	41.6	34.0	3.8	2.9	2.3
Second	83	16.3	16.9	14.7	3.7	2.6	2.5
Third	111	18.5	20.8	22.3	5.3	4.8	3.9
Fourth	139	8.8	10.5	9.4	6.3	5.7	4.7
Fifth	167	1.6	3.0	2.6	6.8	7.1	7.9
Sixth	195	0.0	0.0	0.0	5.9	6.7	7.9
Seventh	223	0.0	0.0	0.0	5.7	6.4	7.2
Total		89.0	92.8	82.9	37.4	36.2	36.4

Description of components		Irrigation, I (mm) Water table depth (cm)			Total water use, E (mm) Water table depth (cm)		
Cut	Days	60	90	120	60	90	120
First	62	1.1	1.1	1.1	48.7	45.6	37.4
Second	83	1.1	1.1	1.1	21.1	20.6	18.3
Third	111	1.1	1.1	1.1	24.9	26.7	27.3
Fourth	139	1.1	1.1	1.1	16.2	17.3	15.2
Fifth	167	1.1	1.1	1.1	9.5	11.2	11.5
Sixth	195	1.1	1.1	1.1	7.0	7.8	9.0
Seventh	223	1.1	1.1	1.1	6.8	7.5	8.3
Total		7.7	7.7	7.7	134.1	136.6	127.0

Table 5.2.2 : Water use by perennial ryegrass (1992) from different saline water tables

Description of components		C/E Water table depth (cm)				
Cut	Days	60	90	120		
First	62	0.078	0.064	0.061		
Second	83	0.174	0.126	0.138		
Third	111	0.214	0.181	0.142		
Fourth	139	0.390	0.328	0.311		
Fifth	167	0.716	0.633	0.684		
Sixth	195	0.842	0.858	0.878		
Seventh	223	0.838	0.853	0.867		
Overall		0.279	0.265	0.286		

Table 5.2.2: (continued)

Description of components		Soil water depletion, dW (mm) Water table salinity (dS/m)			Groundwater contribution, C (mm) Water table salinity (dS/m)		
Cut	Day	0.4	7.5	15	0.4	7.5	15
C1(All)	45	52.01	30.96	9.66	1.30	3.55	2.65
C2S1	60	23.50			1.35		
C2S2	73		25.92			4.45	
C3S1	81	37.10			1.55		
C4S1	96	18.41			2.60		
C3S2/C2	101		22.56	9.78		6.25	12.10
C5S1	111	14.77			3.45		
C6S1	126	11.69			5.40		
C4S2	129		13.88			8.70	
C7S1	141	2.45			7.70		
C8S1	156	0.00			6.45		
C5S2/C3	157		2.08	3.73		7.50	16.85
C9S1	171	0.00			6.40		
C10S1	186	0.00			5.15		
C11S1	201	0.00			4.65		
C4S3	213			0.00			9.35
C12S1	216	0.00			3.40		
Total		159.93	95.40	23.17	49.40	30.45	40.95

* C & S stand for different crop cuts & salinities respectively

Table 5.2.3 : Water use by perennial ryegrass (1993) for different salinity treatments with equal water table depth.

Description of components		Irrigation (mm)			Total water use , E (mm)		
		Water table salinity (dS/m)			Water table salinity (dS/m)		
Cut	Day	0.4	7.5	15	0.4	7.5	15
C1(All)	45	2.20	2.20	2.20	55.51	36.71	14.51
C2S1	60	1.10			25.95		
C2S2	73		1.10			31.47	
C3S1	81	1.10			39.75		
C4S1	96	1.10			22.11		
C3S2/C2	101		1.10	1.10		29.91	22.98
C5S1	111	1.10			19.32		
C6S1	126	1.10			18.19		
C4S2	129		1.10			23.68	
C7S1	141	1.10			11.25		
C8S1	156	1.10			7.55		
C5S2/C3	157		1.10	1.10		10.68	21.68
C9S1	171	1.10			7.50		
C10S1	186	1.10			6.25		
C11S1	201	1.10			5.75		
C4S3	213			1.10			10.45
C12S1	216	1.10			4.50		
Total		14.30	6.60	5.50	223.63	132.45	69.62
* C & S stand for different crop cuts & salinities respectively							

Table: 5.2.3 (continued)

Description of Components		C/E		
Cut	Day	Water table salinity (dS/m)		
		0.4	7.5	15
C1(All)	45	0.023	0.097	0.183
C2S1	60	0.052		
C2S2	73		0.141	
C3S1	81	0.039		
C4S1	96	0.118		
C3S2/C2	101		0.209	0.527
C5S1	111	0.179		
C6S1	126	0.297		
C4S2	129		0.367	
C7S1	141	0.684		
C8S1	156	0.854		
C5S2/C3	157		0.702	0.777
C9S1	171	0.853		
C10S1	186	0.824		
C11S1	201	0.809		
C4S3	213			0.895
C12S1	216	0.756		
Overall		0.221	0.230	0.588
* C & S stand for different crop cuts & salinities respectively				

Table: 5.2.3 (continued)

However, the groundwater contribution in the present investigation was only around 25 to 30 %.(0.25 to 0.3 mm d^{-1} under average climatic demand of 2.0 mm d^{-1}) of the total water use. The rest of the total water use was attributed to pre-irrigation as soil water reserve. In a climate of high evaporative demand, the soil water reserve by pre-irrigation would be exhausted shortly. Hence, only pre-irrigation along with water table contribution is not sufficient to fulfil the water needs by the entire cropping period.

5.3 Salt accumulation in the root zone

In order to understand the salinity effect, it is important to consider salt accumulation within the soil profile, particularly the salt balance within the root zone. The transport of salt in the soil profile was estimated based on the convective flow of water and the concentration of soil solution.

The measured and the estimated salt profiles at different depths and at different times in the lysimeters at different times are shown in Figs. 5.3.1(a),(b) & (c); 5.3.2(a),(b) & (c) and 5.3.3(a),(b) & (c) for lettuce, ryegrass'92 & ryegrass'93 experiments, respectively. These Figures show that, increases in salinity above the initial salinity were mainly confined to 0-15 cm from the soil surface in all experiments.

The magnitude of salinization in the top 5 cm surface increased the initial salinity of the soil water by 3 to 4 fold of initial salinity of soil water. Initially, salinization built up fast, later slowly and, finally stopped when total potential (matric + osmotic) became equivalent to -1.5 to -2.0 MPa. The salinization below the top cm depth was less than double. Also, Figs. 5.3.3(c) and 5.3.2(b) show that though the initial salinity was around double (15.0 dS m^{-1} and 9.4 dS m^{-1}), the maximum salinization in the top surface in the 15.0 dS m^{-1} treatment was smaller (35.0 and 42.0 dS m^{-1}) within almost equal cropping periods. The weighted average of top 60 cm root zone salinities in all the treatments were within the range of 2 fold of initial salinity over the cropping periods greater than 200 days, where no surface water was applied as irrigation. Even in case of highly saline (15.0 dS m^{-1}) treatment, the average root zone salinity built up less than 2 fold.

It is found from Figs. 5.3.2(a),(b) & (c) that salinization almost stopped after 139 days from the sowing date. In the latter 84 days of cropping, it was likely that there would be a reduction of salinity in the top soil layer(s) due to solute diffusion to the lower layer(s), down to the concentration gradient, but the measured values of salinity did not indicate so.

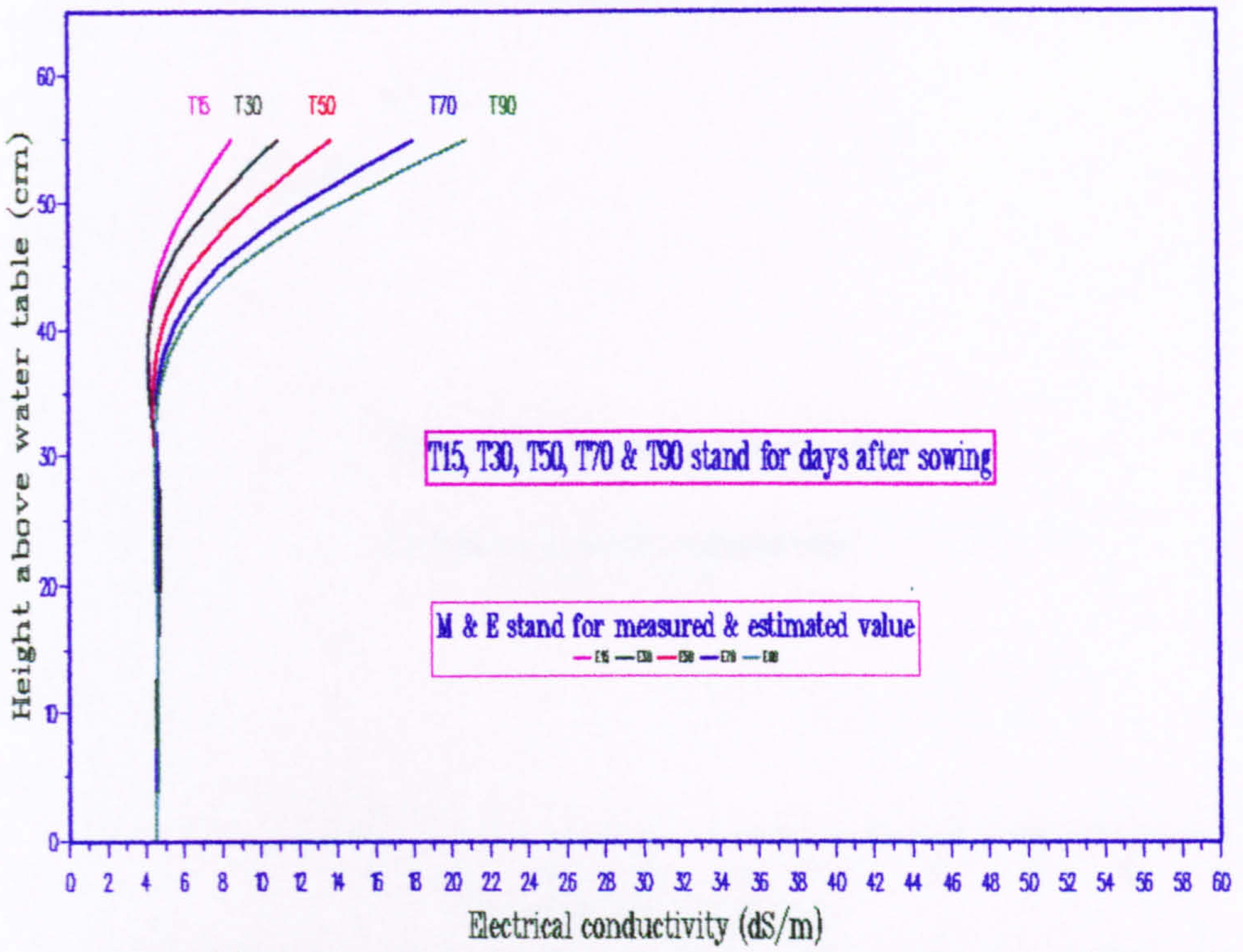


Fig. 5.3.1(a): Salt profile at diff. depths in WT-60 lysimeter (lettuce)

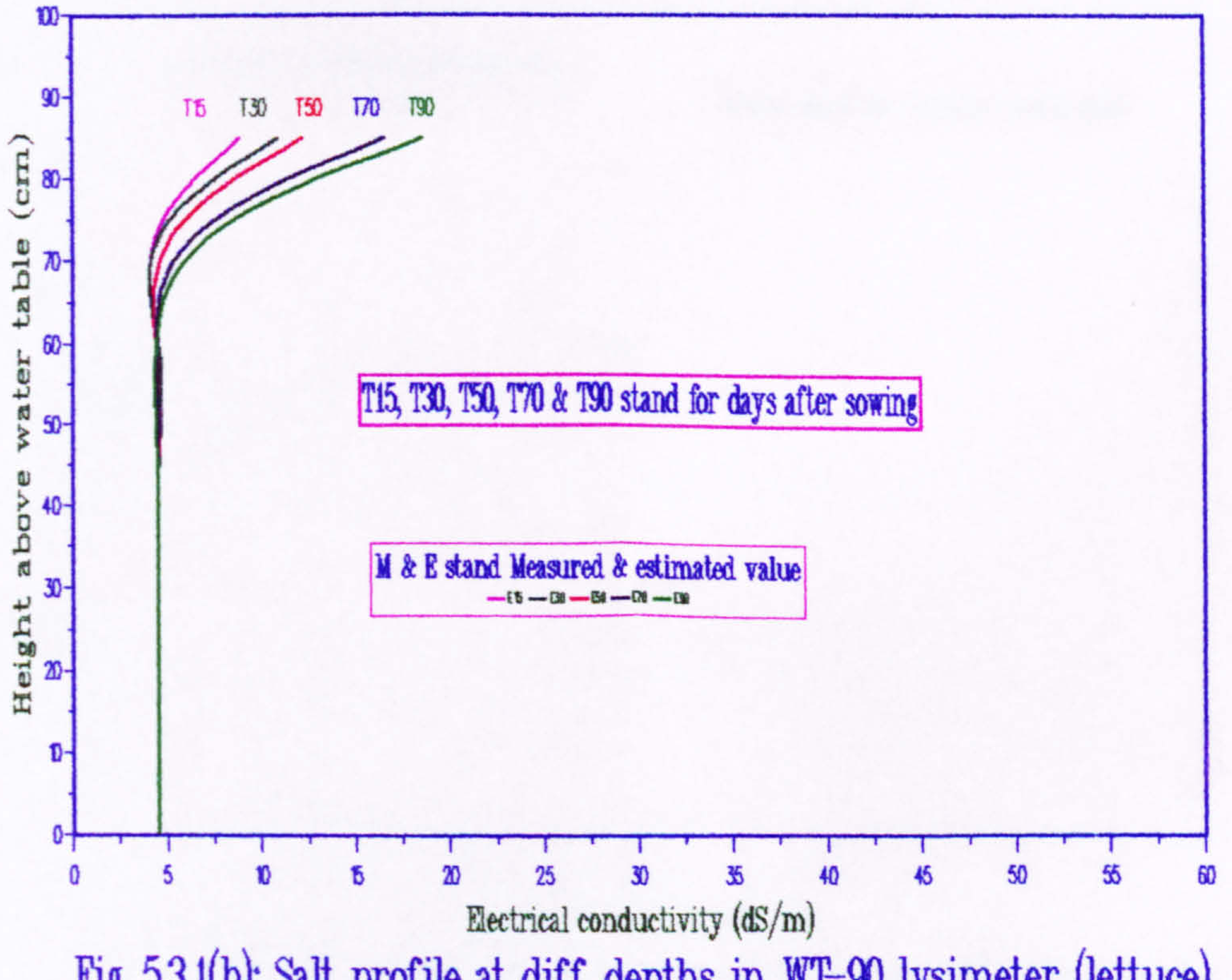


Fig. 5.3.1(b): Salt profile at diff. depths in WT-90 lysimeter (lettuce)

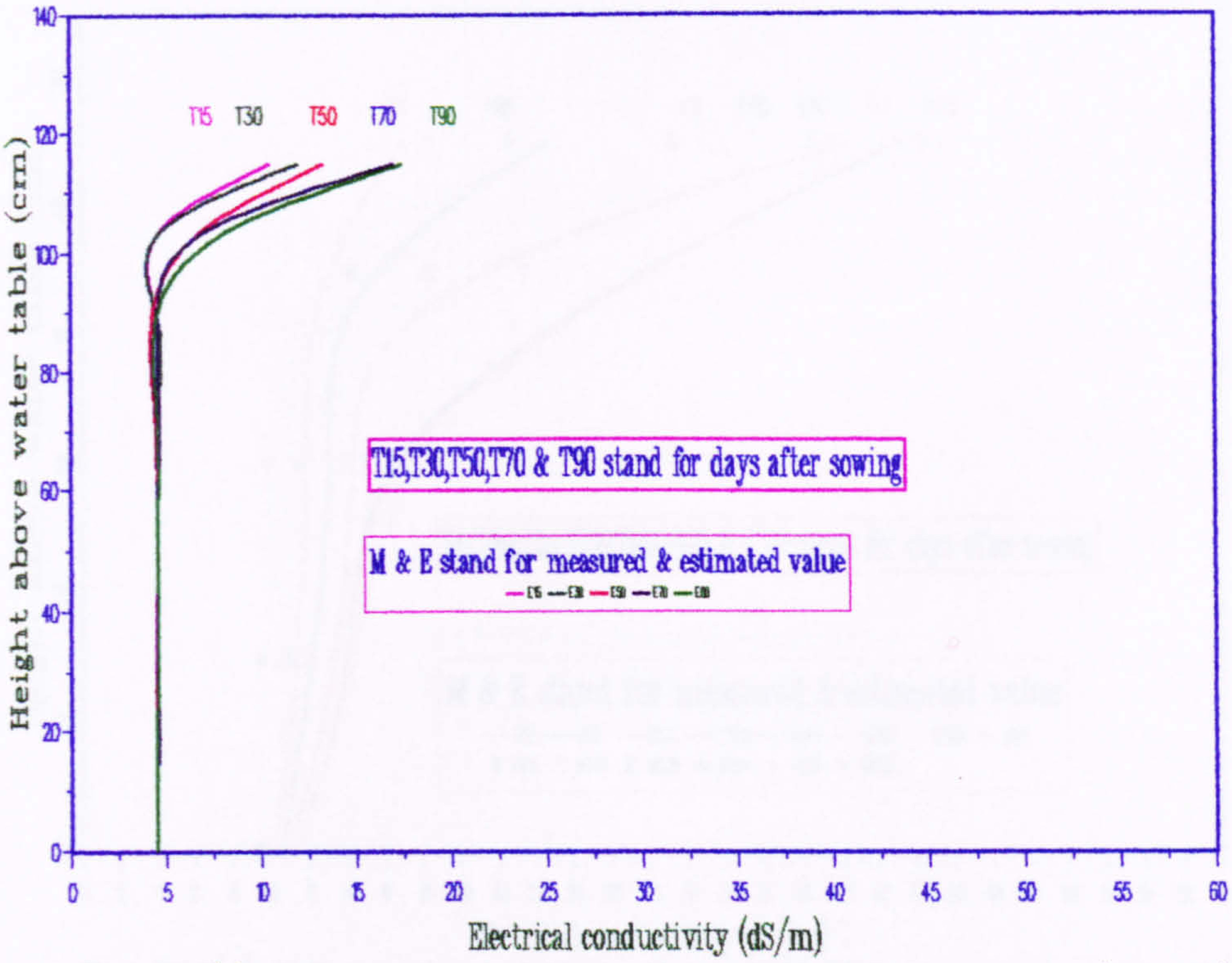


Fig. 5.3.1(c): Salt profile at diff. depths in WT-120 lysimeter (lettuce)

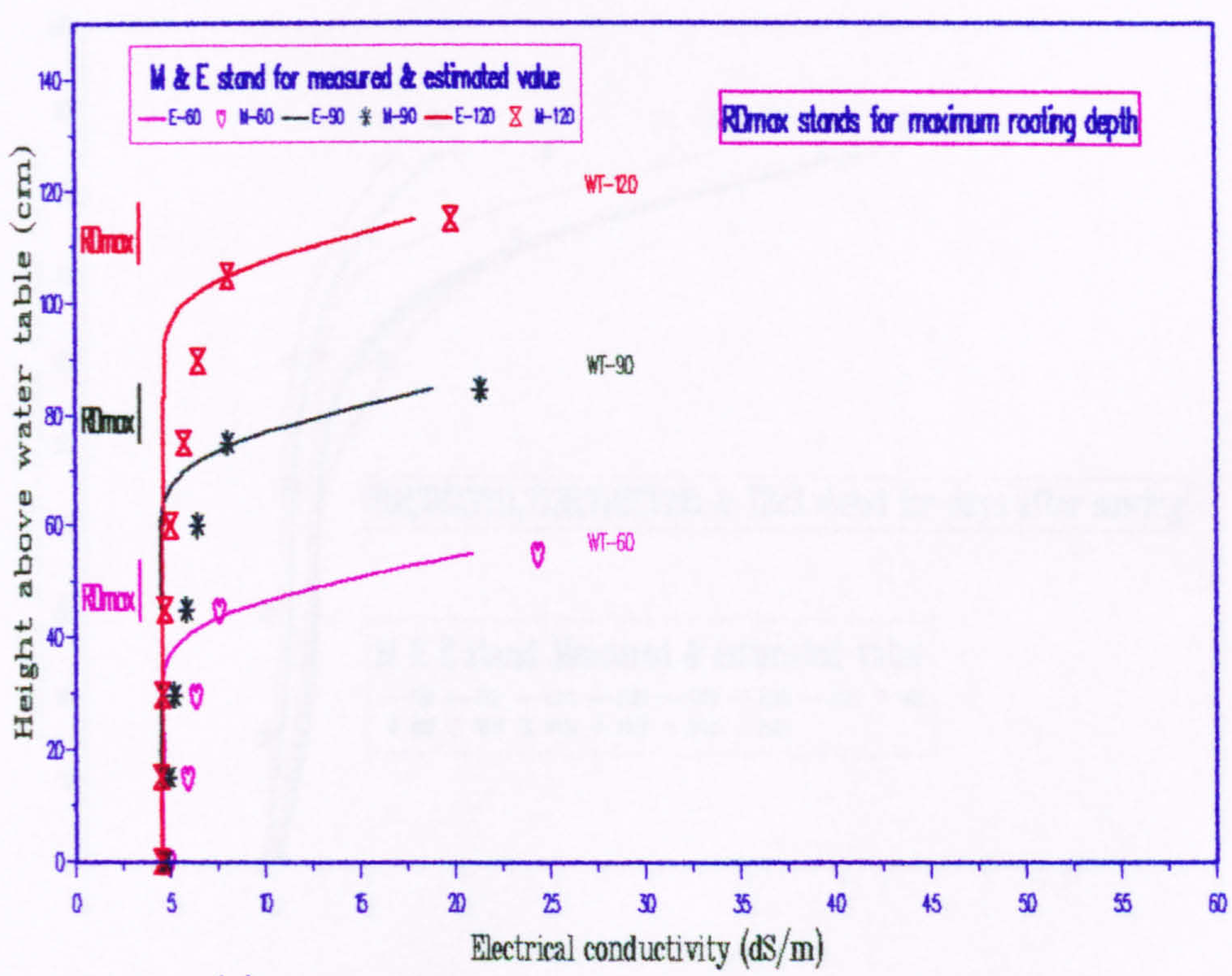


Fig. 5.3.1(d): Salt profile at harvest for diff. water table treatments

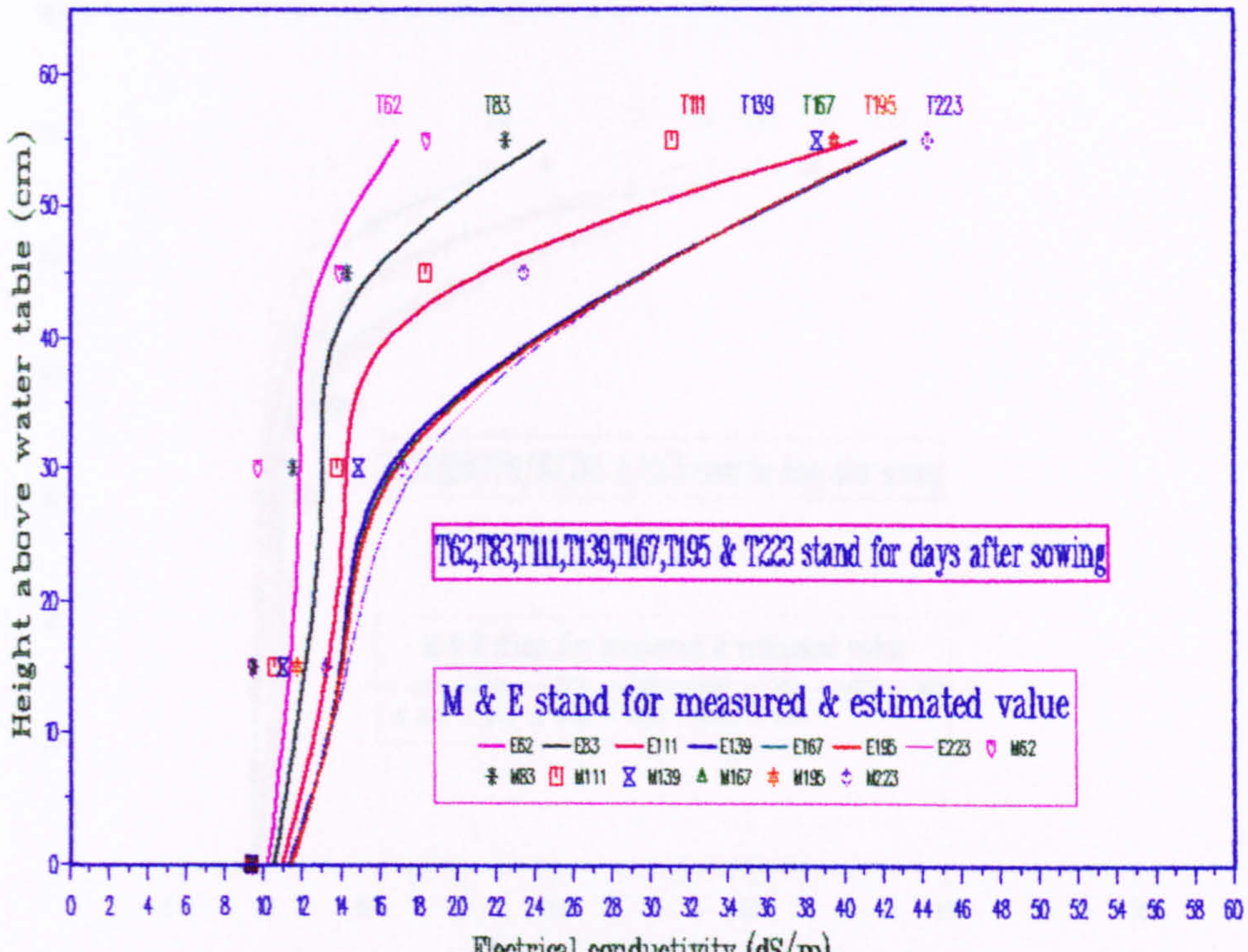


Fig. 5.3.2(a): Salt profile at diff. depths in WT-60 lysimeter (ryegrass'92)

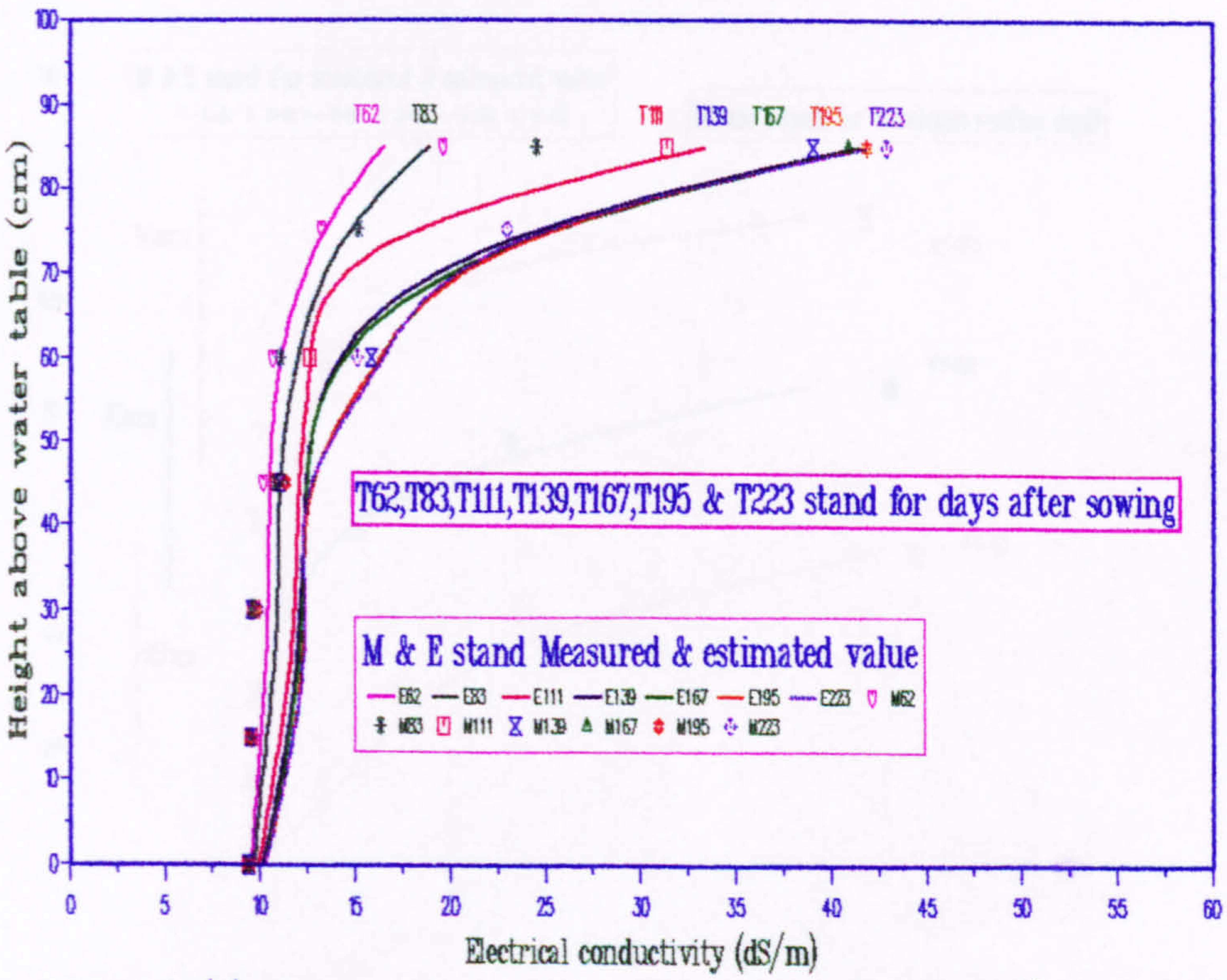


Fig. 5.3.2(b): salt profile at diff. depths in WT-90 lysimeter (ryegrass'92)

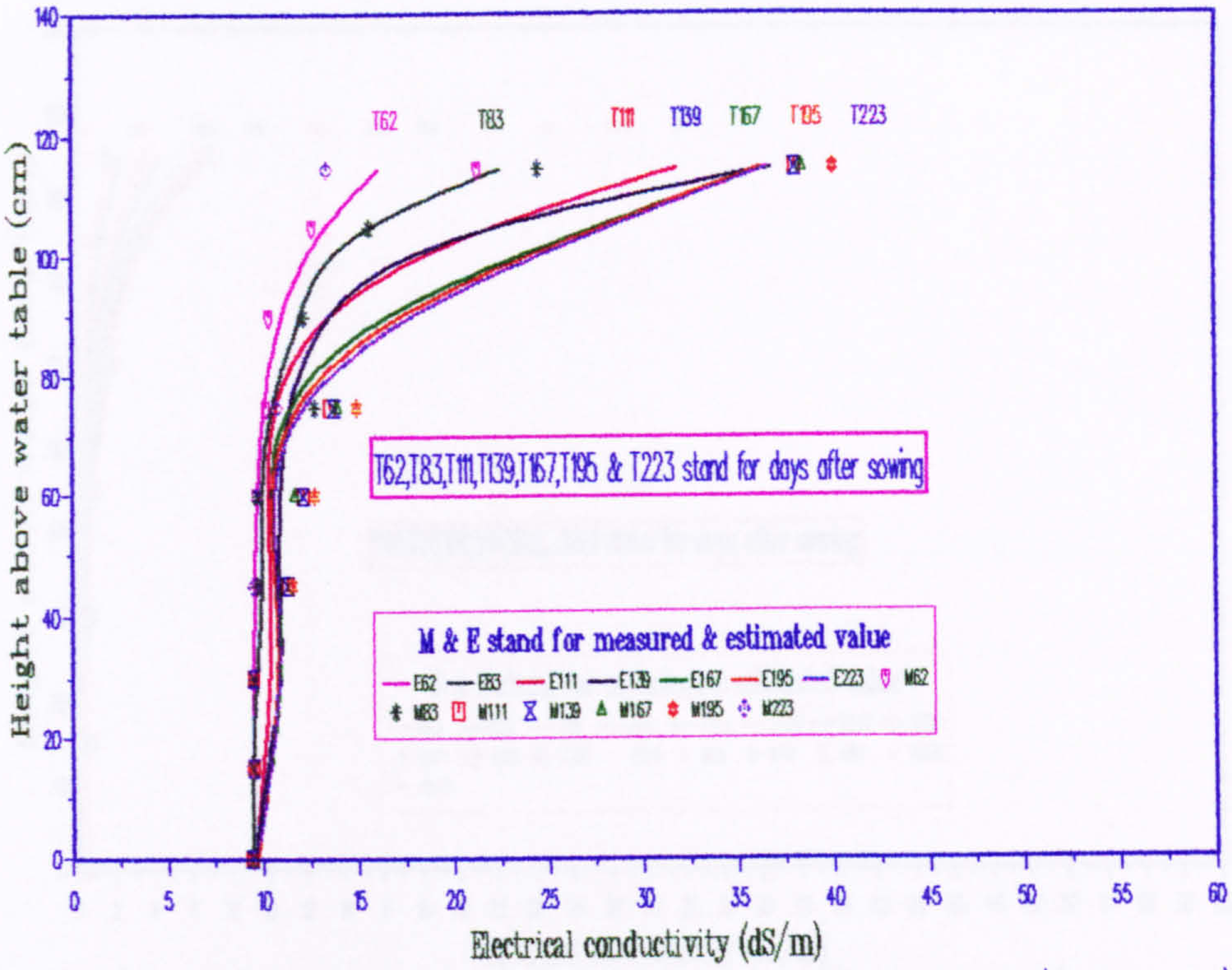


Fig. 5.3.2(c): Salt profile at diff. depths in WT-120 lysimeter (ryegrass'92)

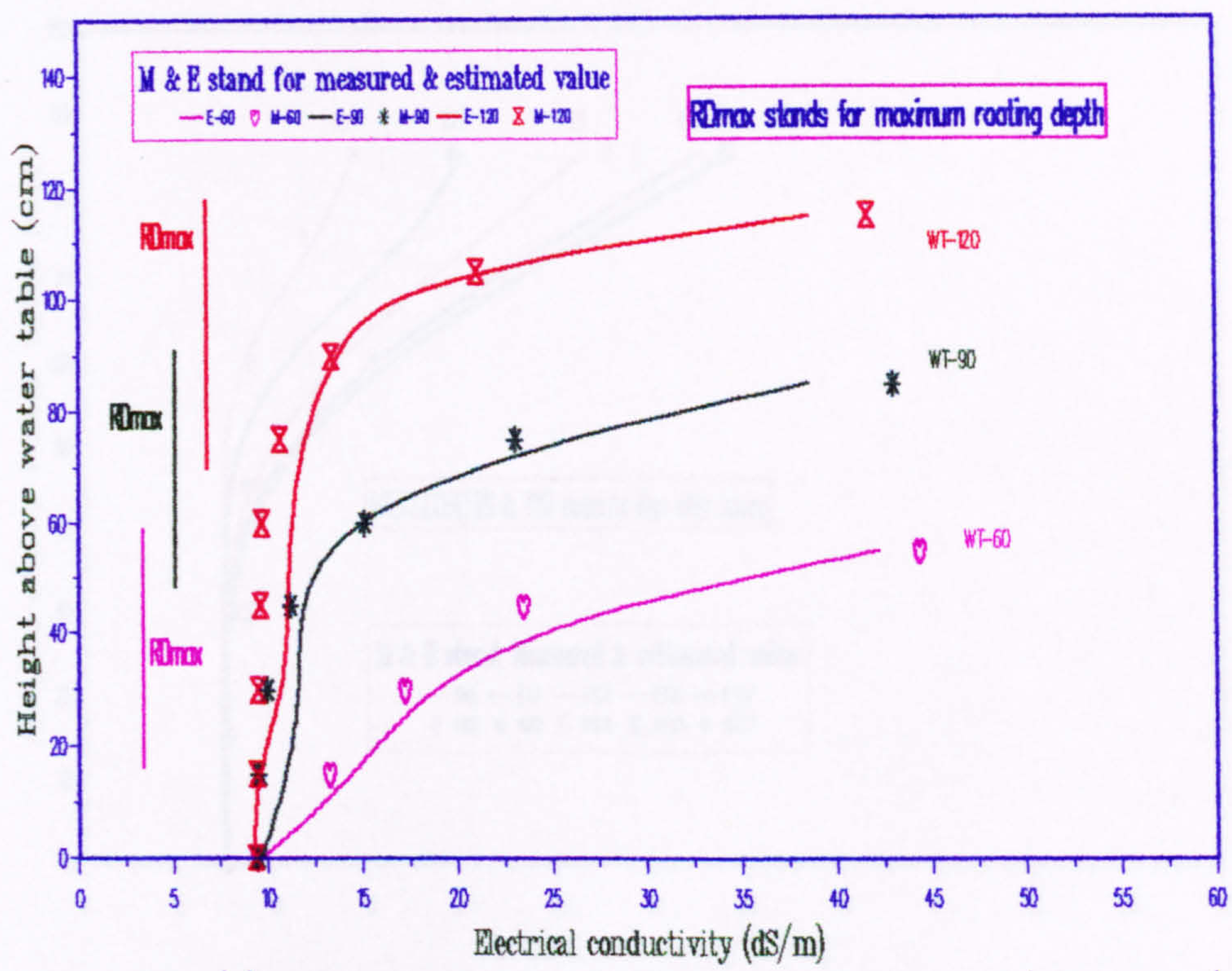


Fig. 5.3.2(d): Salt profile at harvest for diff. water tables (ryegrass'92)

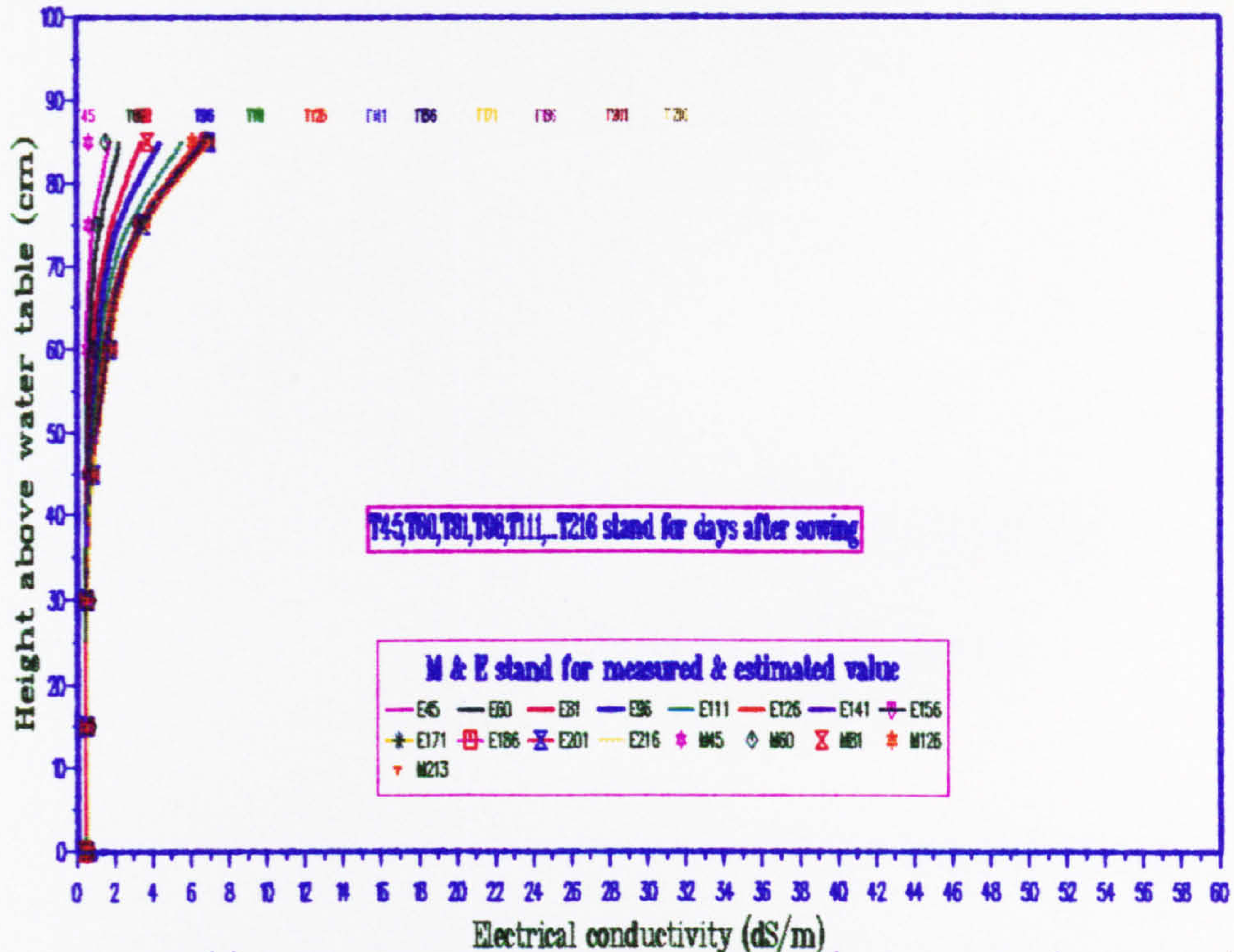


Fig. 5.3.3(a): Salt profile at diff. depths in 0.4 dS/m lysimeter (ryegrass'93)

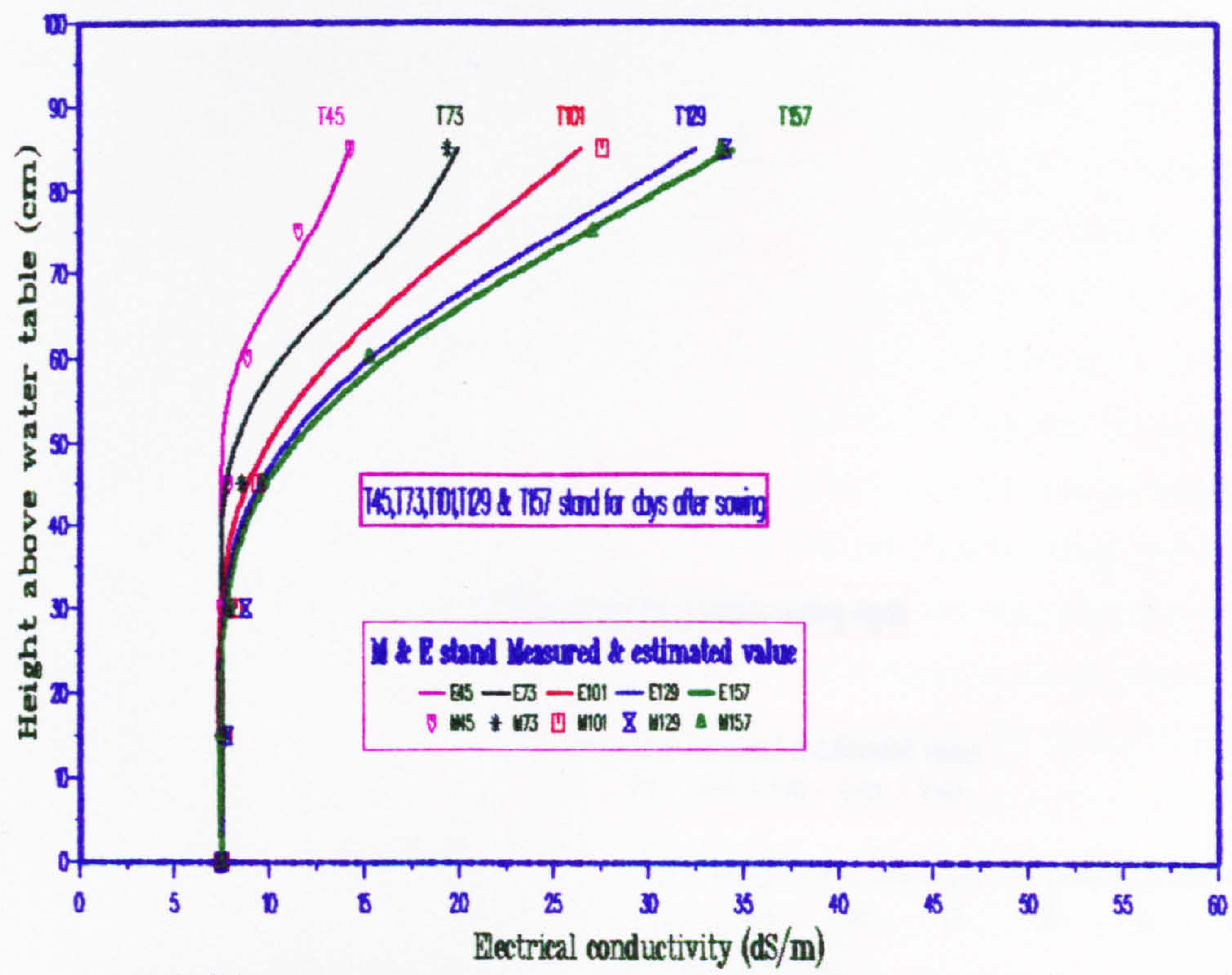


Fig. 5.3.3(b): Salt profile at diff. depths in 7.5 dS/m lysimeter (ryegrass'93)

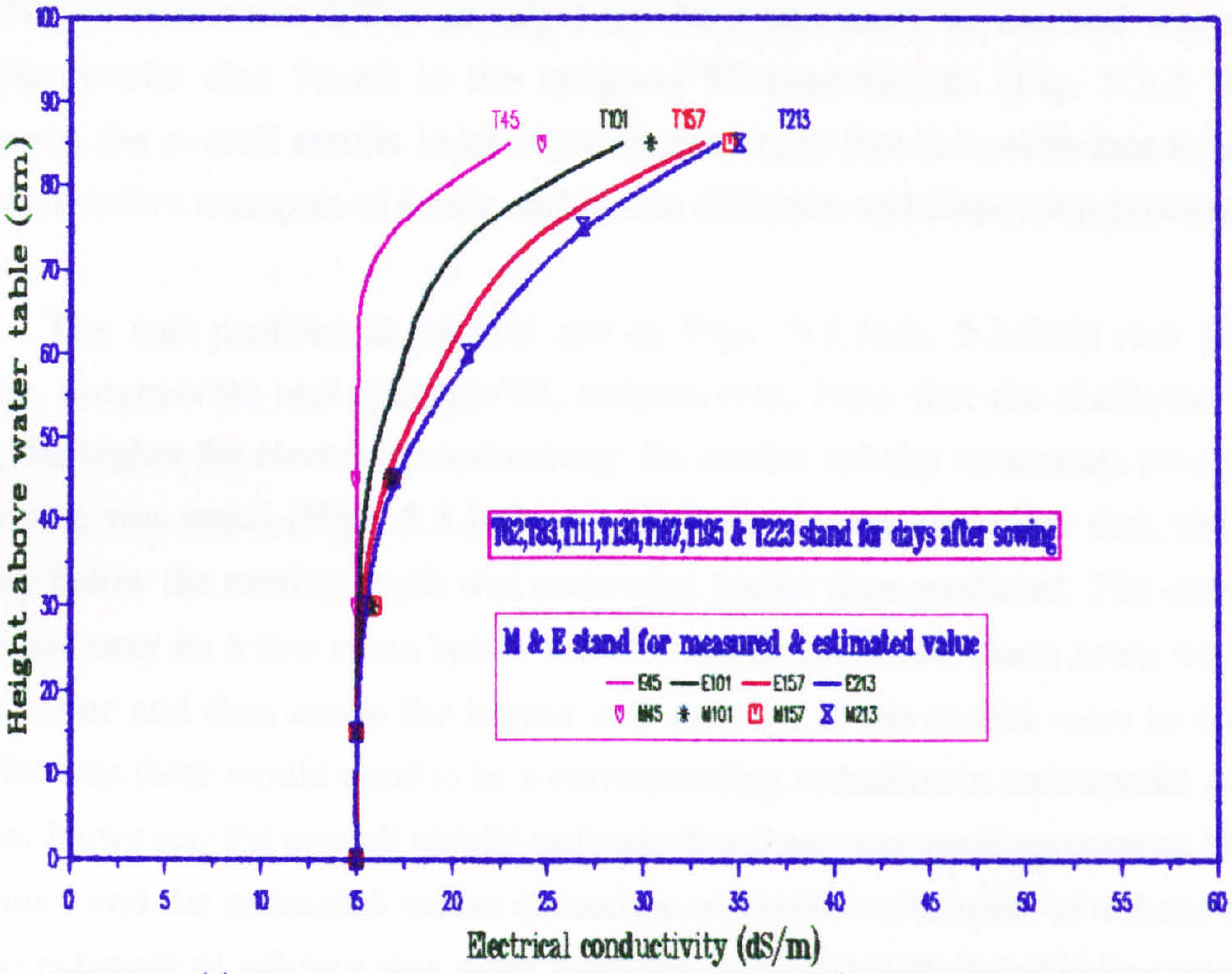


Fig. 5.3.3(c): Salt profile at diff. depths in 15.0 dS/m lysimeter (ryegrass'93)

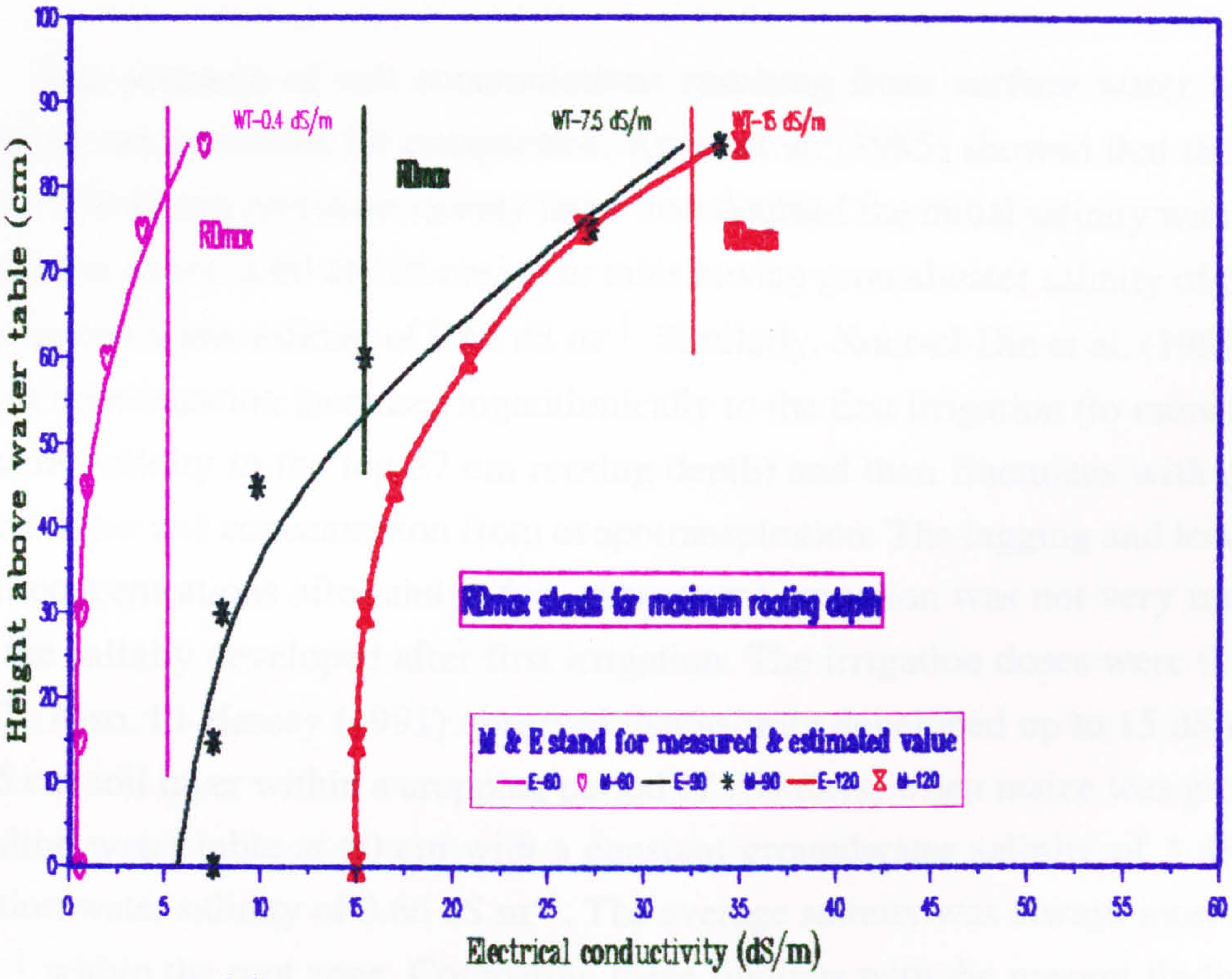


Fig. 5.3.3(d): Salt profile at harvest for diff. salinity treatments

The rate of molecular diffusion might not be pronounced as the soil was rather dry. Similar results also found in the ryegrass'93 experiments [Fig. 5.3.3 (b) & (c)]. However, the overall results in all experiments imply that it is sufficient to account for only convective transport of solute and ignore diffusion and dispersion processes.

The salt profiles at harvest are in Figs. 5.3.1(d), 5.3.2(d) and 5.3.3(d) for lettuce, ryegrass'92 and ryegrass'93, respectively. Note that the shallower the water table, the higher the electrical conductivity for similar salinity treatments though there the difference was small [Figs. 5.3.1(d) and 5.3.2(d)]. It was also clear that, the measured salinity below the rooting depth was somewhat higher than predicted. The reason may be that there may be a few roots below the root depth modelled. Such roots would extract some water and thus cause the higher soil salinity. If the model were to successfully describe this there would need to be a corresponding reduction in root uptake at shallower depths. However, the overall results indicate that there was good agreement between the measured and the estimated values (based on convective transport of solute) of salinity. As the estimate of salinity was done from the mass balance, it could be concluded that simple consideration of convective flow is a reasonable basis for modelling solute movement in designing irrigation and drainage management practices.

The scenario of salt accumulations resulting from surface water application conditions are presented for comparison. Kruse et al. (1985) showed that the weighted average of 0-60 cm root zone salinity more than doubled the initial salinity within 10 days of irrigation above a 60 or 105 cm water table having groundwater salinity of 6.0 dS m^{-1} and irrigation water salinity of 0.66 dS m^{-1}. Similarly, Nour-el-Din et al. (1987) reported that salt concentration increases logarithmically to the first irrigation (to more than twice the initial salinity in the top 57 cm rooting depth) and then fluctuates with dilution of applied water and concentration from evapotranspiration. The lagging and leading range of salt concentrations after and before subsequent irrigation was not very much varied from the salinity developed after first irrigation. The irrigation doses were 0.7, 0.8 and 0.9 ET. Also, El-Hassey (1991) observed that salinity developed up to 15 dS m^{-1} in the top 15 cm soil layer within a cropping period of 155 days, when maize was grown above the saline water table at 60 cm with a constant groundwater salinity of 3 dS m^{-1} and irrigation water salinity of 0.66 dS m^{-1}. The average salinity was always more than the 6 dS m^{-1} within the root zone. Comparing these findings with the present findings on the trend of root zone salinization from sub-irrigation without surface water application, the present management practice may be better than irrigation with highly saline water, unless climatic demand forces irrigation.

5.4 Capillary rise characteristics

The position of the zero-flux plane is identified as the depth at which the hydraulic head gradient is zero (Arya et al., 1975). The position of the zero-flux plane separates upward and downward flux in a soil profile in the presence or absence of a water table and allows the water movement to be partitioned between evaporation and drainage. Therefore, in order to partition between the negative change of moisture content (depletion) and the water table contribution, the zero-flux plane should be at water table unless the crop roots touch the water table or capillary fringe of water table.

Hydraulic head gradients at different depths for lettuce, ryegrass'92 and ryegrass'93 were determined from the tensiometers readings and plotted against time from sowing, [Figs. 5.4.1(a), (b) & (c); 5.4.2(a), (b) & (c) and 5.4.3(a), (b) & (c) respectively). These Figures show that no drainage occurred during the cropping periods indicating that equilibrium soil-water profiles were developed before sowing and hence that the zero-flux planes were at the water tables. They also show that no significant changes in the hydraulic head gradients were found below 30 cm depth and that a low hydraulic gradient occurred even in the upper part of the soil profile in the 15.0 dS m^{-1} treatment. Fig. 5.4.3(d) shows that there is a distinct difference between saline and nonsaline treatments, and the higher the salinity level the lower the hydraulic head gradient developed. Figs. 5.4.1(d) and 5.4.2(d) show that there was little effect of the water table depth on hydraulic head gradients. The Figures also show that, for the ryegrass experiments, there was no increase in hydraulic gradient after 140 to 150 days from sowing i.e. there was no further soil drying after that period.

The movement of water from the water tables started from the day the hydraulic gradients at the bottom soil profile (nearer to water table) started to decrease . In the lettuce and ryegrass'92 experiments, the hydraulic gradients at the bottom profile started to decrease at 10, 25 & 40 and 20, 30 & 35 days respectively from sowing in lysimeters with water tables 60, 90 and 120 cm deep respectively, as shown in Figs. 5.4.1(a), (b) & (c) and 5.4.2(a), (b) & (c). In ryegrass'93 experiments, it started at 25, 25 & 30 days with water table salinities of 0.4, 7.5 & 15.0 dS m^{-1} respectively, as shown in Fig. 5.4.3(a), (b) & (c).

The trend of the matric and osmotic stress effects on capillary rise from water tables for the different experiments are presented in Figs. 5.4.4, 5.4.5 and 5.4.6. Fig. 5.4.4 shows that, up to 150 days, the rate of capillary rise from saline water tables was

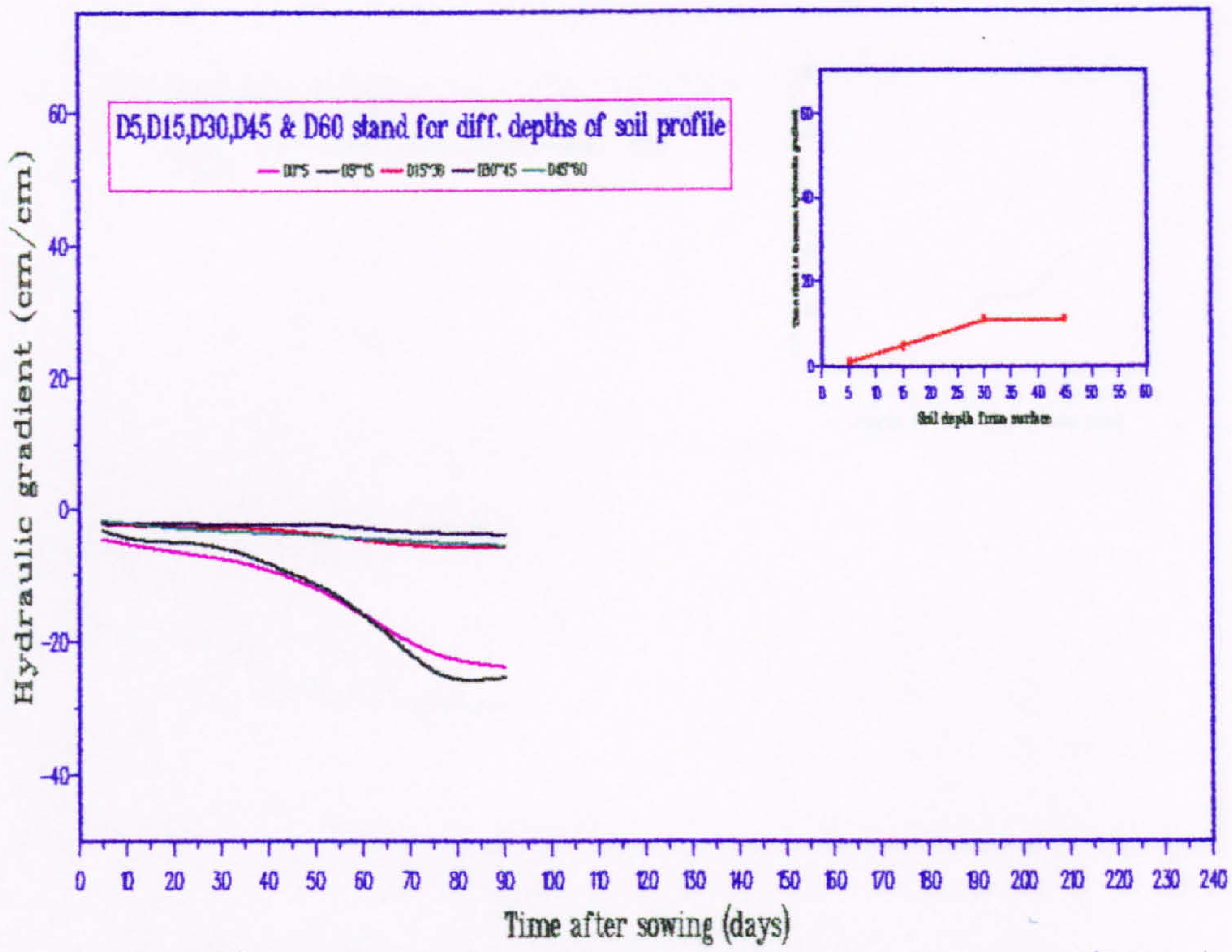


Fig. 5.4.1(a): Variation of hydraulic gradient Vs time for WT-60 (lettuce)

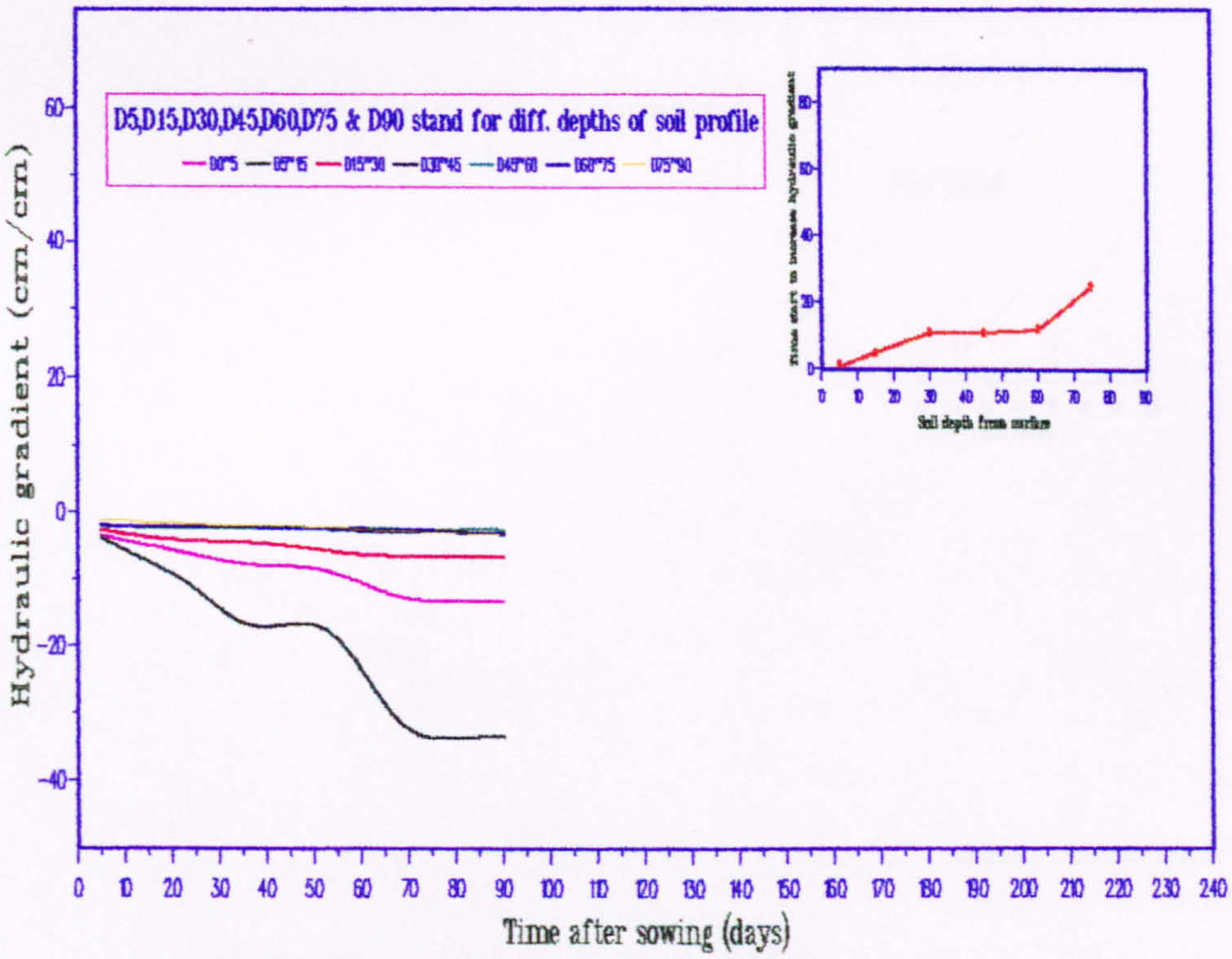


Fig. 5.4.1(b): Variation of hydraulic gradient Vs time for WT-90 (lettuce)

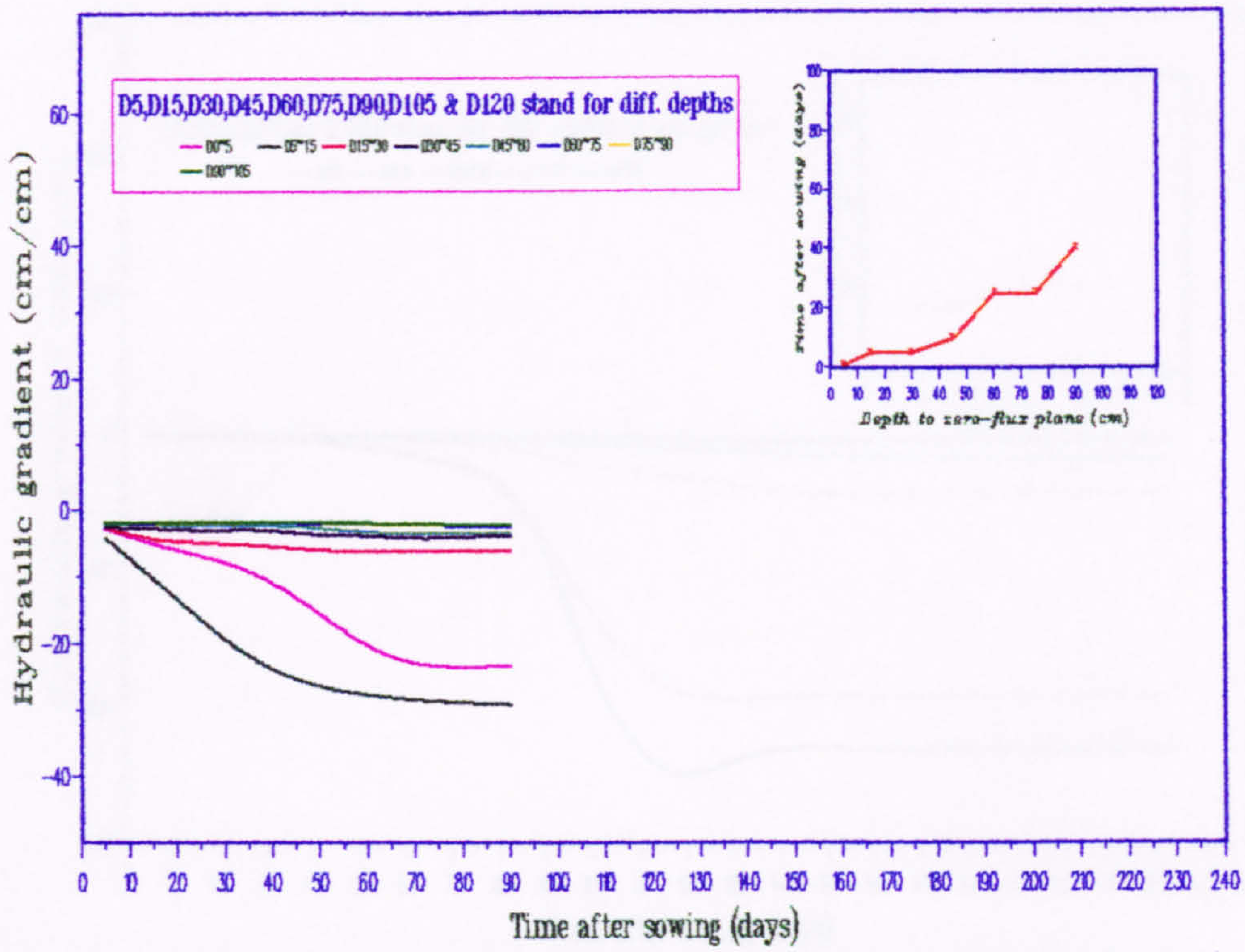


Fig. 5.4.1(c): Variation of hydraulic gradient Vs time for WT-120 (lettuce)

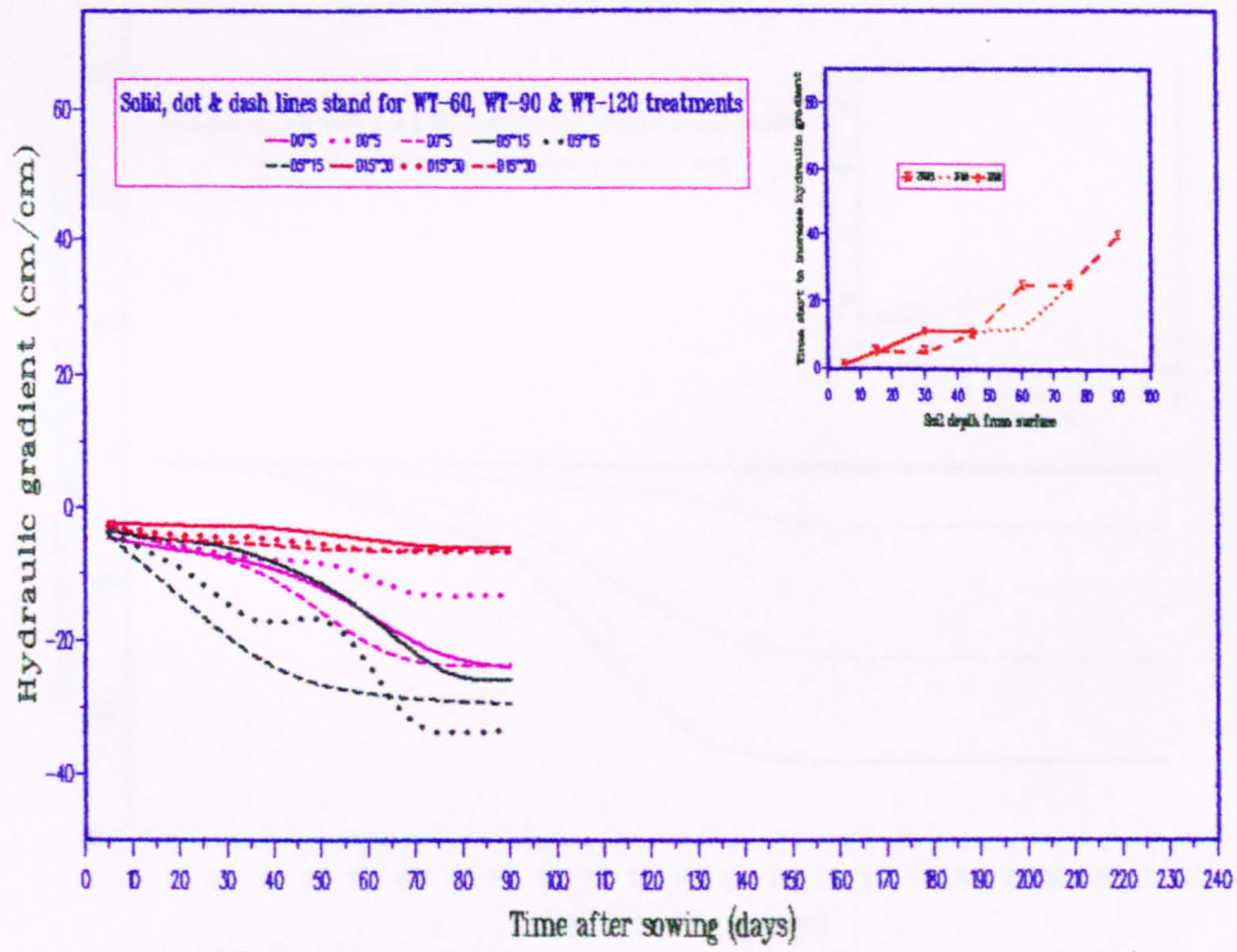


Fig. 5.4.1(d): Comparison of hydraulic gradient for diff. water tables (lettuce)

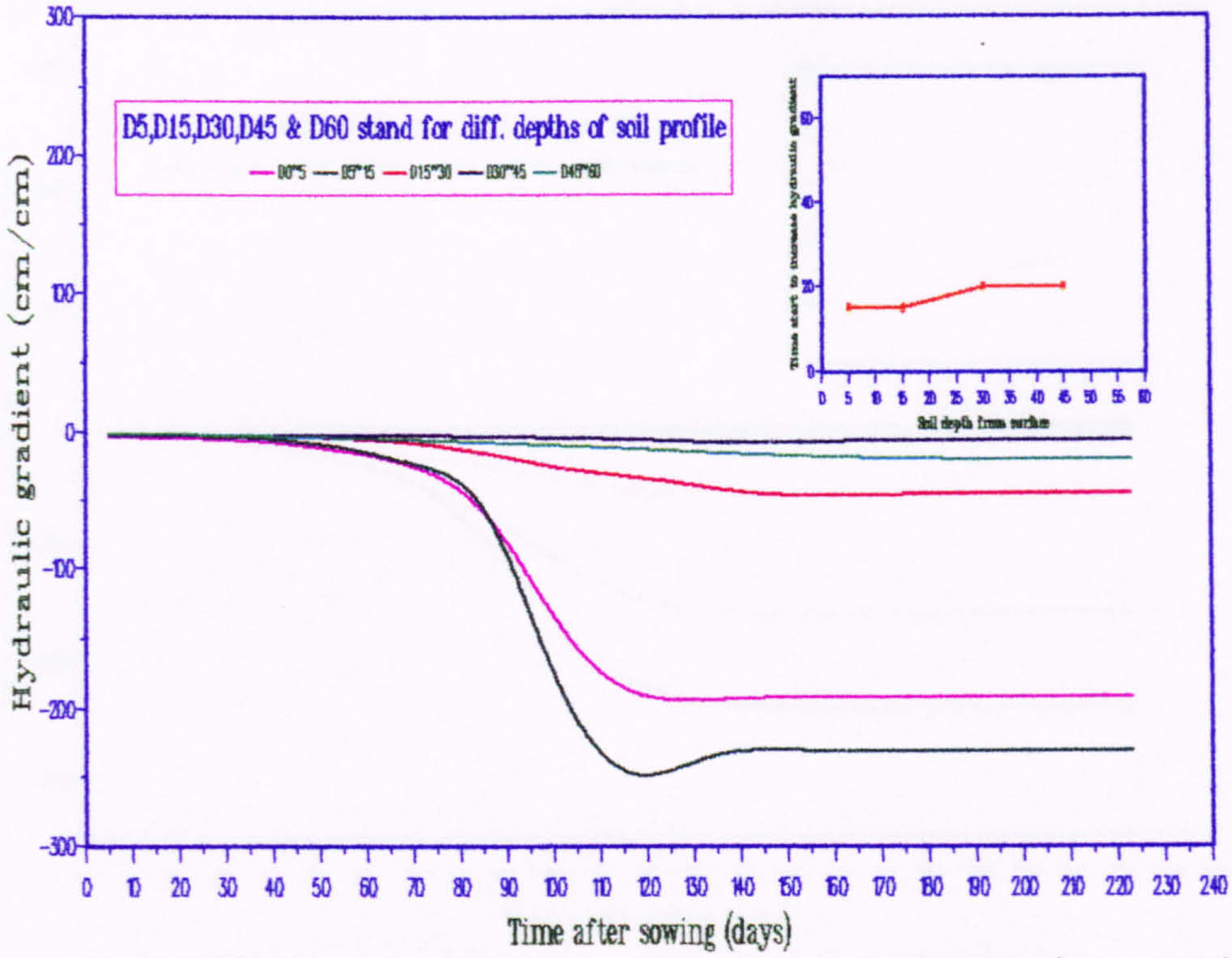


Fig. 5.4.2(a): Variation of hydraulic gradients Vs time for WT-60 (ryegrass'92)

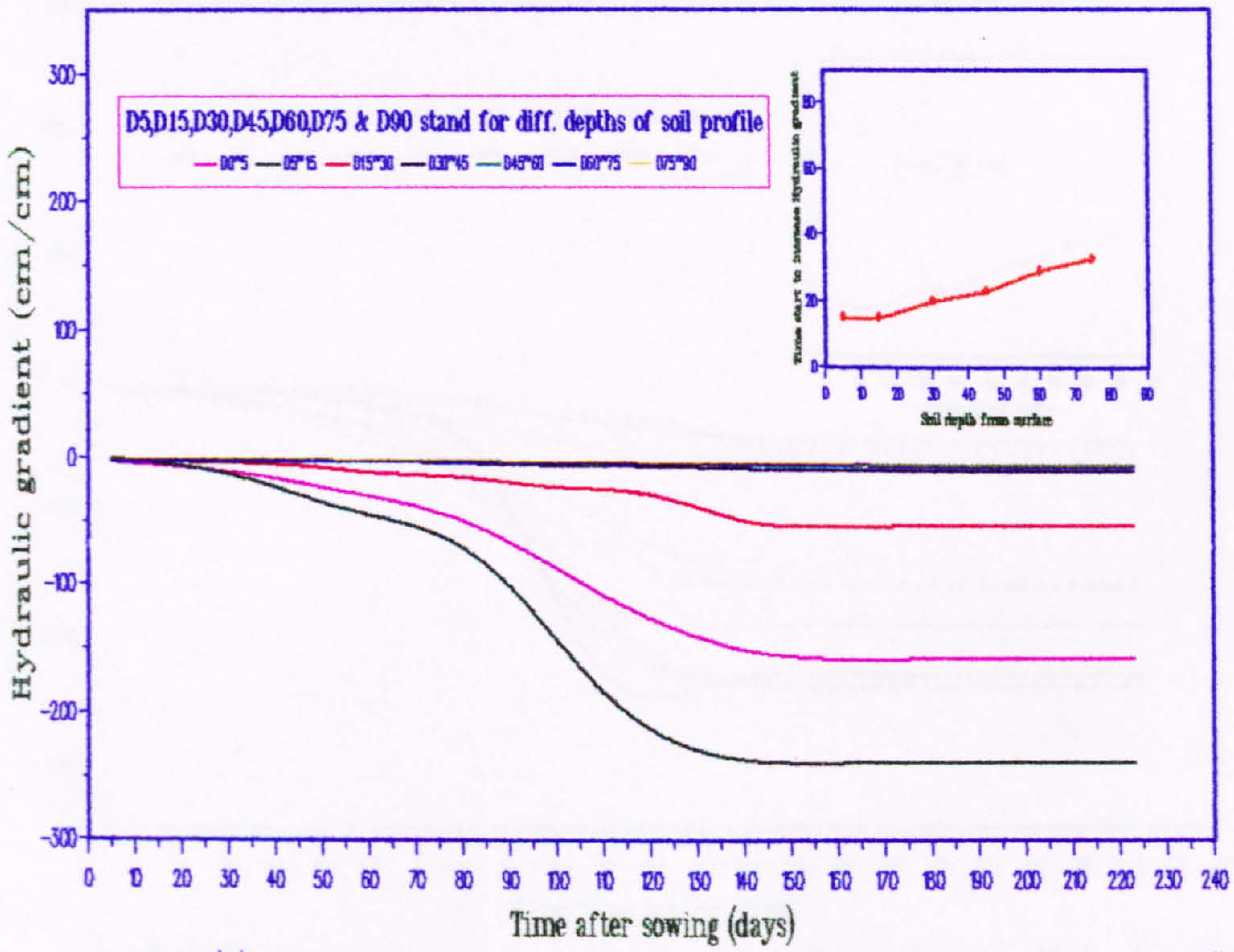


Fig. 5.4.2(b): Variation of hydraulic gradients Vs time for WT-90 (ryegrass'92)

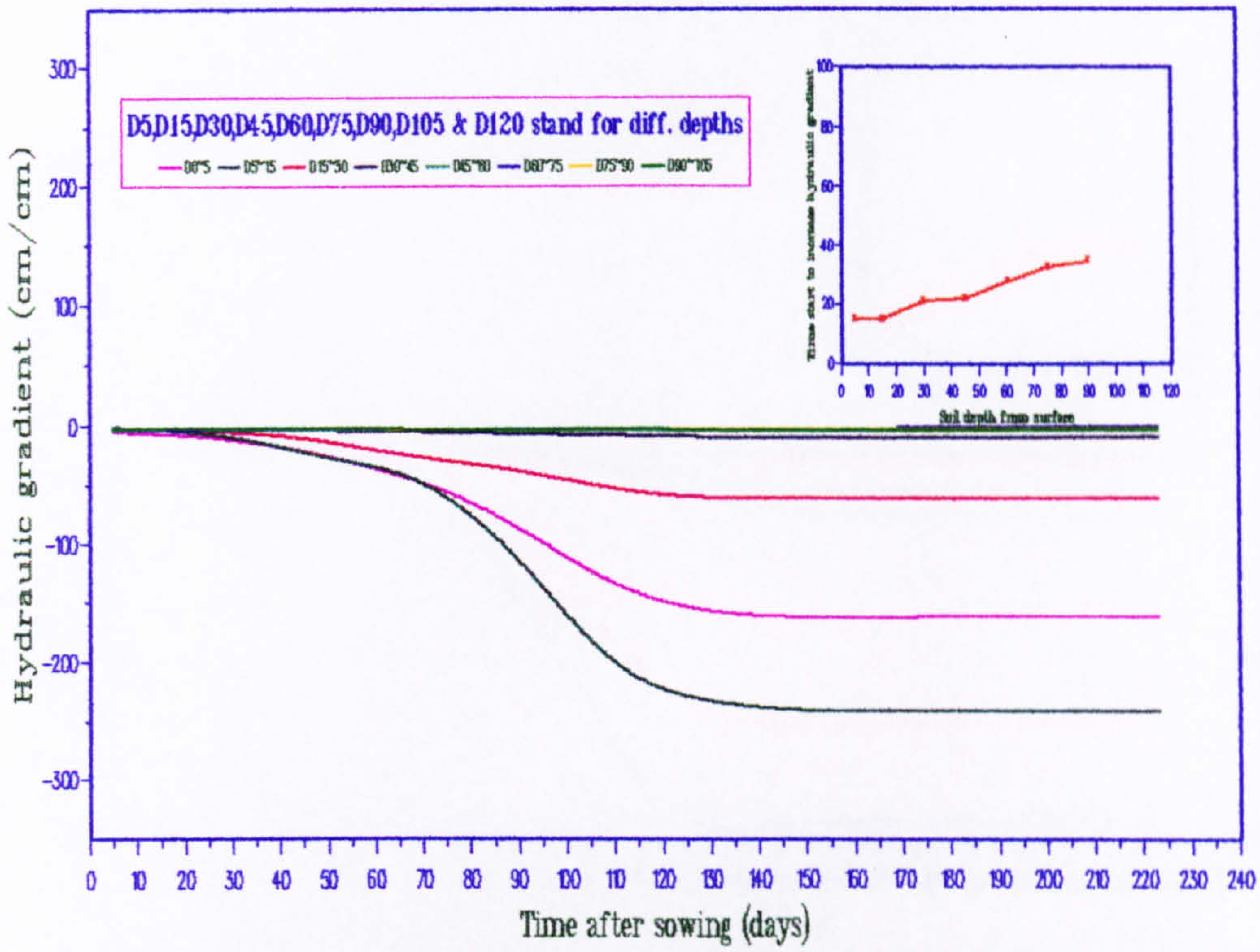


Fig. 5.4.2(c): Variation of hydraulic gradients Vs time for WT-120 (ryegrass'92)

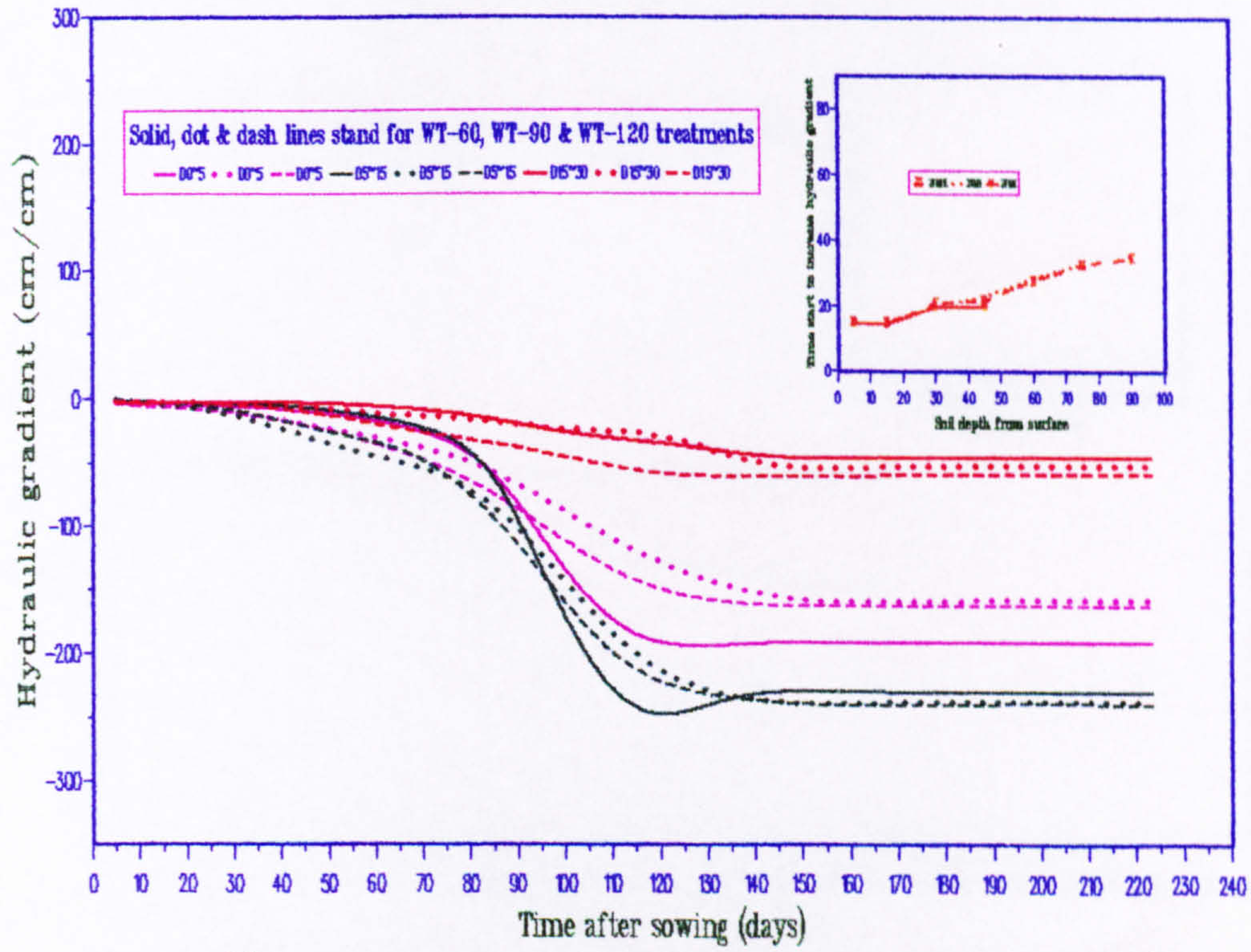


Fig. 5.4.2(d): Comparison of hydraulic gradients for diff. water tables (rye'92)

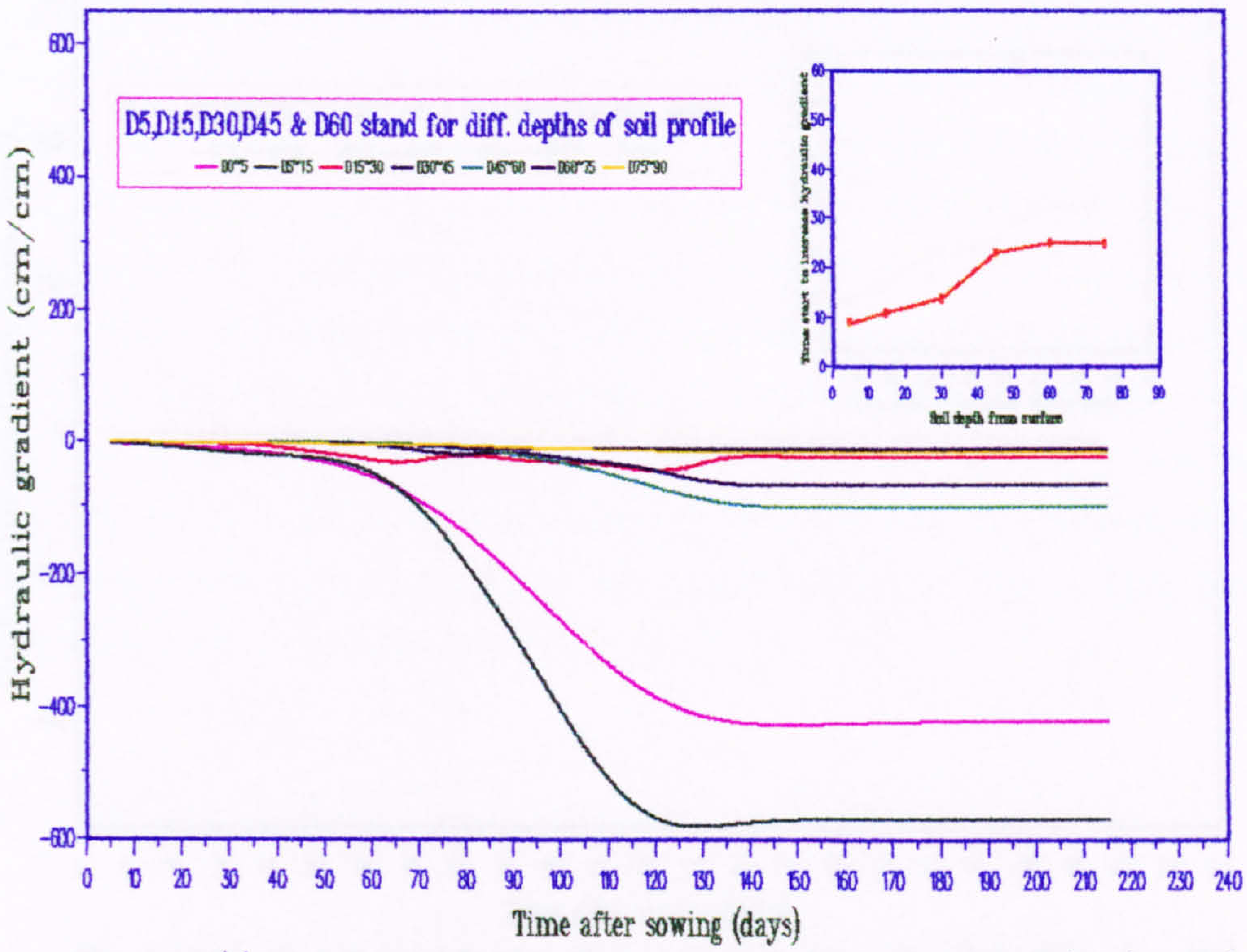


Fig. 5.4.3(a): Variation of hydraulic gradient Vs time for 0.4 dS/m (rye'93)

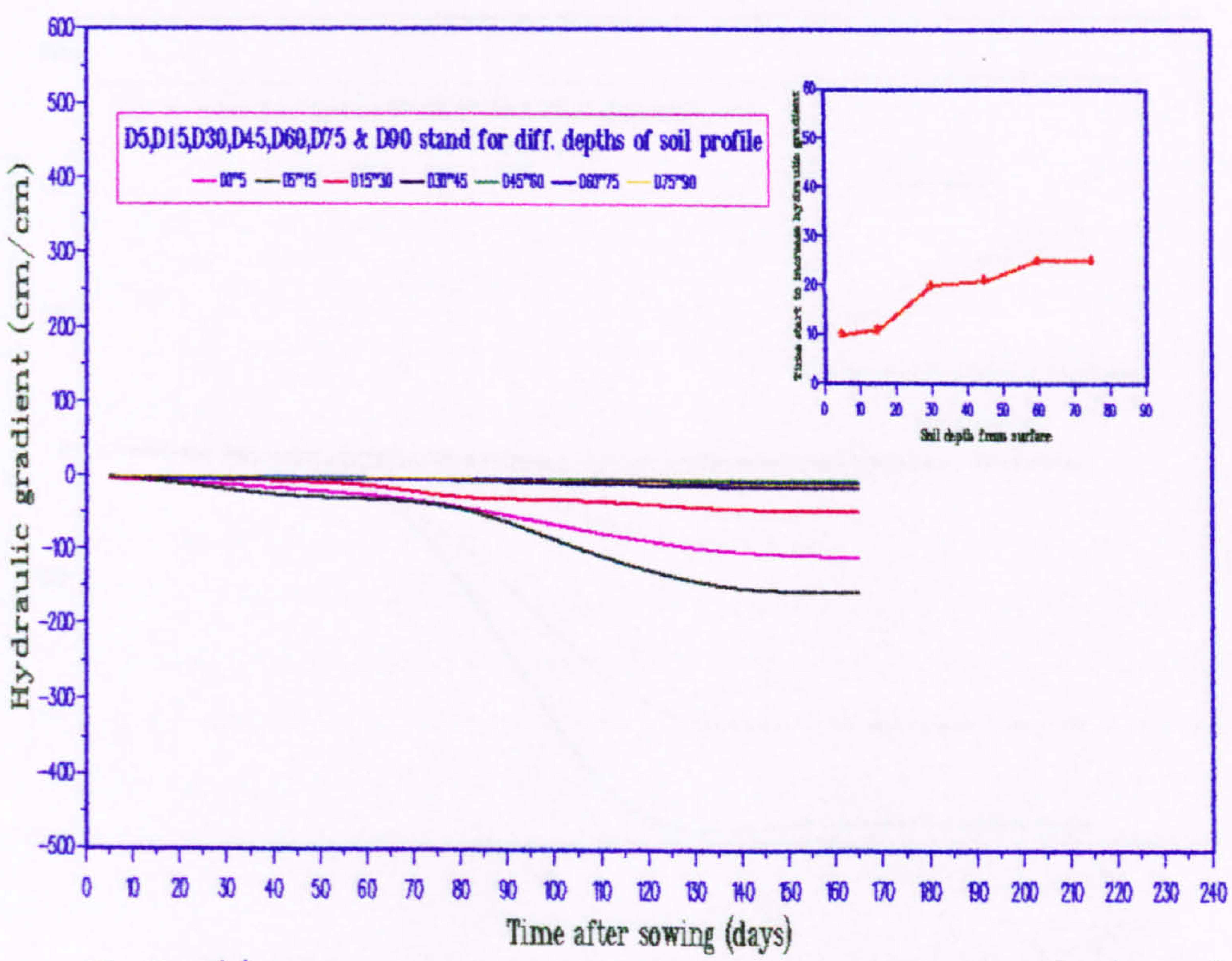


Fig. 5.4.3(b): Variation of hydraulic gradient Vs time for 7.5 dS/m (rye'93)

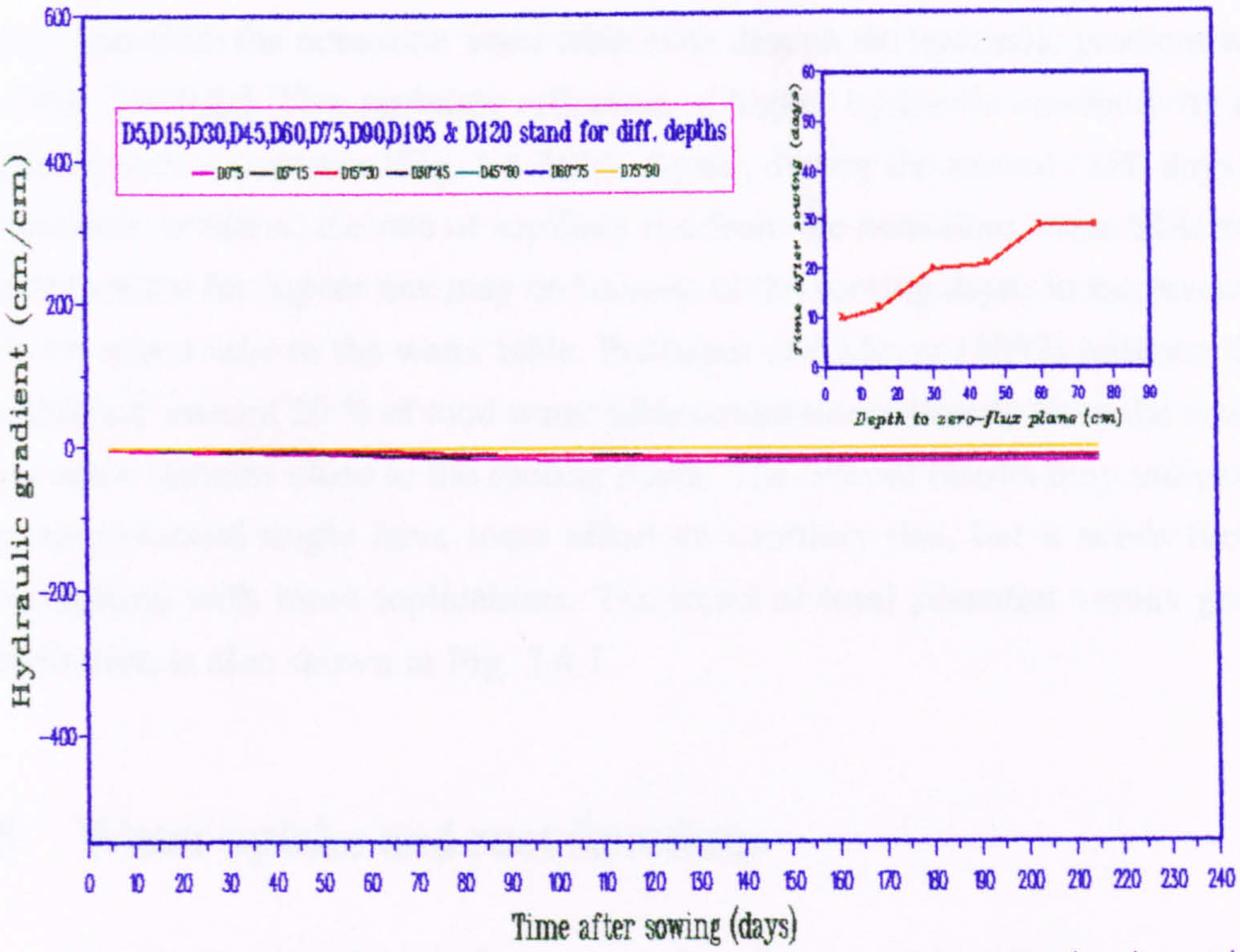


Fig. 5.4.3(c): Variation of hydraulic gradient Vs time for 15.0 dS/m (rye'93)

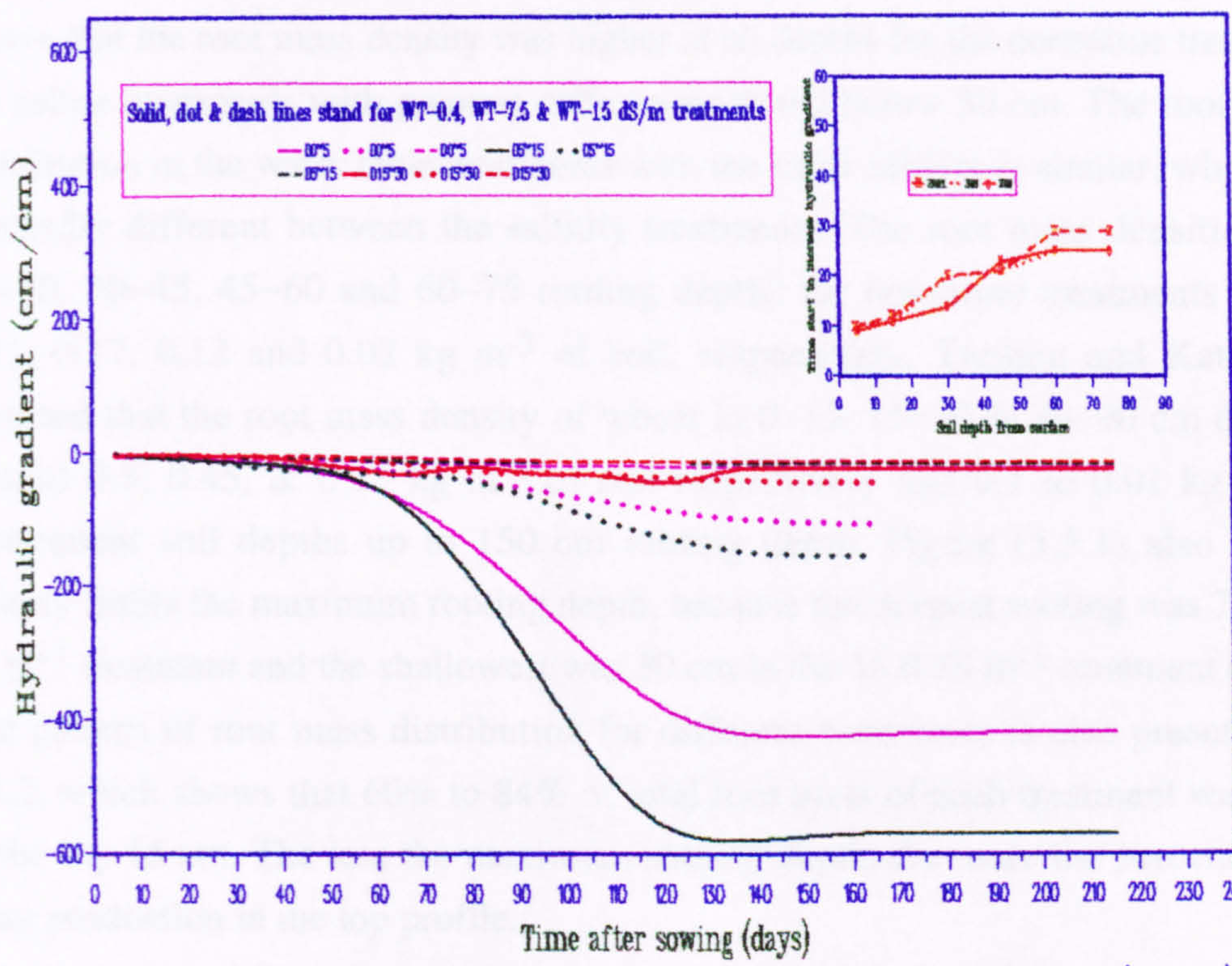


Fig. 5.4.3(d): Comparison of hydraulic gradient for diff. salinities (rye'93)

higher than from the nonsaline water table even though the hydraulic gradient was greater [see Fig. 5.4.3(d)]. This probably reflection of higher hydraulic conductivity associated higher moisture contents [Fig. 5.1.3(d)]. Again, during the period (150 days onwards) of hydraulic gradient, the rate of capillary rise from the nonsaline water table was higher. The reason for the higher rate may be because of the rooting depth in the nonsaline water table extended near to the water table. Prathapar and Meyer (1992) indicate that, plants can abstract around 20 % of total water table contribution directly from the water table, if water table remains close to the rooting depth. The overall results may indicate that, the osmotic potential might have some effect on capillary rise, but it needs further detail investigation with more replications. The trend of total potential versus groundwater contribution is also shown in Fig. 5.4.7.

5.5 Water uptake and root functions

Fig. 5.5.1 shows the density of root mass at different depths for different ryegrass treatments. It shows that the root mass density was much greater in the top 15 cm for all treatments and then decreased exponentially in the lower rooting zone. It also shows that the root mass density was higher at all depths for the nonsaline treatment than the saline treatments with greatest differences found below 30 cm. The rooting density distribution in the water table treatments with the same salinity is similar, whereas it was markedly different between the salinity treatments. The root mass densities in 0~15, 15~30, 30~45, 45~60 and 60~75 rooting depths for nonsaline treatments were 0.71, 0.21, 0.17. 0.12 and 0.02 kg m^{-3} of soil, respectively. Tardieu and Katerji (1990) reported that the root mass density of wheat in 0~15, 15~30 & 30~90 cm depths were around 0.5, 0.45, & 0.25 kg m^{-3} of soil respectively and 0.1 to 0.01 kg m^{-3} in the subsequent soil depths up to 150 cm rooting depth. Figure (5.5.1) also shows that salinity limits the maximum rooting depth, because the deepest rooting was 75 cm in 0.4 dS m^{-1} treatment and the shallowest was 30 cm in the 15.0 dS m^{-1} treatment of ryegrass. The pattern of root mass distribution for different treatments is also presented in Fig. 5.5.2, which shows that 60% to 84% of total root mass of each treatment was produced in the top 15 cm. The less the maximum rooting depth, the more the percentage of root mass production in the top profile.

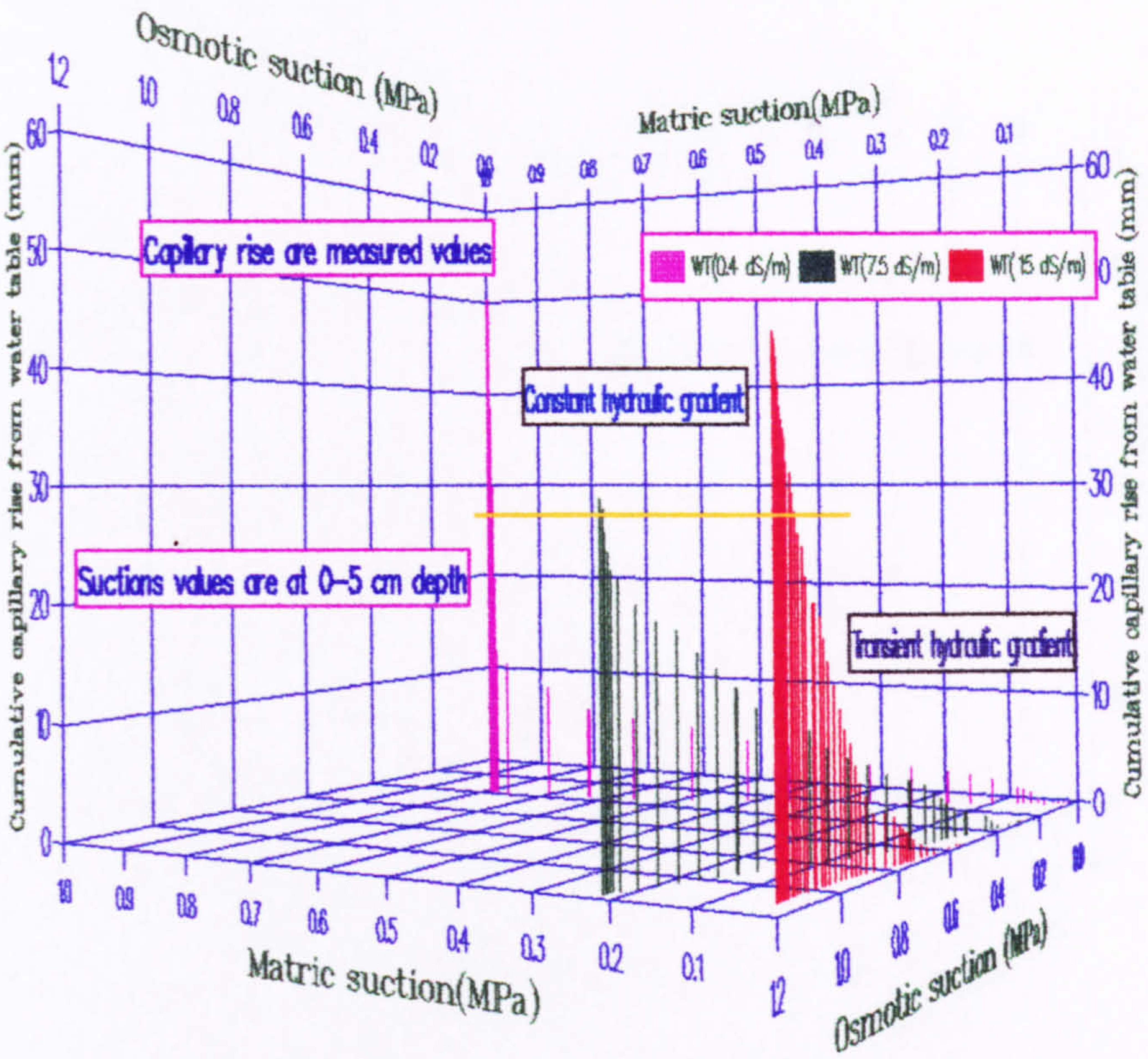


Fig. 5.4.4: Capillary rise versus osmotic & matric suction (ryegrass'93)

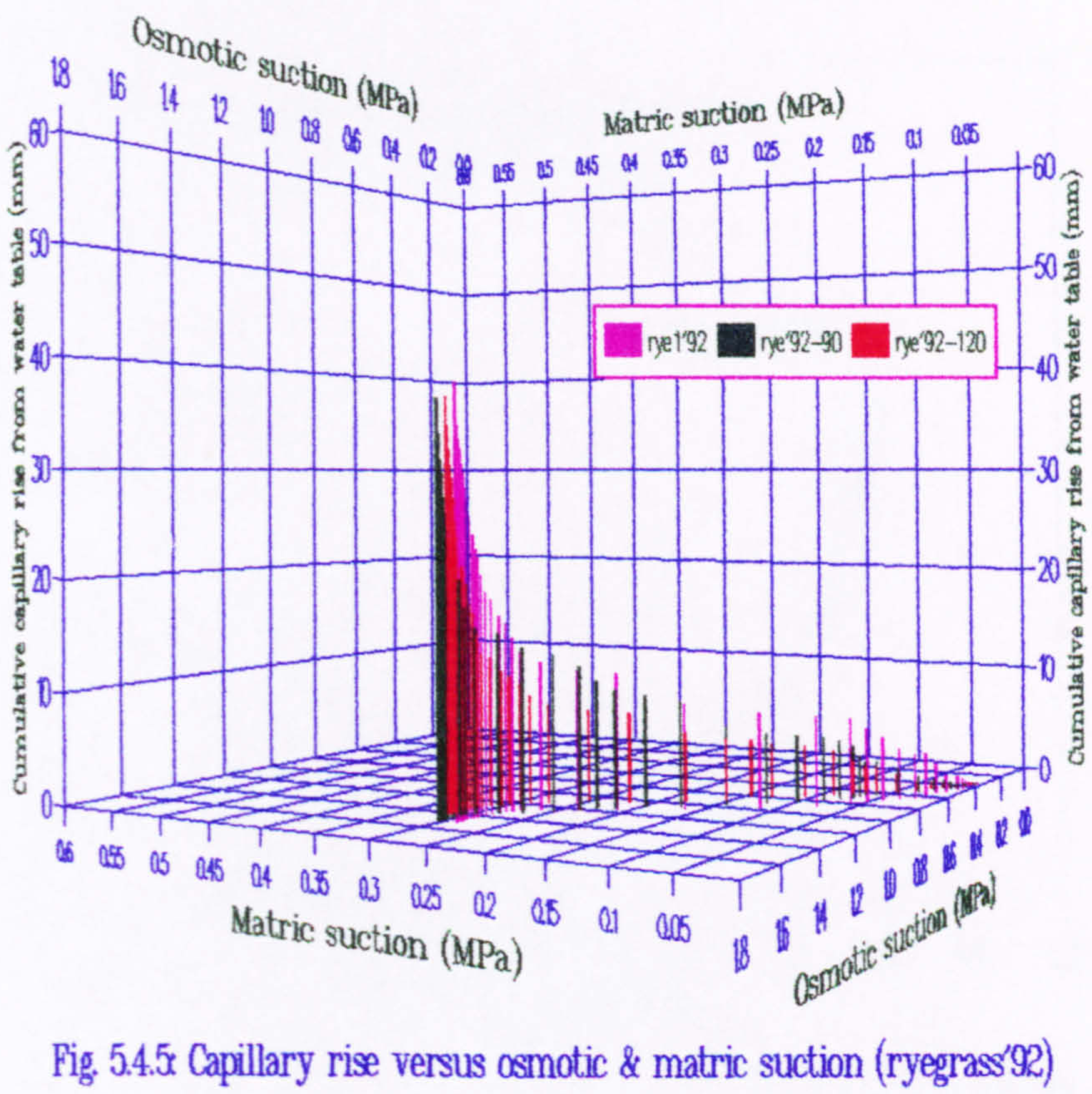


Fig. 5.4.5: Capillary rise versus osmotic & matric suction (ryegrass'92)

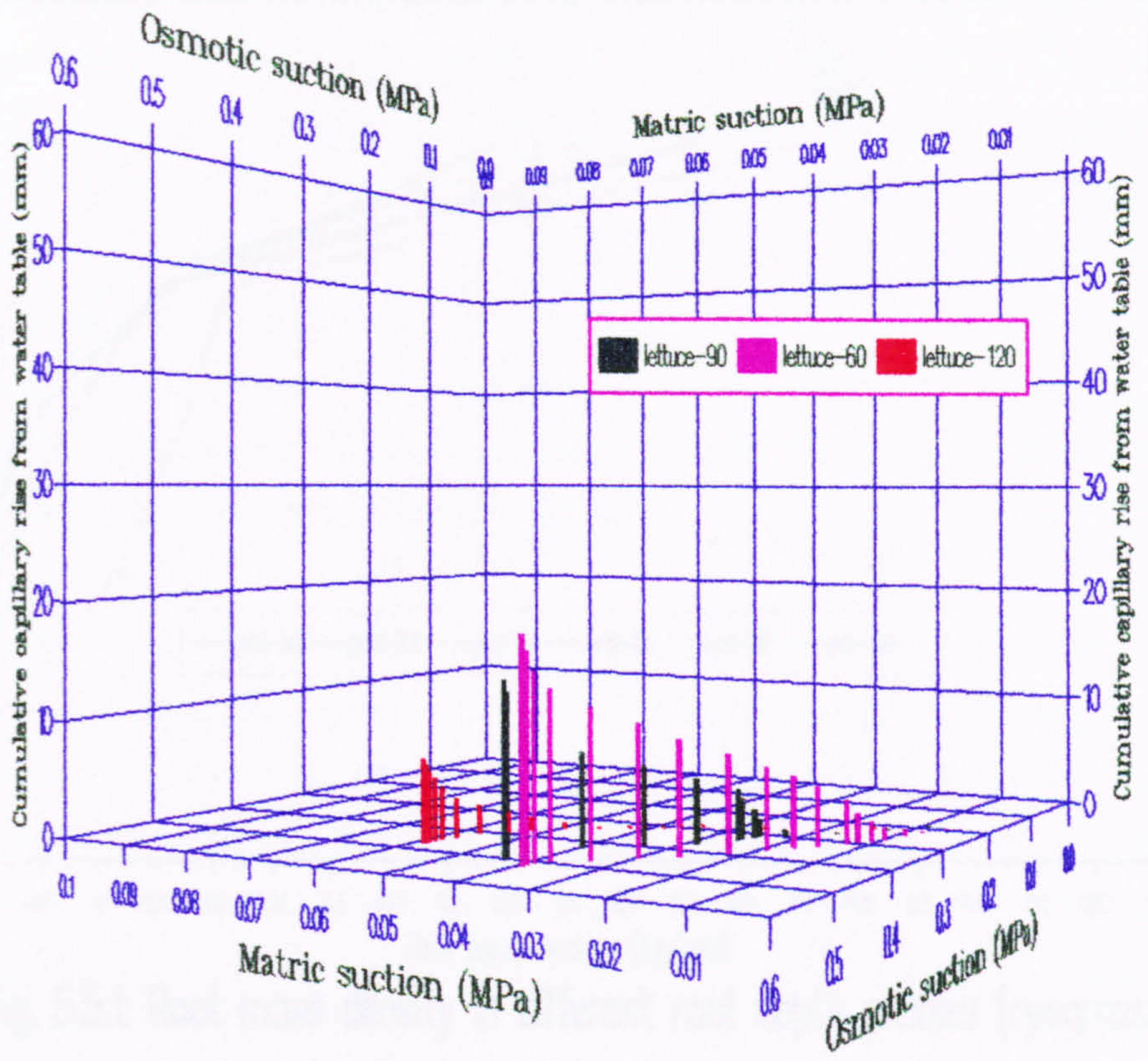


Fig. 5.4.6: Capillary rise versus osmotic & matric suction (lettuce)

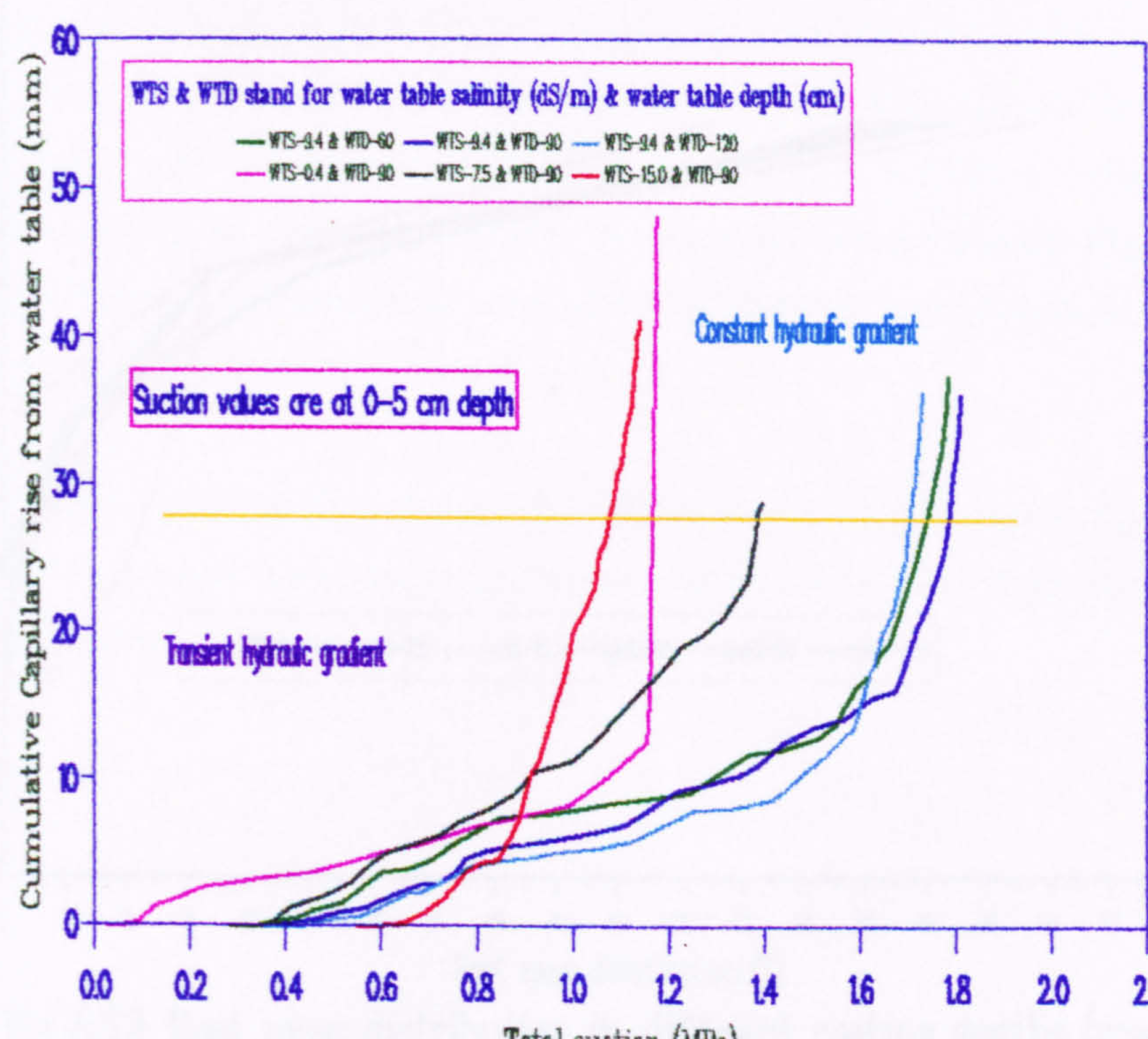


Fig. 5.4.7: Capillary rise from water table versus total suction (ryegrass)

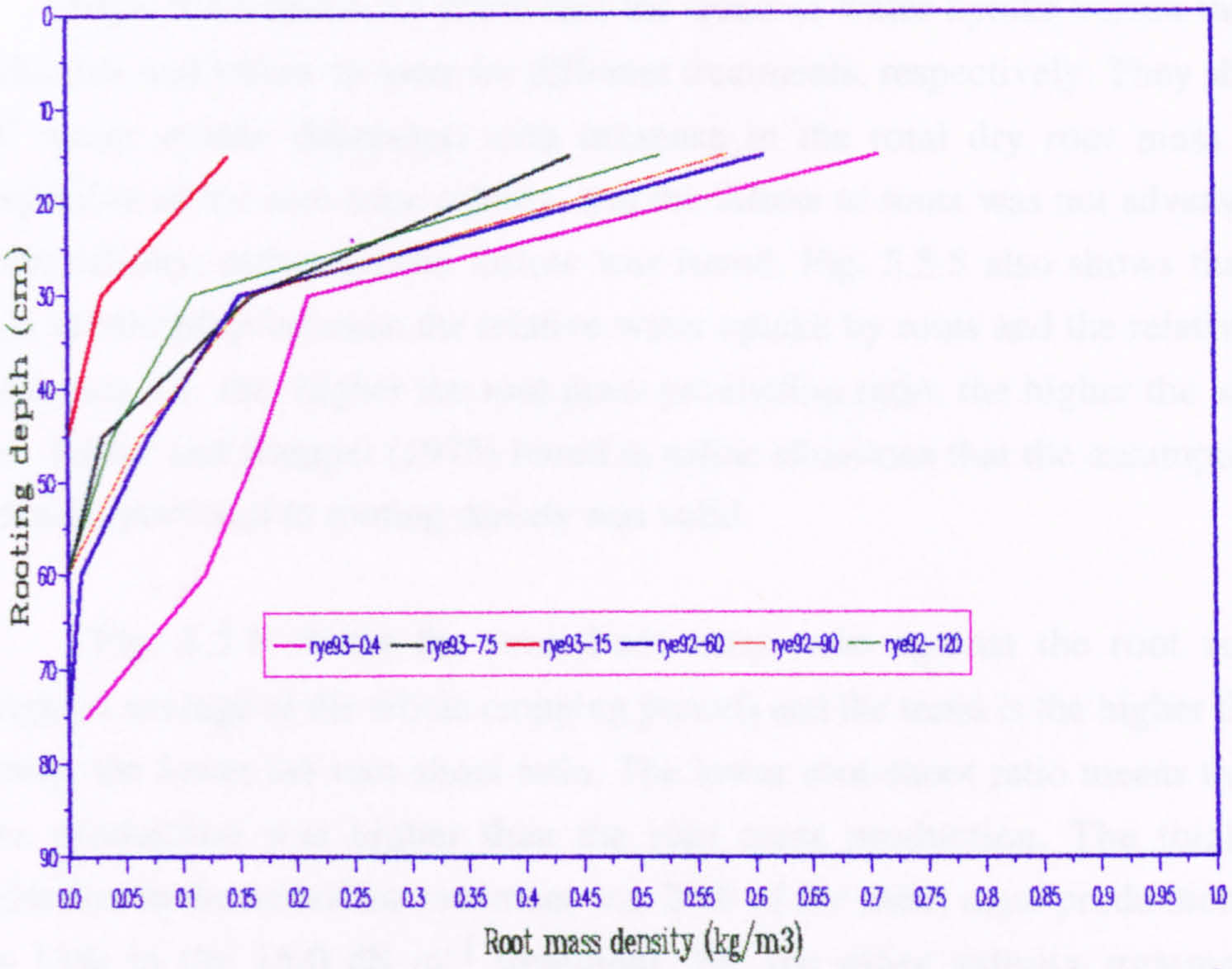


Fig. 5.5.1: Root mass density in different root depth profiles (ryegrass)

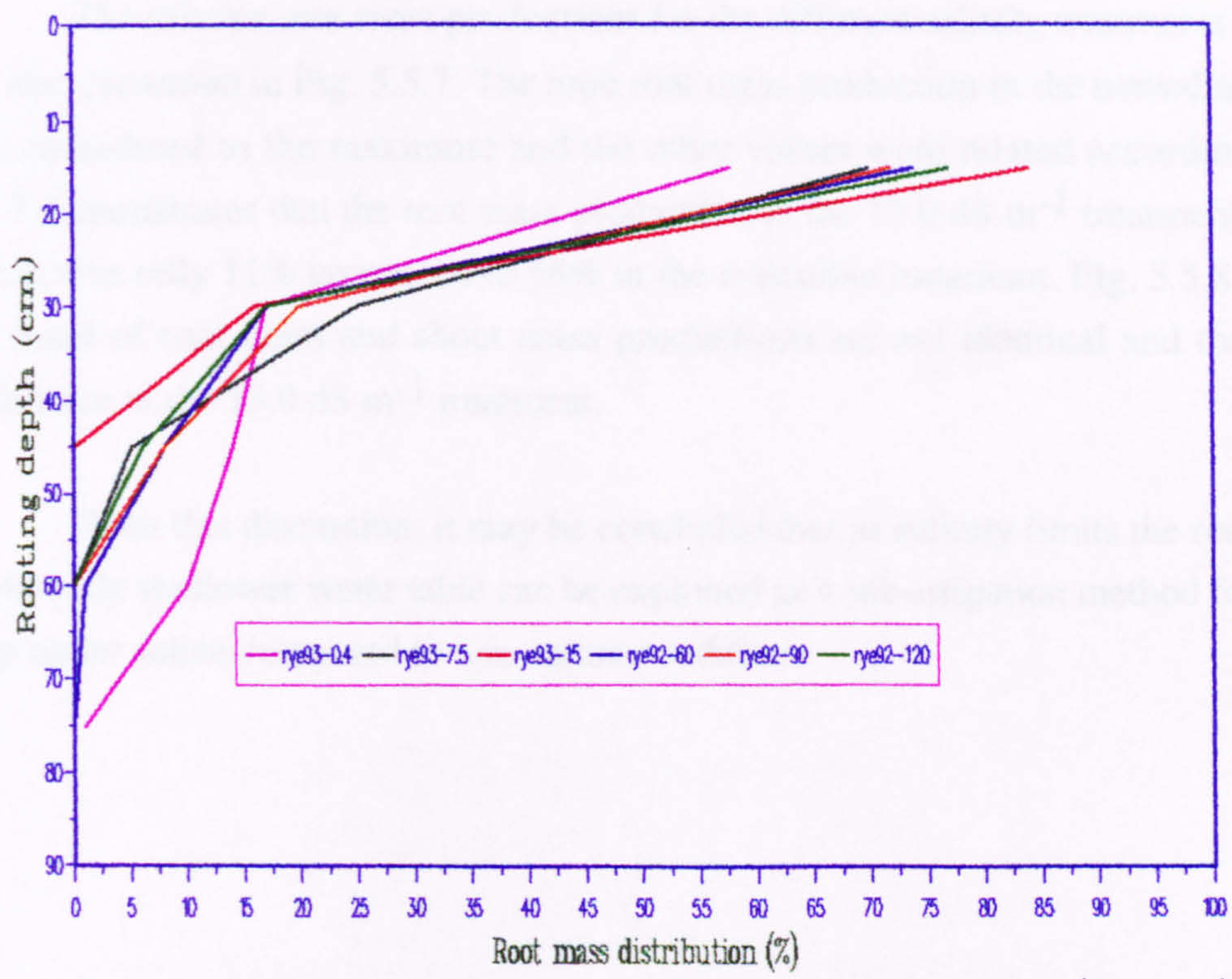


Fig. 5.5.2: Root mass distribution in different rooting depths (ryegrass)

Figs. 5.5.3 and 5.5.4 show also the trend of water uptake versus the root mass production and inflow to roots for different treatments, respectively. They show that the total water uptake decreased with decrease in the total dry root mass production irrespective of the root zone salinity, and the inflow to roots was not adversely affected by the salinity; rather greater inflow was found. Fig. 5.5.5 also shows that there is a linear relationship between the relative water uptake by roots and the relative root mass production, i.e. the higher the root mass production ratio, the higher the water uptake ratio. Tayler and Klepper (1975) found in saline situations that the assumption of water uptake proportional to rooting density was valid.

Fig. 5.5.6 shows the root-shoot mass ratio against the root zone salinity (weighted average of the whole cropping period) and the trend is the higher the root zone salinity, the lower the root-shoot ratio. The lower root-shoot ratio means that the shoot mass production was higher than the root mass production. The total root mass production in the nonsaline treatment was 20% of the shoot mass production, whereas it was 13% in the 15.0 dS m^{-1} treatment; for the other salinity treatments, it was intermediate.

The relative root mass productions for the different salinity treatments of ryegrass are also presented in Fig. 5.5.7. The total root mass production in the nonsaline treatment was considered as the maximum and the other values were related accordingly. Figure 5.5.7 demonstrates that the root mass production in the 15.0 dS m^{-1} treatment within top 15 cm was only 11% compared to 58% in the nonsaline treatment. Fig. 5.5.8 shows that the trend of root mass and shoot mass productions are not identical and there is a big difference in the 15.0 dS m^{-1} treatment.

From this discussion, it may be concluded that as salinity limits the rooting depth, a relatively shallower water table can be exploited as a sub-irrigation method for a specific crop under saline compared to non-saline condition.

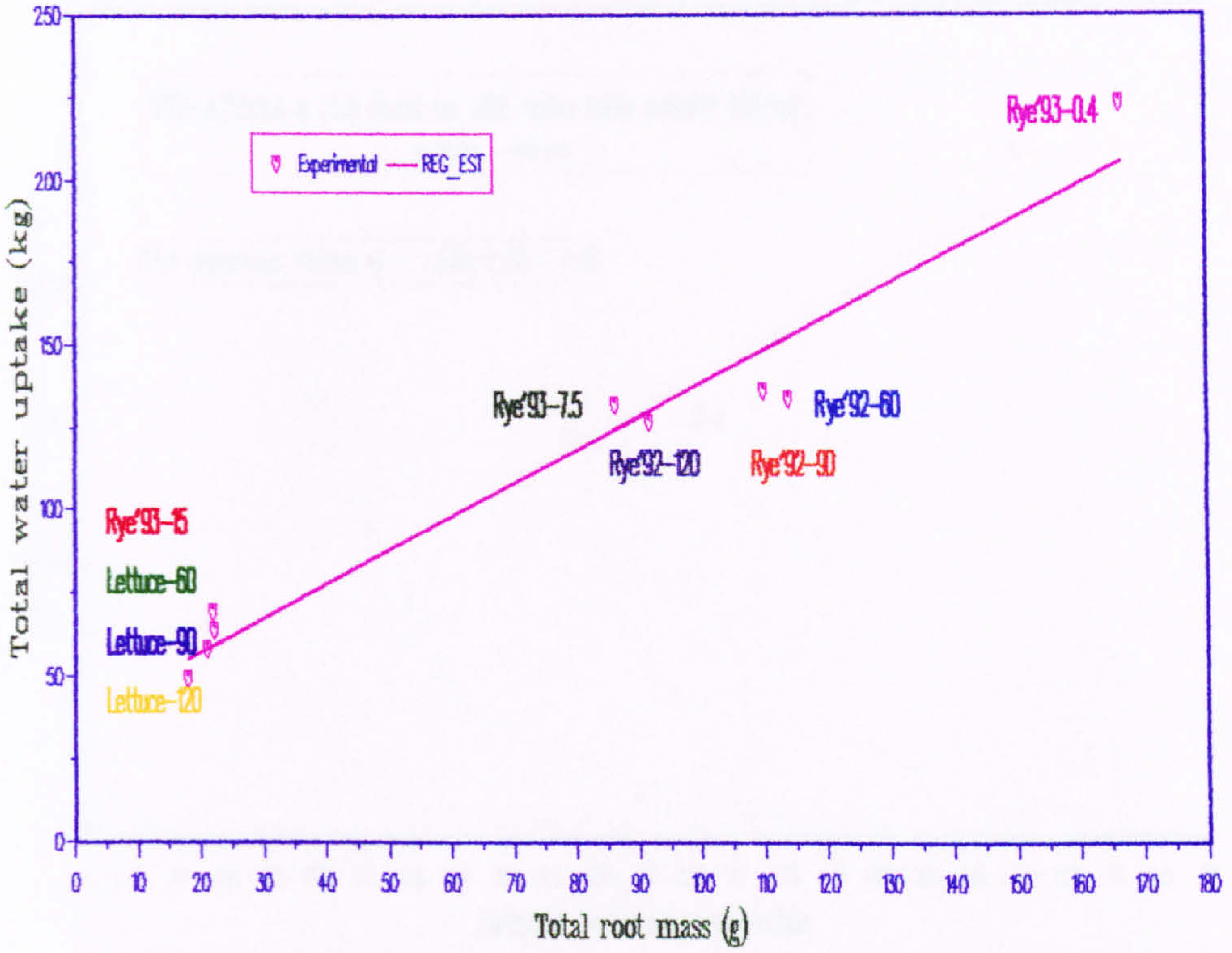


Fig. 5.5.3: Water uptake versus root mass production (ryegrass & lettuce)

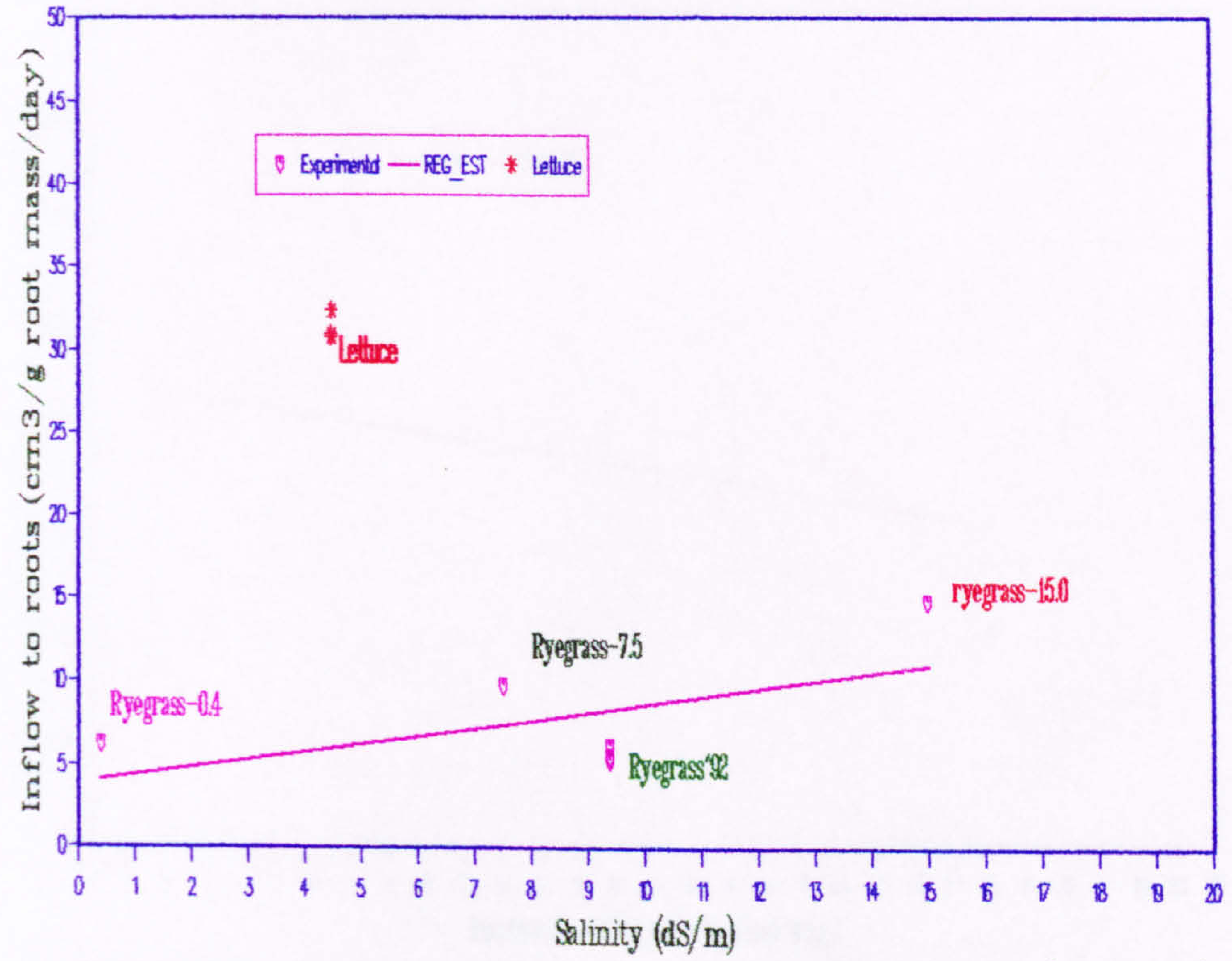


Fig. 5.5.4: Inflow to roots for diff. salinity treatments (ryegrass)

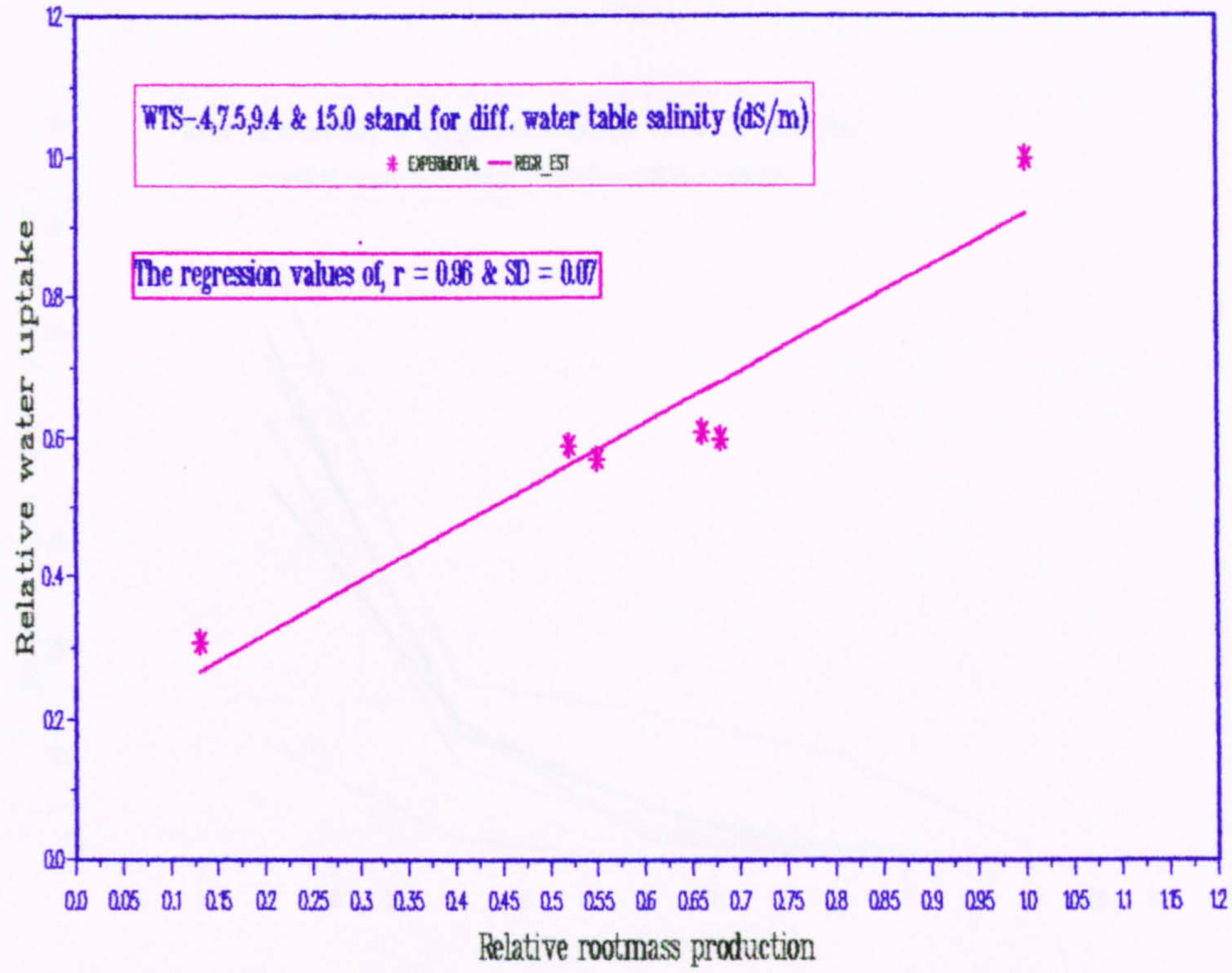


Fig. 5.5.5: Relative water uptake versus root mass production (ryegrass)

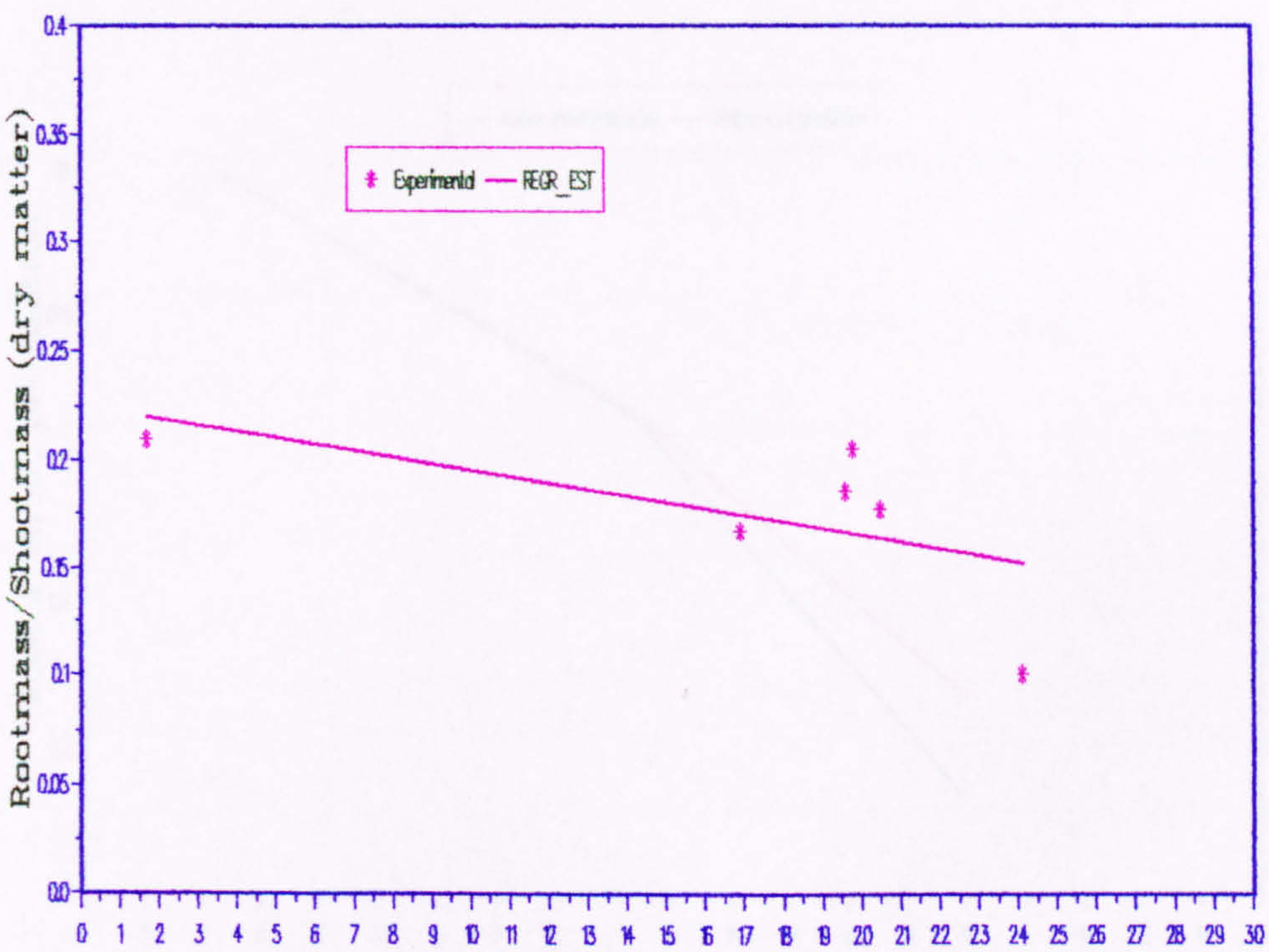


Fig. 5.5.6: Rootmass & shootmass production ratio Vs rootzone salinity

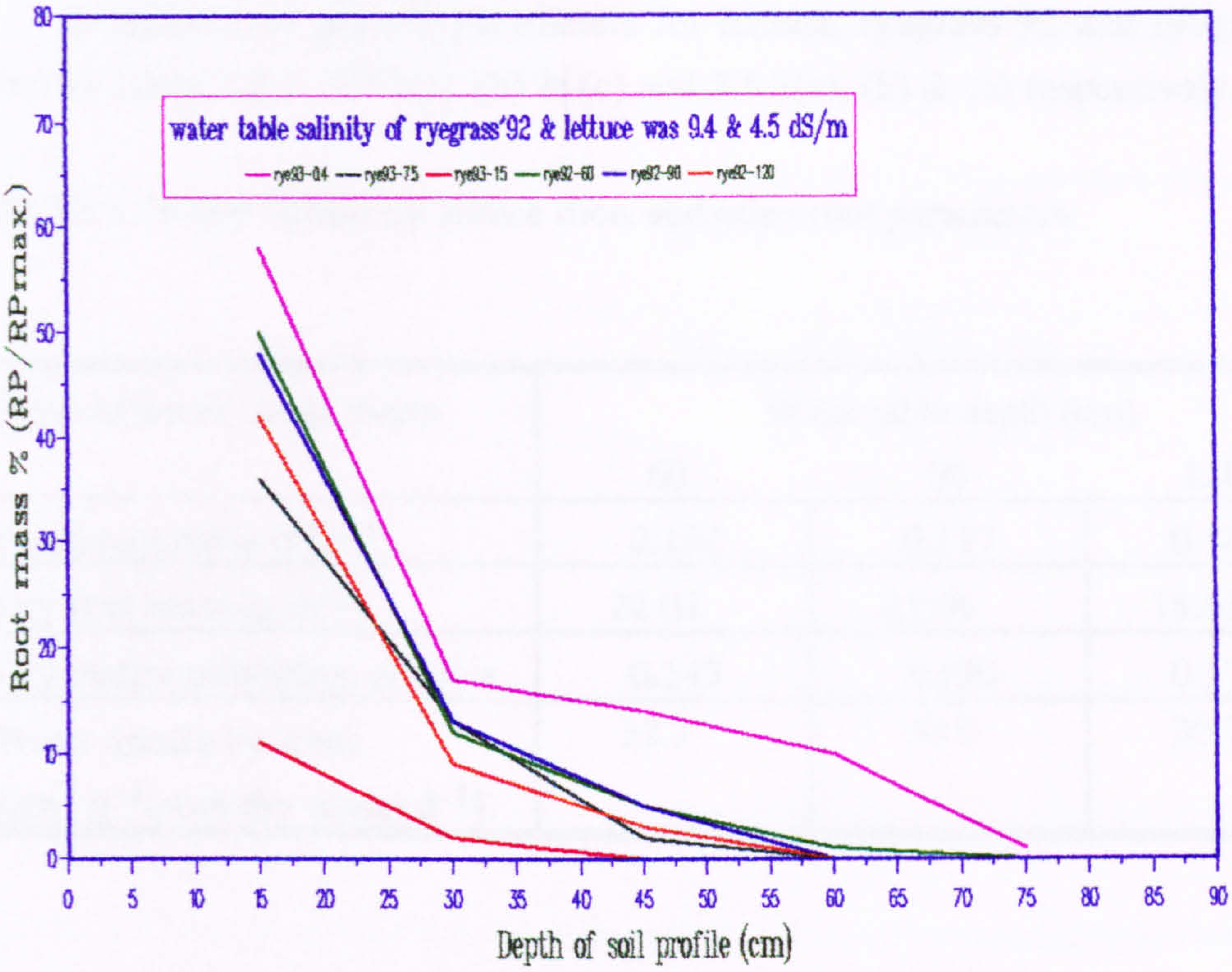


Fig. 5.5.7: Percentage of root mass production in different soil depths

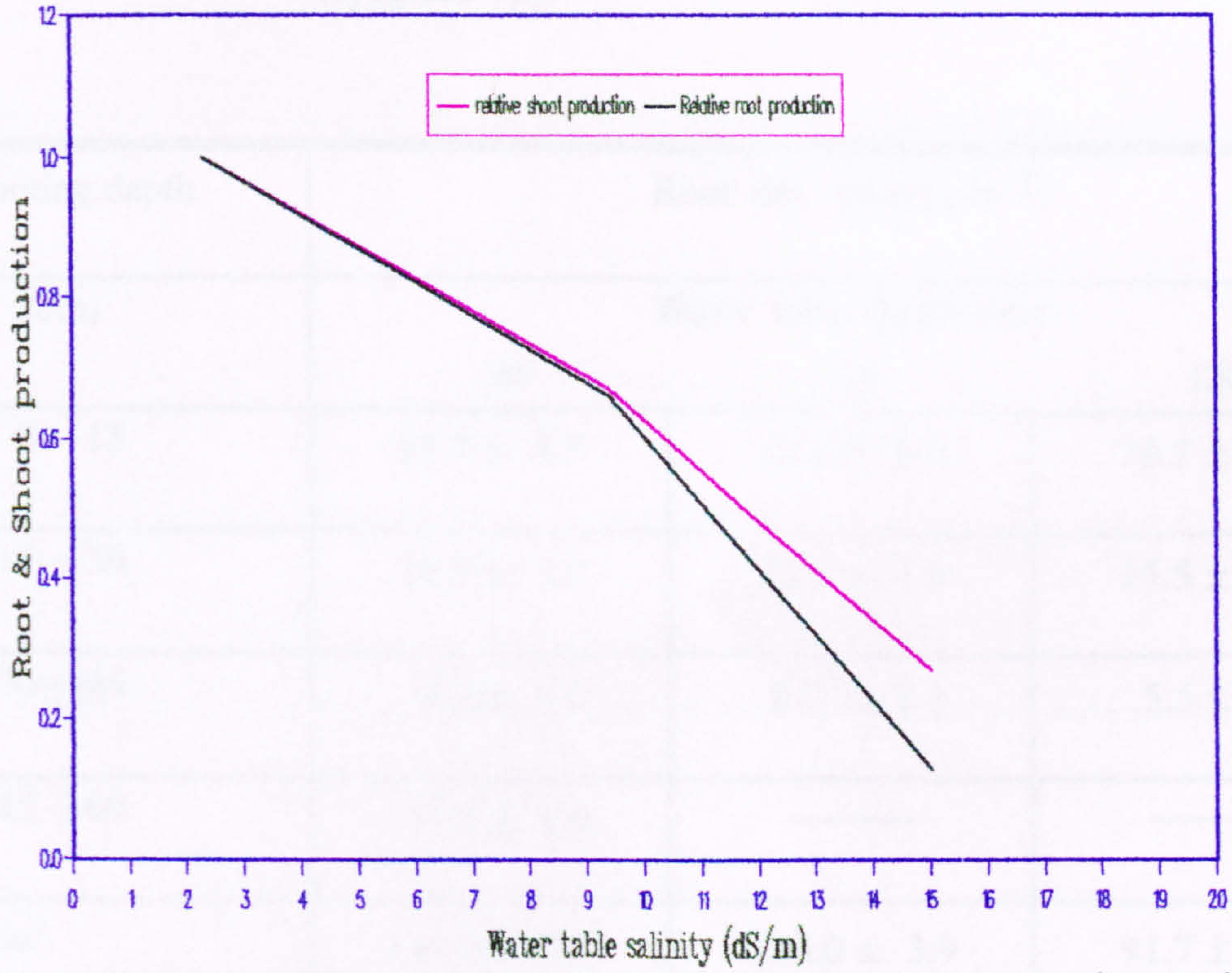


Fig. 5.5.8: Comparison of root & shoot production trend Vs salinity (ryegrass)

Detailed root growth parameters for lettuce, ryegrass'92 and ryegrass'93 are shown in Table 5.5.1, 5.5.2(a), (b) & (c) and 5.5.3(a), (b) & (c) respectively.

Table 5.5.1: Water uptake by lettuce roots and other root parameters.

Description of components	Water table depth (cm)		
	60	90	120
Fresh root mass (t ha^{-1})	0.152	0.117	0.103
Dry root mass (g m^{-2})	22.04	21.06	18.03
Dry matter proportion of roots	0.145	0.180	0.175
Water uptake by roots [cm^3 g^{-1}(root dry mass) d^{-1}].	32.3	30.9	30.7

Table 5.5.2(a): Root distribution at different depth of root zone above the different water table depth with equal groundwater salinity treatments (ryegrass'92).

Rooting depth	Root dry mass (g m^{-2})		
(cm)	Water table depth (cm)		
	60	90	120
0—15	83.7 ± 5.5	79.2 ± 8.0	70.7 ± 6.2
15—30	19.8 ± 7.2	22.1 ± 1.5	15.5 ± 0.4
30—45	9.0 ± 1.0	8.7 ± 1.1	5.5 ± 0.9
45—60	*1.5 ± 2.6	———	———
Total	114.0 ± 2.8	110.0 ± 3.9	91.7 ± 3.2

* Root mass existed in one observation out of three.

Table 5.5.2(b): Percentage of root mass production at different depth of root zone (ryegrass'92).

Rooting depth (cm)	Root dry mass (%)		
	Water table depth (cm)		
	60	90	120
0--15	74	72	77
15--30	17	20	17
30--45	8	8	6
45--60	1	———	———

Table 5.5.2(c): Root parameters for different water table treatments (ryegrass'92).

Parameters	Water table depth (cm)		
	60	90	120
Rooting depth (cm)	45.0	45.0	45.0
Total root dry mass ($g\ m^{-2}$)	114.0	110.0	91.7
Water uptake by roots [$cm^3\ g^{-1}$(root dry mass) d^{-1}]	5.3	5.6	6.2

Table 5.5.3(a): Root distribution at different depth of root zone above the different salinity treatments with equal water table depth (ryegrass'93).

Rooting depth	Root dry mass (g m^{-2})		
(cm)	Salinity level(dS m^{-1})		
	0.4	7.5	15.0
0--15	97.5 ± 9.6	60.7 ± 12.4	18.5± 2.1
15--30	28.2 ± 5.5	21.4 ± 7.8	3.5± 0.6
30--45	23.0± 3.8	4.1 ± 2.2	———
45--60	15.9 ± 3.0	———	———
60--75	*2.1 ± 1.9	———	———
Total	166.7± 3.0	86.2± 5.1	22.0± 1.1

* Root mass existed in two observations out of three.

Table 5.5.3(b): Percentage of root mass production at different depth of root zone (ryegrass'93).

Rooting depth	Root dry mass (%)		
(cm)	Salinity level (dS m^{-1})		
	0.4	7.5	15.0
0--15	58	70	84
15--30	17	25	16
30--45	14	5	———
45--60	10	———	———
60--75	1	———	———

Table 5.5.3(c): Different root parameters for different water table treatments (ryegrass'93).

Parameters	Salinity level (dS m^{-1}) 0.4	7.5	15.0
Rooting depth (cm)	75.0	45.0	30.0
Total root dry mass (g m^{-2})	166.7	86.2	22.0
Water uptake by roots [cm^3 g^{-1}(root dry mass) d^{-1}]	6.2	9.8	14.8

Table 5.5.3(c) shows that the higher the salinity of the soil water, the lower the root dry mass, and the smaller the maximum rooting depth [Table 5.5.3(b)]. Table 5.5.3(c) shows that root water uptake rate (mm of water per g dry root mass) was higher in the saline soils. The greater difference in root mass production is obviously due to salinity, because the matric stress was much higher in the nonsaline treatment. Similar trends were found in the different water table treatments (where the salinity was same), but the difference due to water table depth was much smaller [Tables 5.5.2(a) & (c)].

Presumably, the prediction of time-dependent root elongation and the maximum rooting depth for a cropping period is very important in proper irrigation scheduling and is more important in irrigation with saline water, but the prediction of such a dynamic parameter(s) is difficult to achieve. Though actual measurements can be done by root sampling in the field, it is very often laborious and destructive. Thus the present information on the behaviour of roots might be useful in the modelling of root growth and water uptake, especially the reduction factor due to salinity.

5.6 Water-salinity production functions

The functional relationship between crop yield and water use is called the water-production function. Water use may reasonably be represented by: the depth of irrigation, the total field water supply, the consumptive use of water estimated as evapotranspiration (or, more precisely transpiration). However, relating yield directly to irrigation is not always satisfactory. Under saline conditions, the effect of salinity on the crop yield and actual evapotranspiration (ET_a) is implied by its effect on decreasing ET_a as salinity increases.

De Wit (1958) developed a relation between transpiration and yield for climates with large percentage of bright sunshine duration as

$$Y = m\frac{T}{E_p} \qquad \text{....................(5.6.1)}$$

where,

Y	dry matter yield,
T	actual transpiration,
E_p	potential (free water) evaporation, and
m	crop specific proportionality coefficient.

However, Hank (1974) mentioned that, when only the effects of limited water application were of interest, the model of DeWit (1958) could be reduced to:

$$\frac{DM}{DM_m} = \frac{T}{T_m} \qquad \text{..........(5.6.2)}$$

where,

DM	dry matter yield,
DM_m	maximum or potential dry matter yield,
T	crop transpiration, and
T_m	maximum or potential transpiration when soil does not limit yield.

5.6.1 Yield-evapotranspiration relationships

In this investigation, actual evapotranspiration (ET_a) is measured as the sum of soil moisture depletion (ΔW) and capillary rise (C) from the water table. The yield corresponding to evapotranspiration of ryegrass'93 at 0.4 dS m^{-1} treatment was considered as the potential for calculating the relative yield and evapotranspiration for the salinity treatments.

Figs. 5.6.1 shows that salinity proportionately affected the evapotranspiration, but not the water use efficiency, i.e unit dry matter yield per unit of water use. Rather the salinity treatments showed better water use efficiency. The results of lettuce have been presented here with respect to the same (ryegrass at 0.4 dS m^{-1}) which showed that more water was evapotranspired per unit of drymatter yield than ryegrass. The reason may be that evaporation was higher because of smaller crop cver (32 plants for lettce and 432 to 1067 for ryegrass, see Table 4.2), especially in the early stages of growth. However, the economic importance with respect to water use of lettuce can be better judged by its fresh yield, not the drymatter yield.

The detailed relative yield and relative evapotranspiration for different cuts of ryegrass'93 and ryegrass'92 are presented in Fig. 5.6.2 & 5.6.3. Both the yield and evapotranspiration proportions were calculated on the basis of the total yield and evapotranspiration of the respective treatments. Fig. 5.6.2 shows that, for 1st cut, the proportion of water use is higher than the proportion of drymatter yield because of the more soil evaporation because of less crop cover in the early growth stages. From the second cut and onwards, the drymatter proportions are higher than the proportions of the water use, and are greater in the higher salinity treatments. Fig. 5.6.2 also shows that after 170 days, the drymatter proportions somewhat slowed down in the non-saline treatment. The reason may be because of inadequate aeration in the active root zone as the rooting depth touched the water table. In addition, the seasonal effect of Autumn (the months of October and November) may have some effect in all experiments. The ryegrass'92 results (Fig. 5.6.3) showed that the 60 cm water table treatment had higher proportions drymatter up to the 3rd cut when salinity in the root zone was higher (see Fig. 5.3.3a, b & c in chapter 5). For the following cuts, the drymatter proportions became smaller than for the deeper water tables, while roots extracted water from the water table & partly from lower soil depth upto 150 days where salinity was much smaller. On the other hand, in the deeper water table treatment's root zone (45 cm for both) was far above the water tables where salinity was higher compared to the 60 cm water table salinity in the latter periods. The overall reason for more drymatter yield may be because of more aeration in the root zone due to the soil dryness. The reason for higher drymatter proportions in the salinity treatments than nonsaline treatment, I do not know. However, this is a positive side of the present management of saline water.

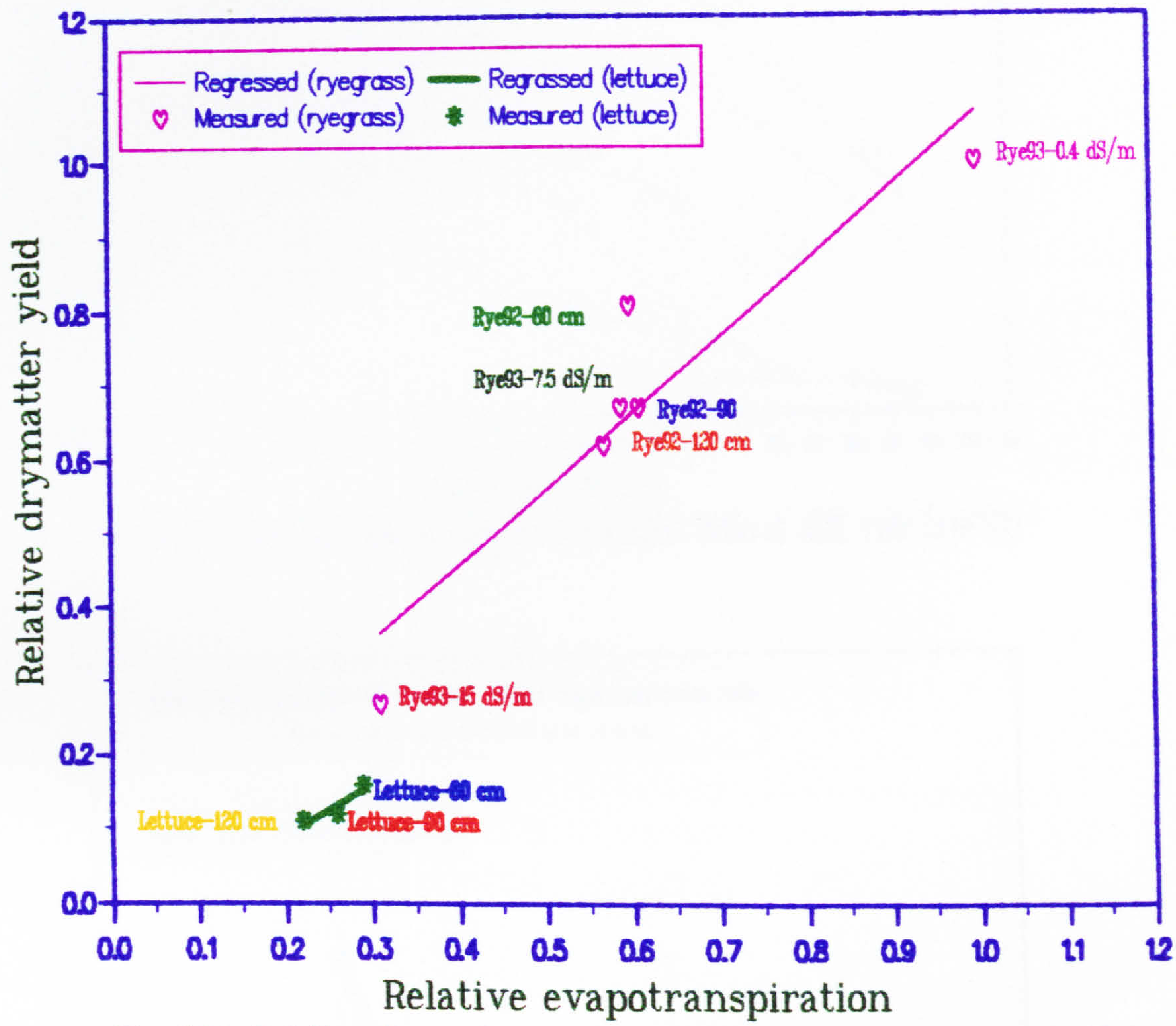


Fig. 5.6.1: Relative drymatter yield versus relative evapotranspiration.

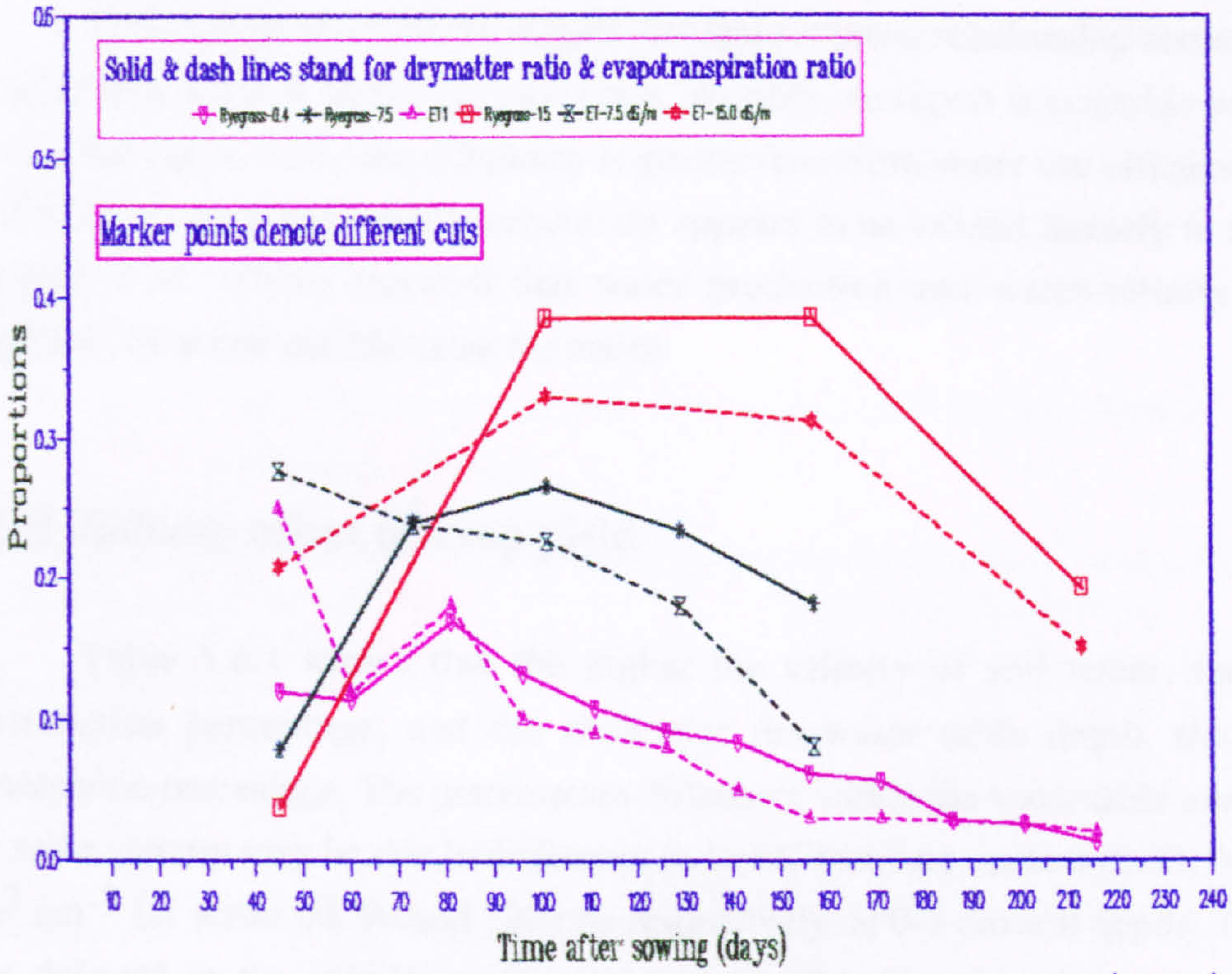


Fig. 5.6.2: Drymatter ratio & evapotranspiration ratio at diff. cuts (rye'93)

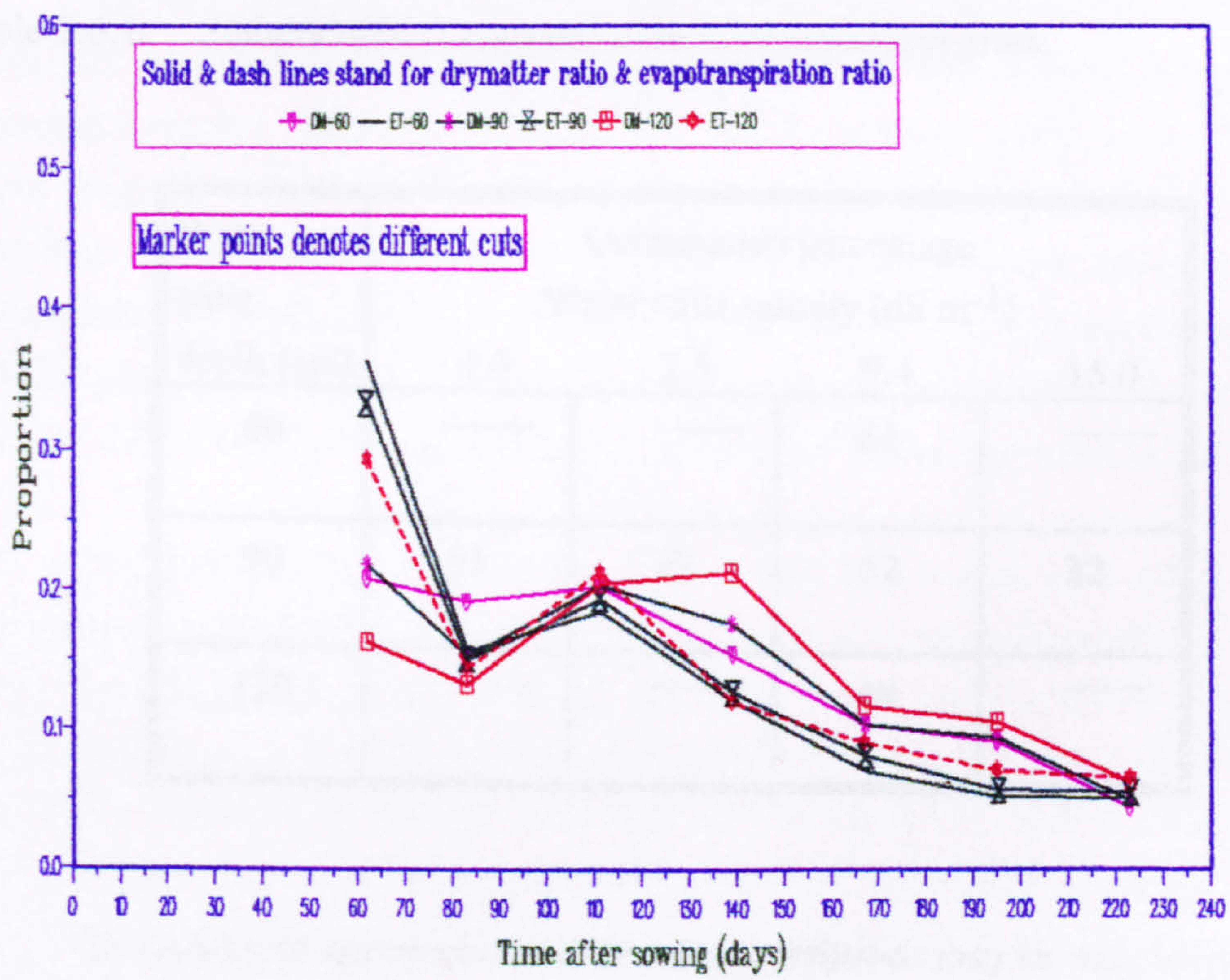


Fig. 5.6.3: Drymatter ratio & evapotranspiration ratio at diff. cuts (rye'92)

The experimental results suggest that there is linear relationship between yield and evapotranspiration in nonsaline conditions. Possibly, no report is available yet where it is shown that saline water use efficiency is greater than fresh water use efficiency. Hoffman (1985) concluded that evapotranspiration appears to be related linearly to soil salinity. Stewart et al. (1980) reported that water production and water-salinity production functions were one and the same for maize.

5.6.2 Salinity effect on crop yield

Table 5.6.1 shows that the higher the salinity of soil water, the lower the germination percentage, and the shallower the water table depth, the higher the germination percentage. The germination difference within the water table treatments with the same salinity may be due to difference in initial moisture content (0.38, 0.36 and 0.34 cm^3 cm^{-3} for water 60, 90 and 120 cm, respectively, at 0-5 cm soil depth). Germination was delayed in the salinity treatments compared to the nonsaline treatment with a difference of 5 days at 15.0 dS m^{-1}.

Table 5.6.1: Salinity effect on germination of perennial ryegrass.

Water table depth (cm)	Germination percentage Water table salinity (dS m^{-1})			
	0.4	7.5	9.4	15.0
60	——	——	61	——
90	91	79	52	32
120	——	——	49	——

The density of germination under saline conditions may be increased by heavier seeding than for the nonsaline condition. FAO/UNESCO (1973) reported that adverse effect of soil salinity on germination can be minimized by heavier seeding. If germination is reduced by 50% at the salt level existing in a soil, a full stand can be obtained by using a seeding rate of twice the amount required on a nonsaline soil. They also mentioned that high seeding rate also aid in reducing the harmful effect of the delay in germination and emergence on crop stand. They added that in general, crops are more tolerant in cool climate than in a hot one.

Detailed fresh top yield, dry matter yield and water use efficiency are presented in Tables 5.6.2, 5.6.3 and 5.6.4 for lettuce, ryegrass'92 and ryegrass'93 respectively.

Table 5.6.2 : The yield components and water use efficiencies for different water table treatments (lettuce).

Description of components	Water table depth (cm)		
	60	90	120
Fresh top yield (t ha^{-1})	2.370	1.796	1.614
Dry top yield (g m^{-2})	0.126	0.095	0.090
Dry matter proportion of tops	0.053	0.053	0.056
Water use efficiency (g / kg of water).	1.97	1.62	1.81

The total yield of drymatter from nonsaline treatment under 90 cm water table depth was 7.95 t ha^{-1} (Table 5.6.4), which may be considered as usual potential yield of perennial ryegrass. Bailey (1990) reported that the average yield, over 13 years data under irrigated and non-irrigated conditions were 9.7 and 7.7 t ha^{-1} in the United Kingdom. The growing period of his reported yields was 8 months, whereas the present experimental growing period is 7 months. Munro et al. (1992) reported that the annual herbage production (dry matter yield) of perennial ryegrass in UK, totalling of 9 cuts with 3-4 weeks cutting intervals, was around 10 t ha^{-1} under nonsaline conditions.

The total drymatter yield for ryegrass'93 at 7.5 and 15.0 dS m^{-1} salinity treatments were 67%, 27% respectively from 0.4 dS m^{-1} treatment at 90 cm water table depth. The ryegrass'92 at 60, 90 & 120 cm water tables yielded 80, 67 & 62% (Table 5.6.3) from the same 0.4 dS m^{-1}. Note that, significant yield differences are found in the ryegrass'92 experiments though total water use by them was almost same (see Table 5.2.2 in section 5.2). The reason may be that the soil temperature in 60 cm water table lysimeter was less than the others, because it was surrounded by the other two lysimeters as well as neighbouring shades while the other were exposed to sunlight (See Plate 5.6.1). The precaution (wrapping with kitchen foils) for preventing sunshine might not be good enough. The yield potential at 60 cm water table depth indicate that this shallow saline water table is preferable for ryegrass under saline situations.

Description of components		Fresh yield (t/ha)			Drymatter yield (t/ha)		
		Water table depth (cm)			Water table depth (cm)		
Cut	Day	60	90	120	60	90	120
First	62	4.95	3.56	2.31	0.80	0.63	0.42
Second	83	4.62	2.40	1.83	0.74	0.44	0.34
Third	11	5.03	3.29	2.60	0.78	0.59	0.53
Fourth	139	3.69	2.69	2.66	0.59	0.51	0.55
Fifth	167	2.58	1.81	1.75	0.40	0.30	0.30
Sixth	195	2.23	1.57	1.57	0.35	0.27	0.27
Seventh	223	1.08	0.75	0.84	0.17	0.14	0.16
Total		24.18	16.07	13.56	3.83	2.88	2.57
Surface		5.82	5.60	5.60	2.57	2.47	2.37
G. total		30.00	21.67	19.16	6.40	5.35	4.94

Description of components		Drymatter proportion			Water use efficiency (g/kg of water)		
		Water table depth (cm)			Water table depth (cm)		
Cut	Day	60	90	120	60	90	120
First	62	0.162	0.177	0.182	1.64	1.38	1.12
Second	83	0.160	0.183	0.186	3.51	2.14	1.86
Third	11	0.155	0.179	0.204	3.13	2.21	1.94
Fourth	139	0.160	0.190	0.207	3.64	2.95	3.63
Fifth	167	0.155	0.166	0.171	4.22	2.67	2.60
Sixth	195	0.157	0.172	0.172	5.01	3.48	3.00
Seventh	223	0.157	0.187	0.190	2.51	1.87	1.93
Average		0.158	0.179	0.187	2.86	2.11	2.02
Surface		0.442	0.441	0.423	—	—	—
Overall		0.213	0.247	0.258	4.77	3.92	3.89

Table 5.6.3: The yield components and water use efficiency for different water table treatments (ryegrass' 92)

Description of components		Fresh yield (t/ha)			Drymatter yield (t/ha)		
		Water table salinity (dS/m)			Water table salinity (dS/m)		
Cut	Day	0.4	7.5	15	0.4	7.5	15
C1(All)	45	5.05	1.92	0.22	0.63	0.27	0.04
C2S1	60	4.45	------	------	0.62	------	------
C2S2	73	------	4.65	------	------	0.82	------
C3S1	81	6.37	------	------	0.92	------	------
C4S1	96	4.89	------	------	0.72	------	------
C3S2/C2S	101	------	4.84	1.65	------	0.91	0.40
C5S1	111	3.90	------	------	0.59	------	------
C6S1	126	3.23	------	------	0.50	------	------
C4S2	129	------	3.96	------	------	0.80	------
C7S1	141	2.97	------	------	0.46	------	------
C8S1	156	2.13	------	------	0.34	------	------
C5S2/C3S	157	------	2.98	1.54	------	0.62	0.40
C9S1	171	1.92	------	------	0.31	------	------
C10S1	186	0.99	------	------	0.16	------	------
C11S1	201	0.84	------	------	0.14	------	------
C4S3	213	------	------	0.73	------	------	0.20
C12S1	216	0.43	------	------	0.07	------	------
Total		37.17	18.35	4.14	5.45	3.42	1.04
surface		6.60	4.56	2.46	2.50	1.94	1.13
G. total		43.77	22.91	6.60	7.95	5.36	2.17
* C & S stand for different crop cuts & salinities respectively.							

Table 5.6.4 : The yield components and water use efficiencies for different salinity treatments with equal water table depth (ryegrass' 93).

Description of components		Drymatter proportion			Water use efficiency (g/kg of water)		
		Water table salinity (dS/m			Water table salinity (dS/m)		
Cut	Day	0.4	7.5	15	0.4	7.5	15
C1(All)	45	0.125	0.141	0.182	1.135	0.735	0.276
C2S1	60	0.139	------	------	2.389	------	------
C2S2	73	------	0.176	------	------	2.606	------
C3S1	81	0.144	------	------	2.314	------	------
C4S1	96	0.147	------	------	3.256	------	------
C3S2/C2S	101	------	0.188	0.242	------	3.042	1.741
C5S1	111	0.151	------	------	3.054	------	------
C6S1	126	0.155	------	------	2.749	------	------
C4S2	129	------	0.202	------	------	3.378	------
C7S1	141	0.155	------	------	4.089	------	------
C8S1	156	0.160	------	------	4.503	------	------
C5S2/C3S	157	------	0.208	0.260	------	5.805	1.845
C9S1	171	0.161	------	------	4.133	------	------
C10S1	186	0.162	------	------	2.560	------	------
C11S1	201	0.161	------	------	2.348	------	------
C4S3	213	------	------	0.277	------	------	1.933
C12S1	216	0.160	------	------	1.533	------	------
Average		0.152	0.183	0.240	2.839	3.113	1.449
surface		0.379	0.425	0.459	------	------	------
Overall		0.182	0.234	0.329	3.557	4.047	3.120
* C & S stand for different crop cuts & salinities respectively.							

Table: 5.6.4 (continued)

Plate: 5.6.1(a)

60 cm / 0.4 dS m^{-1} 90 cm / 7.5 dS m^{-1} 120 cm/15 dS m^{-1}

Plate: 5.6.1(b)

Plate 5.6.1: The surroundings of the lysimeters in Moorbank glasshouse

The established plant population densities were smaller in the higher salinity treatments and might be somewhat minimized by heavier seeding during plantating. This is likely because, after germination, the plants in all the salinity treatments survived well and continued to grow for a long period of 223 or 216 days, except that growth was slower and there were insect-pest infestations. Note that the plants under 7.5 dS m^{-1} treatment were more heavily infested by pests than under the 15.0 dS m^{-1} treatment.

Fig. 5.6.4 shows the comparison of the experimental and estimated yield versus the water table salinity at 90 cm water table depth. Estimated yields were determined from the reported regression equation of Maas (1986) on the relationship between yield and salinity of perennial ryegrass and the equation is:

$$Y = 100 - Y_r(ECe - EC_t) \qquad \text{.........} \quad (5.6.3)$$

when $EC_e \geq EC_t$

where,

Y = percentage of yield, and

ECe = electrical conductivity of saturated soil paste (dS m^{-1}).

ECt = threshold salt tolerance of ryegrass = 5.6 dS m^{-1},

Y_r = percentage of yield reduction of ryegrass (7.6 % per unit increase in salinity, dS m^{-1}, from the threshold value).

In the estimation of yield, initial soil salinity and constant groundwater salinity were considered and these were the same. Figure 5.6.4 shows that there was no big difference between the experimental and the estimated yield potential. This finding suggests that though root salinity became much higher, yield reductions were not bigger than predicted using the initial soil salinities.

Dry matter proportions versus weighted average root zone salinity for different treatments are presented in Fig. 5.6.5. The average root zone salinity is calculated for each cut by taking the time and depth average of the salinity within the maximum rooting depth. Fig. 5.6.5 also shows that the dry matter proportions in the salinity treatments are higher than the nonsaline treatment, and that the 15.0 dS m^{-1} treatment produced the highest dry matter proportion.

Noble et al. (1989) studied the irrigation management of perennial pasture using saline irrigation water in Southeast Australia, and reported that ryegrass yield was reduced by 15% when irrigated with groundwater at 4.5 dS m^{-1} initially and that subsequent irrigation water salinity was 2.4 dS m^{-1}. Mehanni and Repsys (1986) found

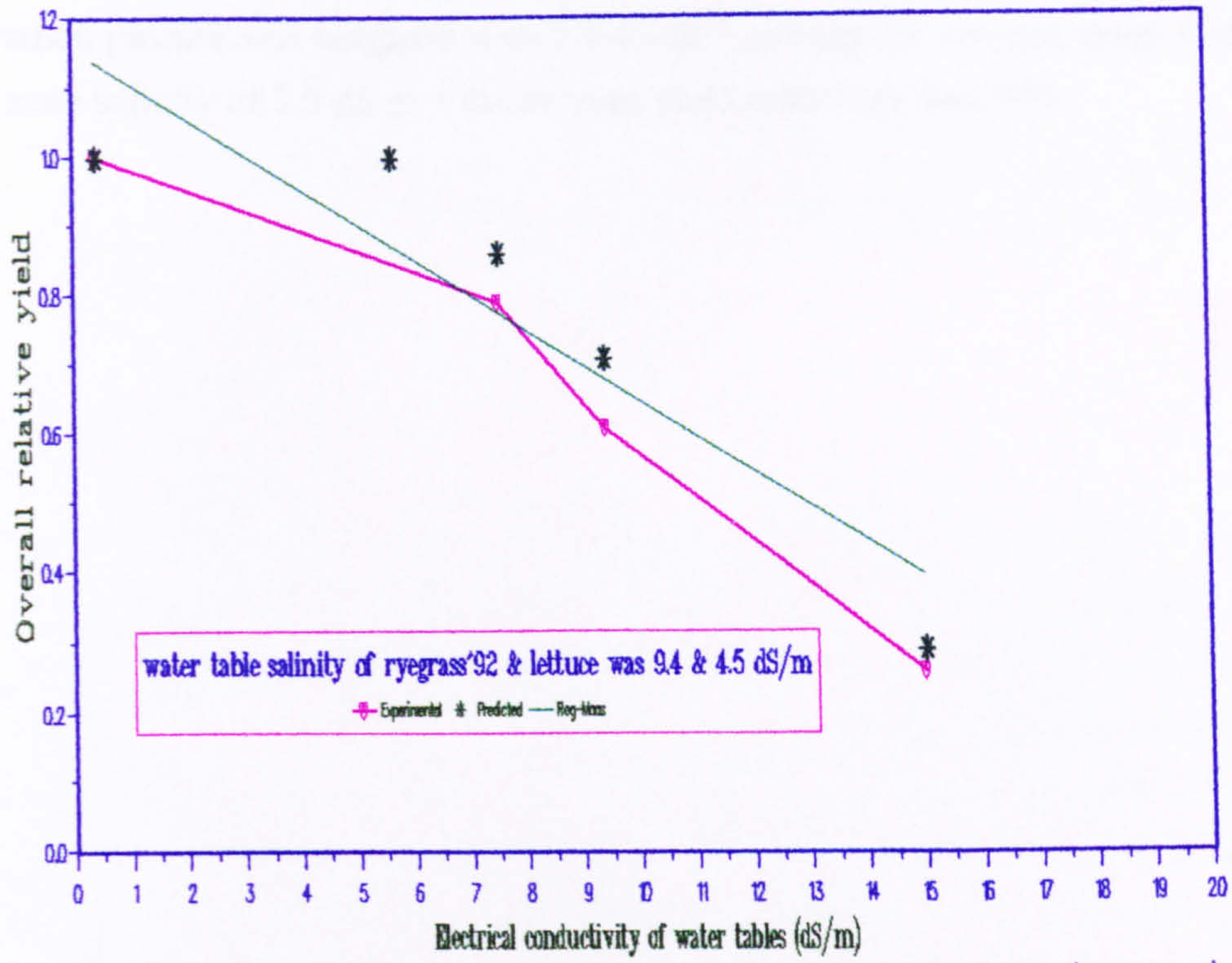


Fig. 5.6.4: Comparison of yield potential for diff. treatments (ryegrass)

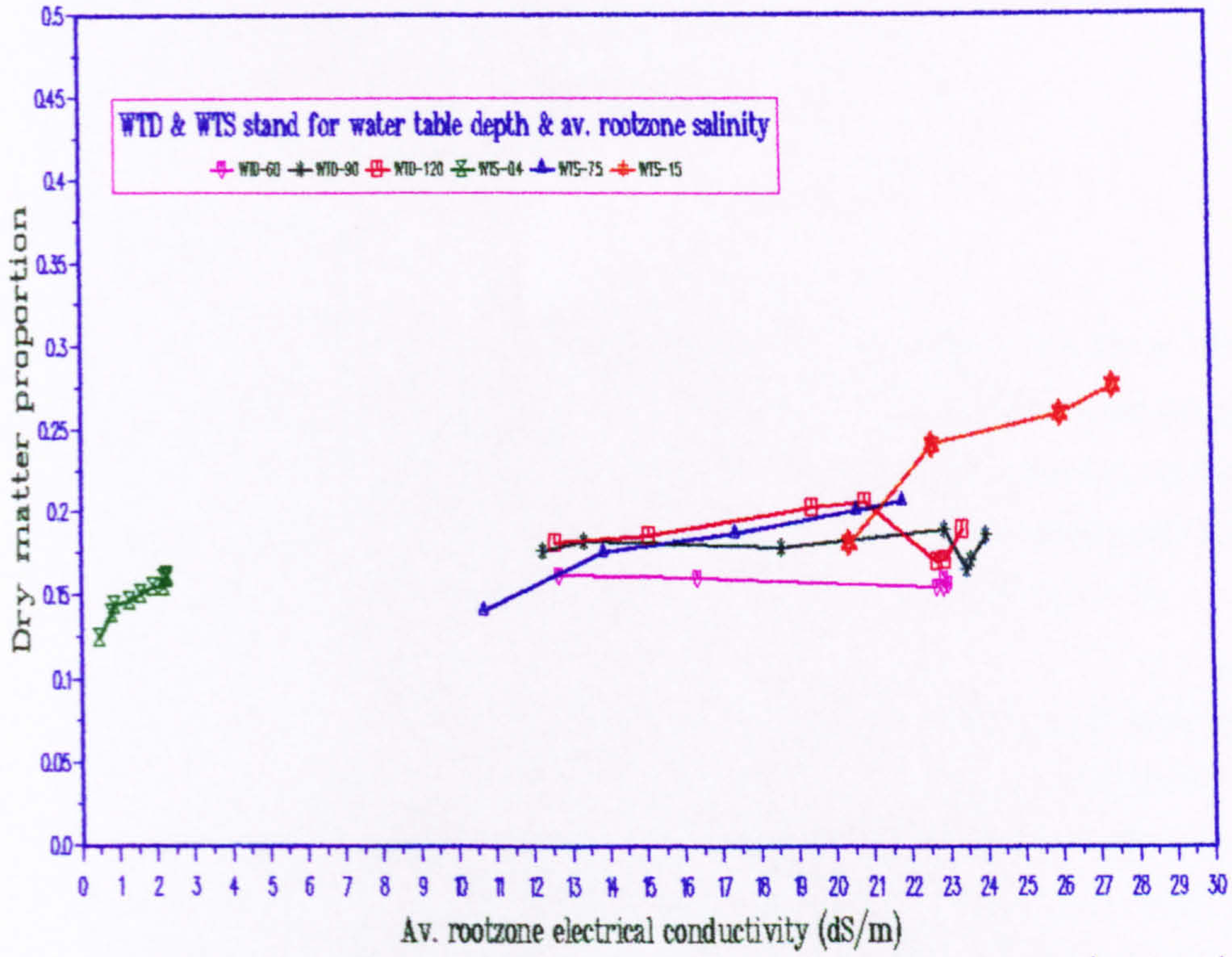


Fig. 5.6.5: Average rootzone salinity versus drymatter proportion (ryegrass)

that when pasture was irrigated with 2.4 dS m^{-1} salinity for several years with average root zone salinity of 5.5 dS m^{-1} the average yield reduction was 30% .

Chapter: 6

MODELLING WATER UPTAKE BY ROOTS UNDER SALINE WATER TABLE MANAGEMENT

6.1 Introduction

Water in the soil is extracted by active roots first in the upper layers creating a soil water matric potential gradient which results in upward flow of water from the water table. Generally, the upper few cm of the root zone cannot be used to extract water, but can be used by allowing soil dryness there-in to drive groundwater upwards. The potential for saline groundwater supply to the crop water needs can be evaluated by knowing the water extraction pattern by roots under wide range of salinity stress. Therefore, the present effort is to simulate the water uptake behaviour under stress conditions based on our experimental results. The information generated on the effect of stress on root water uptake can also be utilized in proper irrigation scheduling.

6..2 Sink term: concept and development

6.2.1 Sink term concept

The widely used concept of defining water uptake by roots by a 'sink' term is based on a macroscopic view. The sink term generally is assumed to be proportional to the difference in potential between the soil water and the root interior, to the hydraulic conductivity of soil and to some effective root function. It is an additional function used with the continuity equation of soil water flow (equation 3.6 in Chapter 3).

Two different terminologies are used for the microscopic and macroscopic approaches, because of the differences in handling the water uptake by roots. The microscopic approach considers the moisture flow process in the vicinity of a single root and is then integrated with respect to the whole domain of root zone. This approach is based on the mathematical solution to a single root, which is considered as a hollow cylinder of infinite length with uniform diameter and water extraction properties. In radial co-ordinates, the flow equation is expressed as :

$$\frac{\partial\theta}{\partial t}=\frac{1}{r}\frac{\partial}{\partial r}\left[rD(\theta)\frac{\partial\theta}{\partial r}\right] \qquad \text{..................(6.1)}$$

where,

r = radial distance from the axis of the root (L),

θ = volumetric moisture content ($L^3\ L^{-3}$),

D = hydraulic diffusivity ($L^{-2}\ T^{-1}$), and

t = time (T).

In contrast, the macroscopic approach refers to the flow process in an entire root zone, i.e. the extraction of water from each differential volume of the root zone is assumed at some rate, the 'sink'. Therefore, the sink term is presented as a negative source in the soil water flow equation. In this approach, the root water extraction is inferred from the measurements of soil physical properties. The flow equation is modified to account for a sink term that represents the roots and it is defined as follows:

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial z}\left[K\frac{\partial H}{\partial z}\right]-S \qquad \text{..................(6.2)}$$

where,

θ = volumetric water content ($L^3\ L^{-3}$),

H = hydraulic potential (L),

K = hydraulic conductivity ($L\ T^{-1}$),

t = time (T), and

S = sink term ($L^3\ L^{-3}\ T^{-1}$).

The root extraction or sink term, S, depends in part on the root length density distribution (the fraction of total active roots per unit volume of soil). However, the comcept of macroscopic approach (Equation 6.2) will be used in the present modelling.

Basically, the sink term is used to distribute the atmospheric demand over the root zone and decrease the water extracted by plant roots according to the stress conditions (either matric or osmotic or both) of soil water in the root zone.

6.2.2 Rooting depth consideration

The models developed (section 6.2.4) usually use the maximum rooting length and the time of reaching the maximum length used as input to the model. The increase in rooting length from root initiation to the maximum length is calculated by linear, or sigmoidal, or exponential approximation. The maximum rooting length differs from crop to crop and the rooting length of any crop is restricted by salinity stress. In the present experiments, it is observed that the maximum rooting length and the rooting mass varies inversely with the root zone salinities.

6.2.3 Water extraction pattern by roots

The water extraction pattern by roots is related to the active rooting mass developed within the root zone. The extraction function is sometimes expressed by the usual agronomic rule of thumb, in which the water uptake from each quarter of the root zone (moving downward from the surface) is assumed to be approximately 40, 30, 20 & 10 % respectively, or some equivalent equation(s). This is because the root mass development, diameter of the roots, number of roots developed and physiological maturity (causing water extraction to decline or cease), all vary with time. Even the expression of root length per unit volume of soil is not sufficient to estimate the inflow to roots because of the above mentioned variations. For example, the visual observation in the present experiments is that the diameter of roots in nonsaline and saline treatments differed markedly.

6.2.4 Sink term approaches for saline water management

Childs and Hanks (1975) expressed the sink term for saline water management, which is the modification of the model of Nimah and Hanks (1973), by adding the osmotic potential factor for defining soil water salinity. Their equation is:

$$S = \frac{[H_{Root} + RRES \times Z - \Psi_m - \Psi_o]}{dx.dz}[RDF \times K(\theta)] \quad .....(6.3)$$

where,

H_{root} = root water potential at the soil surface where Z is considered zero (L),

Ψ_m = soil matric potential (L),

Ψ_o = soil water osmotic potential (L),

$K(\theta)$ = soil hydraulic conductivity (LT^{-1}),

RDF = proportion of total active roots in depth increment dz,

Z = depth (L),

dx = distance between plant roots and the point in the soil where Ψ_m and Ψ_o are measured (L),

RRES = head loss coefficient for longitudinal water flow in the root xylem (generally assumed to be 1.05).

Cardon (1990) tested this model for simulating water uptake by alfalfa with a water table salinity of 6.0 dS m^{-1} along with irrigation and found that the model was not very sensitive to salinity, and the overall performance was not satisfactory. The reason is that the negative effect of osmotic potential does not have affect on the hydraulic conductivity, but greater effect on the gradient terms.

Based on the concept of determining a stress coefficient under saline conditions, Van Genuchten (1987) developed an empirical equation including a root distribution function. The relationship between yield and transpiration based on available crop salt tolerance data was the basis of his equation. The equation is:

$$S = \left[\frac{S_{max}}{1 + \left(\frac{a\Psi_m + \Psi_o}{\Psi_{o(50)}} \right)^3} \right] RDF \qquad \text{.......... (6.4)}$$

$$\text{and} \quad RDF = \left| \begin{array}{ll} \frac{5}{3}L & \text{when } z \le 0.2L \\ \frac{25}{12}L\left(1 - \frac{z}{L}\right) & \text{when } 0.2L \le z \le L \\ 0.0 & \text{when } z > L \end{array} \right|$$

where,

$S_{max} = T / L \qquad T = E_o \times K_{cr}$

E_o potential evaporation (LT^{-1}),

T potential transpiration (LT^{-1}),

K_{cr} crop coefficient adapted from available experimental data, and

a a coefficient which equals $\Psi o_{(50)} / \Psi m_{(50)}$, where suffix 50 is used to denote the respective values at 50% yield reduction level,

Ψm soil matric potential (L),

Ψo soil water osmotic potential (L),

L rooting depth (L),

z soil depth (L), and

RDF implies root distribution function related to depth (dimensionless).

The equation for RDF reflects the percentage of crop water uptake from each quarter of the root zone moving downward from the surface is assumed to be approximately 40, 30, 20 & 10 % respectively (Cardon, 1990).

Ismail and Gowing (1990) developed an empirical model for crop response to salinity & irrigation in which the sink term was formulated as:

$$S = \frac{E_a}{ROOT \times 10.0} \times t_{inc} \qquad \text{.............................(6.5)}$$

$$\text{when } E_a = E_p \left[a_1 (c_1 . \Psi_t + d_1) \exp^{-(b_1^2 . \Psi_t . E_p)} \right]$$

$$\text{and} \qquad ROOT = \frac{L_{max}}{1 + e^{6.0 - 12.0 \frac{t_p}{t_{max}}}}$$

where,

E_a = actual evapotranspiration (mm d^{-1}),

t_{inc} = time increment from the planting date (d),

E_p = potential evapotranspiration (mm d^{-1}),

Ψ_t = total potential (matric + osmotic) in soil water (bar),

L_{max} = maximum root depth (cm),

t_{max} = number of days to root profile maturity (d),

t_p = time (d) considering planting date equals 1,

a_1, b_1, c_1 & d_1 = empirical constants

(a = 0.3372, b = 0.06105, c = 0.005747 & d = 2.69).

The model had been validated for cotton. They pointed out the limitations of using the model as:

i) water table was assumed so far from the root zone that is not effective, and

ii) the calculation of E_a / E_p is derived by using data representing maize (with a few modifications), so that it is probable that the model may be less realistic for other crops.

El-Hessy (1991) used the Van Genutchen (1987) equation, but a time dependent rooting depth was calculated as follows:

The root depth calculated using the inputs of i) minimum and maximum root depth, ii) date of planting, and iii) date of reaching maximum root depth was as follows :

$$L_t = \frac{L_{max} - L_i}{t_m - t_{pj}} \times (t_j - t_{pj}) \qquad ....(6.6)$$

where,

L_t = Root depth at day d (L),

L_{max} = maximum root depth (L),

L_i = initial root depth (L),

t_m = day from planting to reach maximum root depth (Julian day),

t_{pj} = planting day (Julian day), and

t_j = present day (Julian day).

Cardon and Letey (1992) modified Van Genuchten's (1987) potential transpiration term into a stress adjusted potential transpiration term using a stress adjusted crop coefficient. The reason for modification was because they found that the Van Genuchten equation was only appropriate for modelling of root water uptake over short time when the maximum transpiration rate, $S_{max,}$ and rooting depth, L, can be considered constant. For season-length simulations, S_{max} and L are time dependent functions dictated by climate and soil profile conditions. Moreover, many crops exhibit differential tolerance to soil moisture deficit and salinity stress at different growth stages. Cardon and Letey expressed the sink term as follows:

$$S = \left[\frac{S'_{max}}{1 + \left(\frac{a\ \Psi_m + \Psi_o}{\Psi_{o(50)}} \right)^3} \right] RDF \qquad .... (6.7)$$

where,

S'_{max} = stress-adjusted value of S_{max} , substituting K_a for K_{cr}.

But, $K_a = K_{cr}\,(CT/CT_a)$;

when, K_a = stress-adjusted crop coefficient,

CT = cumulative potential transpiration (L), and

CT_a = cumulative stress-adjusted potential transpiration (L).

Many forms for describing the root water uptake are available and although they are acceptable for some simple specific conditions, they cannot predict the water uptake function well enough in more complex situations. That is why a gradual development and modification of the model is continuing. Basically, the root uptake function is so dynamic (because it is biologically governed) that any assumption(s) could not account for different situations.

However, the reported saline water management models are mainly related to saline water tables coupled to irrigation with sweet or water that is less saline than the water table salinity. From those models or investigations, the actual trends of water uptake from the water table only were not reflected accurately. Moreover, the main basis of the models was the consideration of matric and osmotic potentials at 50% yield reduction level and though the specific values of osmotic potentials for different crops are available (Rhoades and Loveday, 1990) the specific matric potentials for different values of yield reduction are less available or even scarce. In addition, the water uptake model based on the osmotic potential at 50% yield reduction level [$\Psi o_{(50)}$] (Van Genuchten, 1987) results in greater reduction in water uptake rate when the salinity in the root zone is higher than $\Psi o_{(50)}$.

6.3 Present approach of the sink term

The present model of the sink term attempts to generate a simple, and general approach to water uptake by roots under both saline and nonsaline situations for estimating crop water available from a shallow water table. The main consideration of the model is to estimate the limiting capacity of water uptake by roots when any soil water stress lowers the root uptake potential, i.e. based on available water to plants. It can be expressed by the equation:

$$S = E_p \left[\frac{\Psi_L - \Psi_t}{\Psi_L} \right] \times RDF \qquad ....(6.8)$$

where,

Ψ_L = leaf water potential assumed to be -1.5 MPa (this assumption is discussed in section 3.1.3),

Ψ_t = total potential (matric + osmotic) of soil water in the root zone (MPa),

E_p = potential evapotranspiration (LT^{-1}),

S = water uptake by roots (LT^{-1}) and

RDF = proportion of total active roots, the usual agronomic thumb rule.

6.4 Test of the model

The model as in equation (6.8) was tested under the following conditions:

Measured Pan evaporation:	Varied from 2.5 to 1.0 mm d^{-1} and an average 2.0 mm d^{-1}.
Initial soil water salinity:	0.4, 7.5 & 15.0 dS m^{-1} for ryegrass'93, 9.4 dS m^{-1} for ryegrass'92.
Constant water table salinity:	0.4, 7.5 & 15.0 dS m^{-1} for ryegrass'93, 9.4 dS m^{-1} for ryegrass'92.
Constant water table depth:	90 cm,
Cropping period:	157 days for ryegrass'93-7.5 dS m^{-1} and 215 days for other treatments.
Maximum rooting depth:	75 cm for 0.4 dS m^{-1}, 45 cm for 7.5 & 9.4 dS m^{-1} & 20 cm for 15.0 dS m^{-1} ryegrass treatments.

Initial matric potential of soil profiles: 0 to -120 cm from water table to soil surface.

The time-dependent matric potential and salinity of soil water in the root zone were calculated on the basis of water uptake (S) in the preceding day using the soil moisture characteristic , and upward flow equations (3.9 and 3.16 in chapter 3). The calculated salinity is converted into osmotic potential using a standard conversion factor. The initial matric and osmotic potentials at each depths, pan evaporation, and maximum rooting depth are the inputs in the model. The potential evapotranspiration was approximated from the measured pan evaporation in the experimental location. Though there are some shortcomings in obtaining an estimate of the climatically driven potential evapotranspiration from pan evaporation, it can be used to assess potential evapotranspiration, if the pan is properly sited and maintained (Hillel, 1987). As the present investigations were conducted in a glasshouse where there was little wind and the evaporation rate was less than 2 mm per day, the pan coefficient was assumed to be 0.8 (Doorenbos and Pruitt, 1977). The time to reach the maximum rooting depth is calculated considering uniform root extention at 10 mm d^{-1} for the 0.4 dS m^{-1} treatment. The same time period to reach the maximum rooting depth was considered for salinity treatments and only the root development rate varied with salinity.

6.5 Results and discussion

Fig. 6.1.1(a) & (b) compares predicted and measured cumulative water uptake by roots for these experiments. Fig. 6.1.1(a) shows that there is a very good agreement between the predicted and the measured values for the 0.4 dS m^{-1} treatment up to 150 days, though the predicted uptake is somewhat higher at later times. Figs. 6.1.1(a) & (b) show that, for the other salinity treatments, the model predicted higher water uptake than measured. The over-prediction is greatest in the highest salinity. The predicted results for variations of salinity and water table depth are also shown in Fig. 6.1.2. It shows that the model responds to the effect of variations due to salinity and water table depth. Comparison of water uptake at different leaf water potentials for 9.4 dS m^{-1} at 60 cm water table treatment is presented in Fig. 6.1.3. It shows that a leaf water potential of -1.2 MPa did not allow any water uptake after 130 days and, similarly, -1.5 MPa resulted in a lower uptake than the measured after 150 days, as the slope of the predicted line declined faster than the slope of the measured uptake. Again, considering a leaf water potential of -2.0 MPa, the slope of the predicted and the measured uptake became consistent which may indicate that the osmotic adjustment by the plants happened after 150 days of growth.

The simulated and measured moisture contents at various depths of the soil profile for different treatments are shown in Figs. 6.2.1(a), (b) & (c) and 6.2.2(a), (b) & (c) for ryegrass'93 and ryegrass'92 respectively. These Figures represent good agreement between the simulated and measured soil moisture content as a whole, except for some divergence below 30 cm depth in the early part of the growing seasons. Similar results on the trend of root zone salinization are presented in Figs. 6.3.1(a), (b), & (c) and 6.3.2(a), (b) & (c) for the '92 and '93 experiments with ryegrass. The same over-prediction was found in the higher salinity treatments due to the overprediction of total water uptake. The salinity prediction is based on the convective transport of solutes, and therefore, depends strongly on water uptake. The hydrodynamic dispersion effect was not incorporated in the model.

The evaluation of predicted versus measured water uptake rate (mm d^{-1}) can be made by calculating two objective functions. The first is RMSE (root mean square error), which is calculated as:

$$\text{RMSE} = \left[\frac{\sum_{i=1}^{N} (P_i - M_i)^2}{N} \right]^{0.5} \qquad \text{...........(6.9)}$$

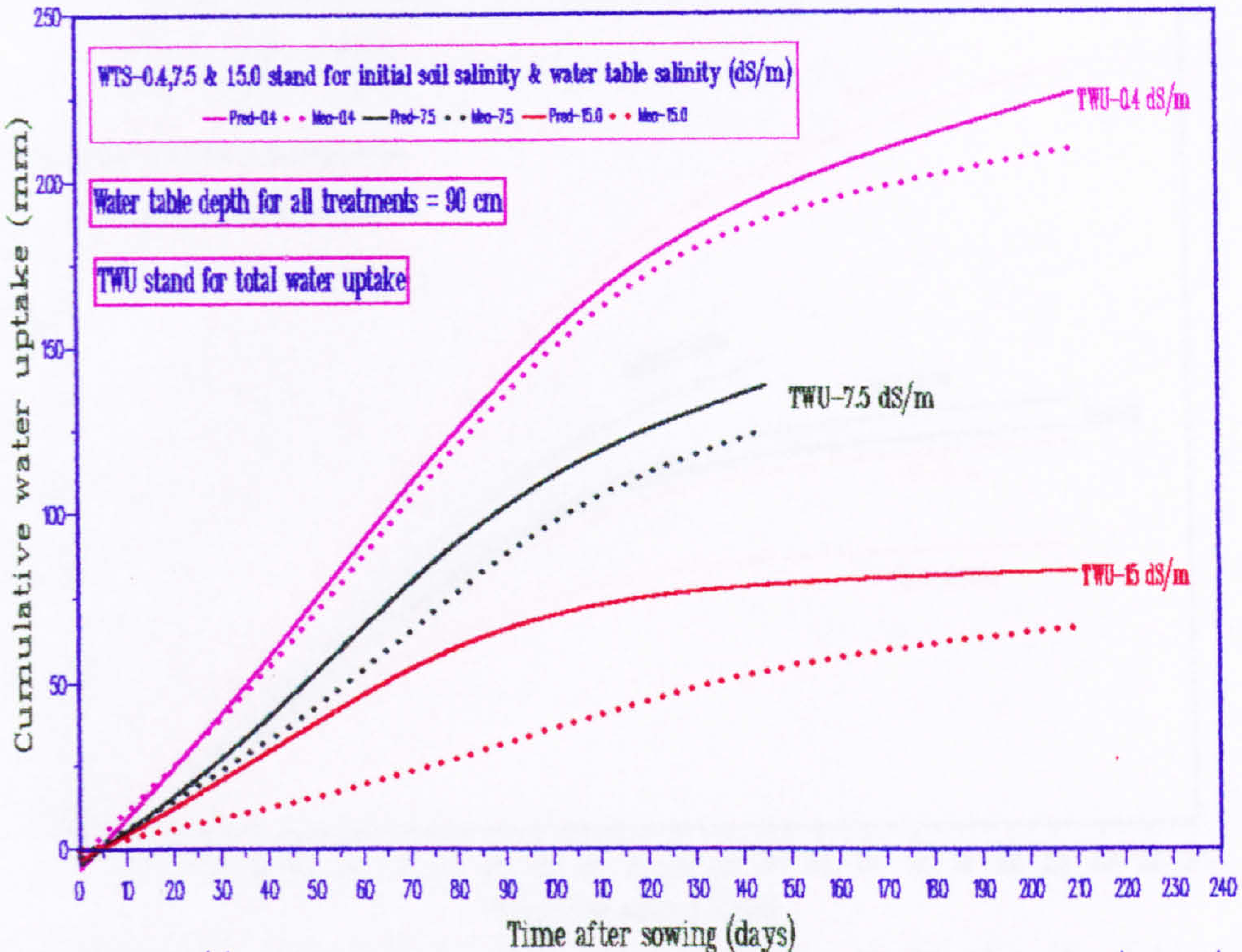


Fig. 6.1.1(a): Predicted & measured water uptake for diff. salinities (rye'93)

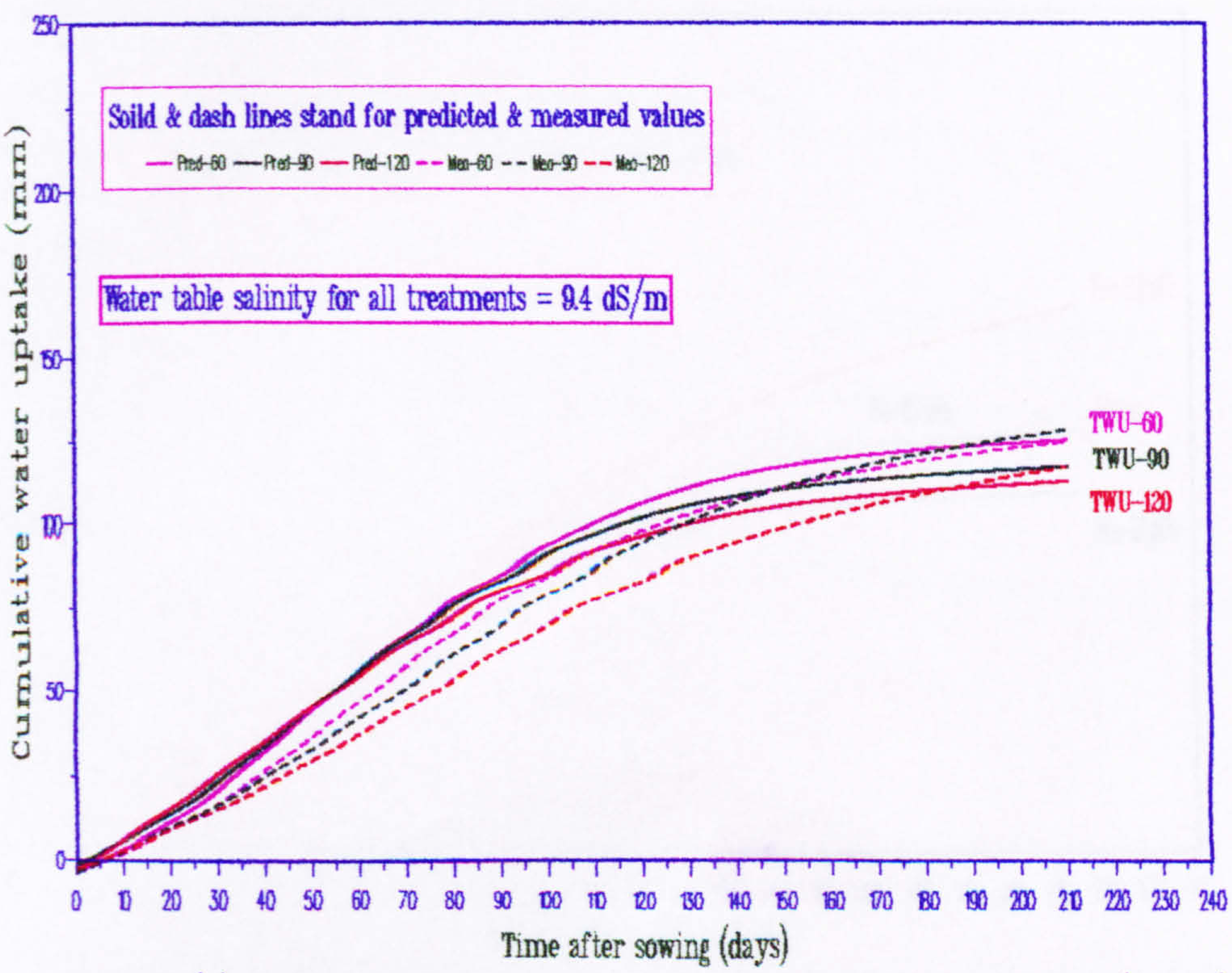


Fig. 6.1.1(b): Predicted & measured water uptake for diff. water table (rye'92)

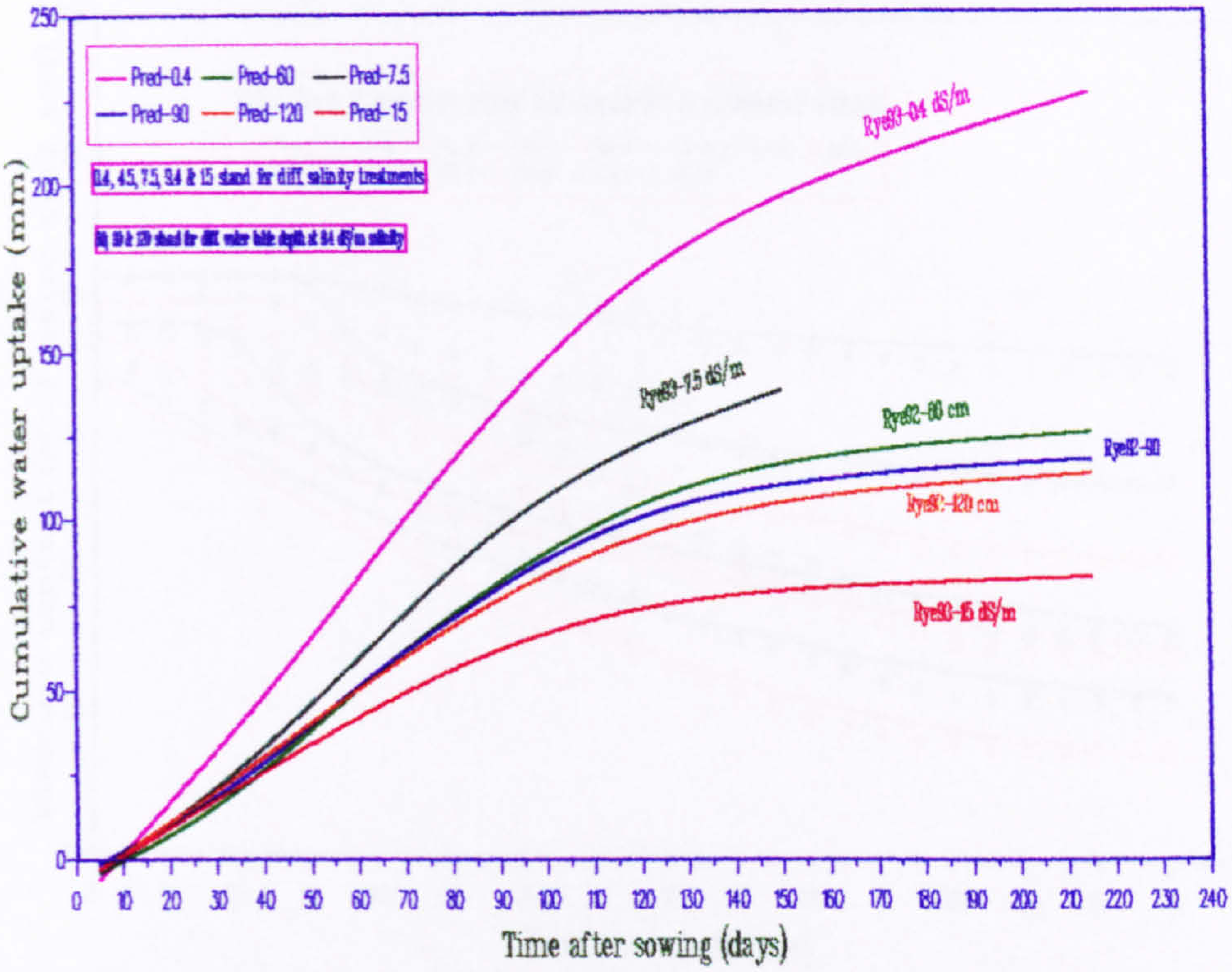


Fig. 6.1.2: Comparison of water uptake at diff. salinities & water tables

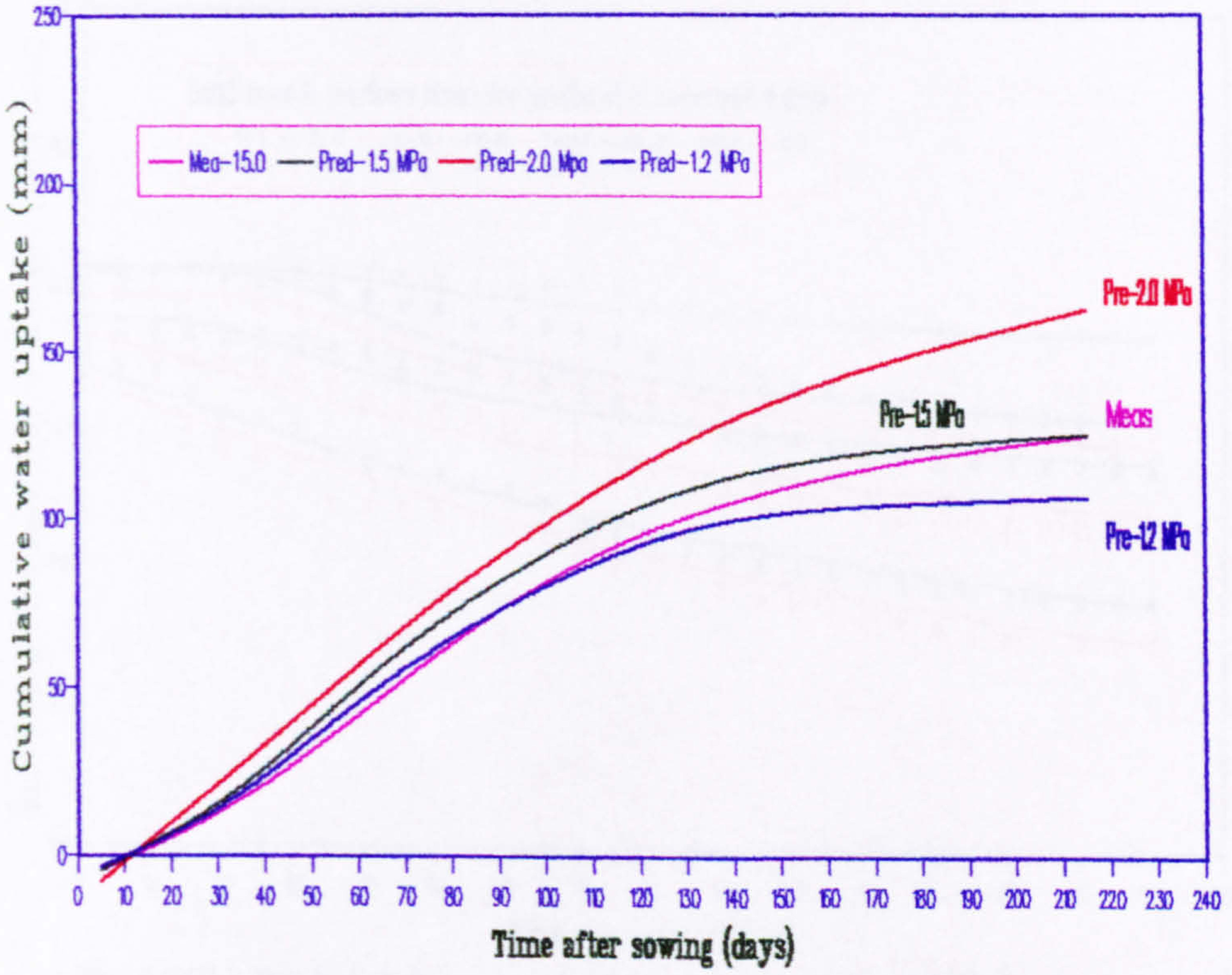


Fig. 6.1.3: Comparison of water uptake at diff. leaf water potential (9.4 dS/m)

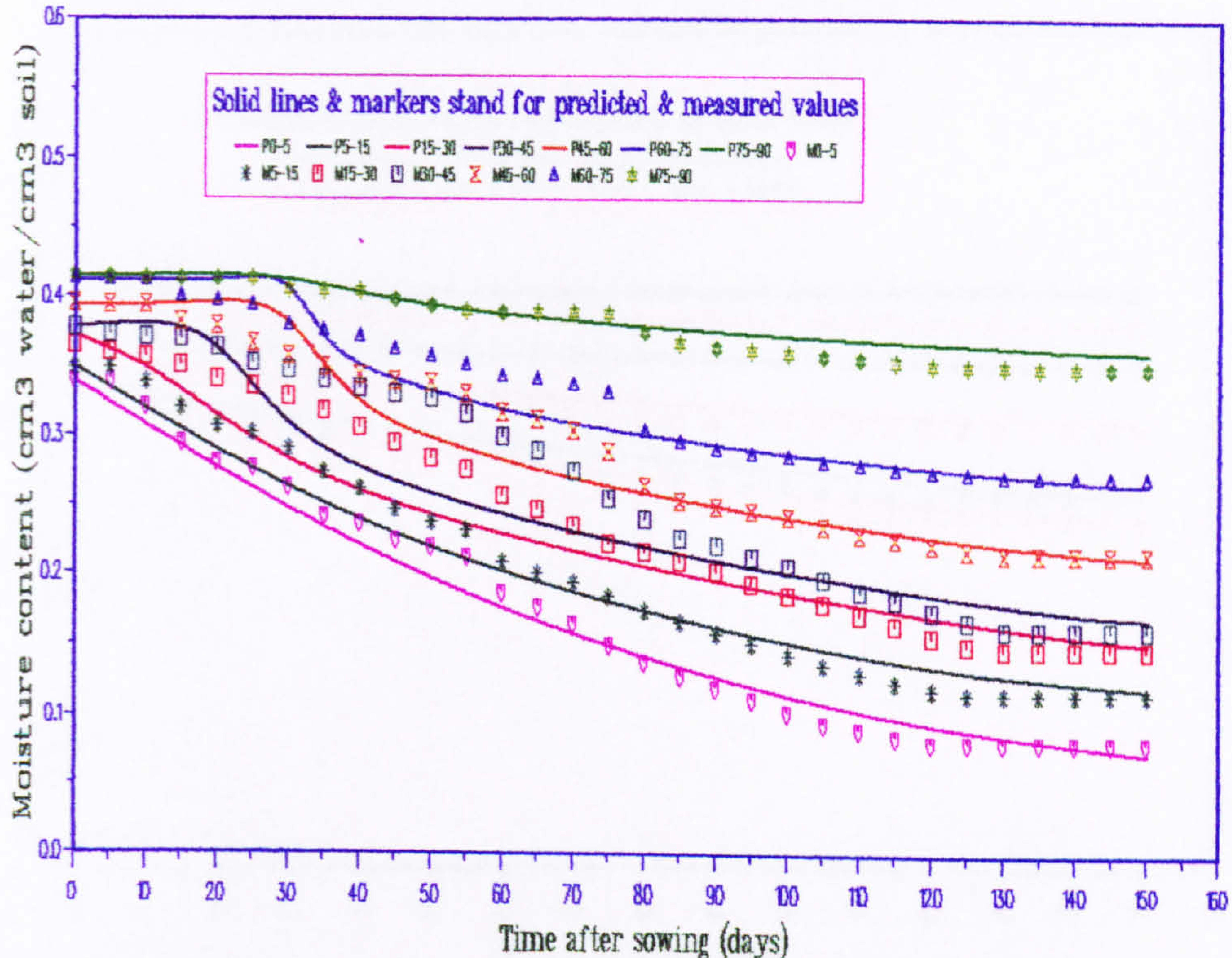


Fig. 6.2.1(a): Predicted & measured moisture content for 0.4 dS/m (ryegrass'93)

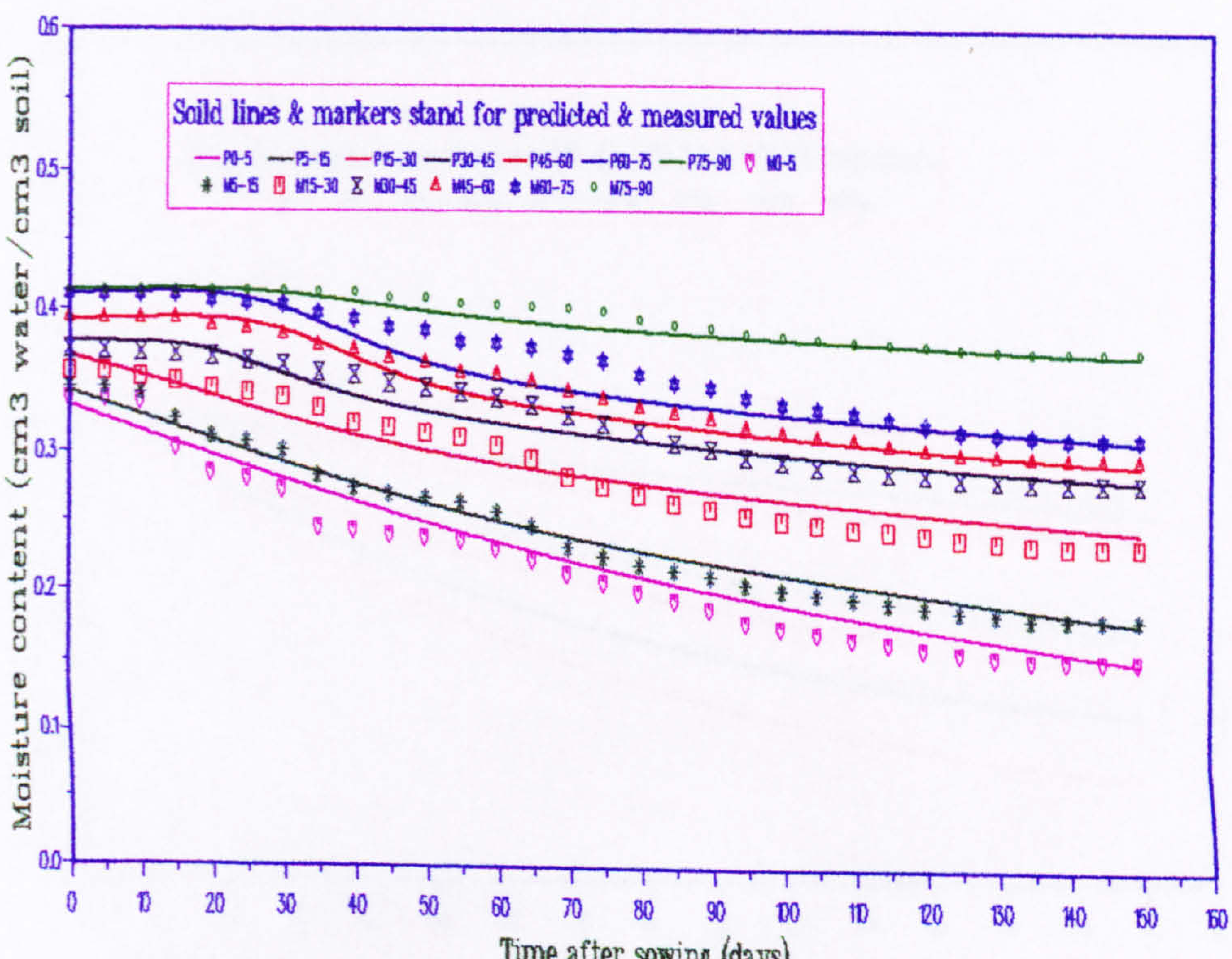


Fig. 6.2.1(b): Predicted & mesured moisture content for 7.5 dS/m (ryegrass'93)

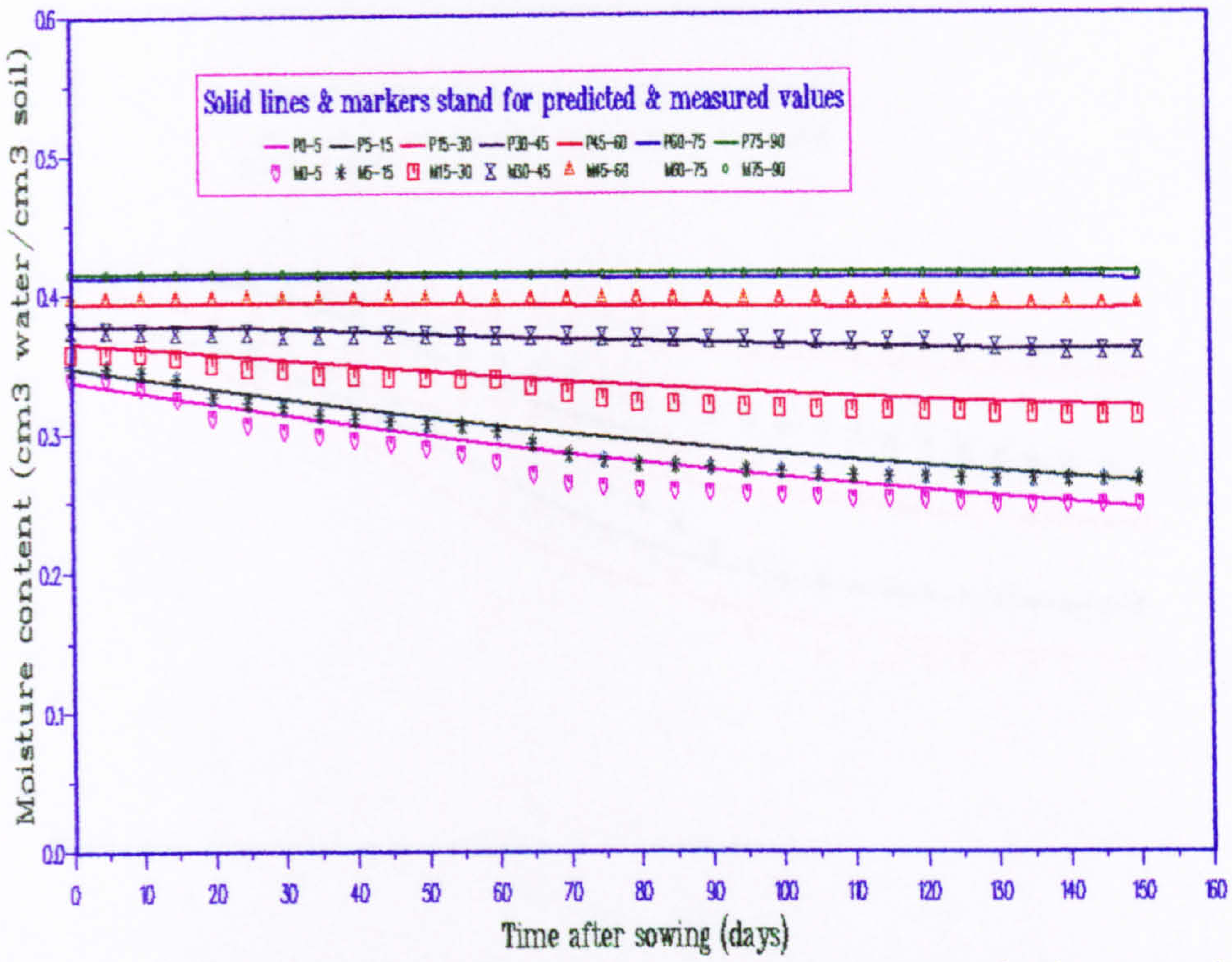


Fig. 6.2.1(c): Predicted & measured moisture content for 15.0 dS/m (ryegrass'93)

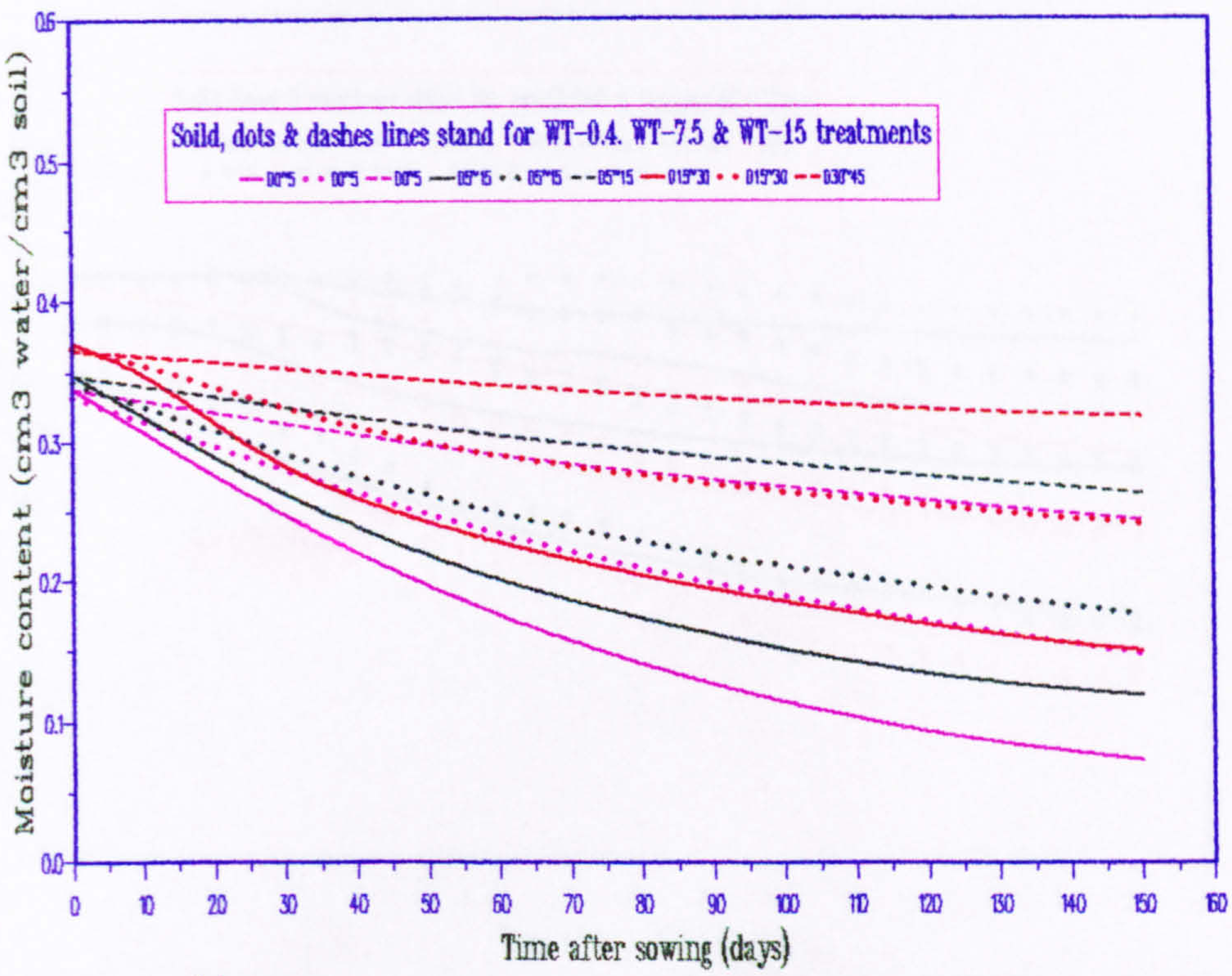


Fig. 6.2.1(d): Comparison of predicted moisture content for salinity treatments

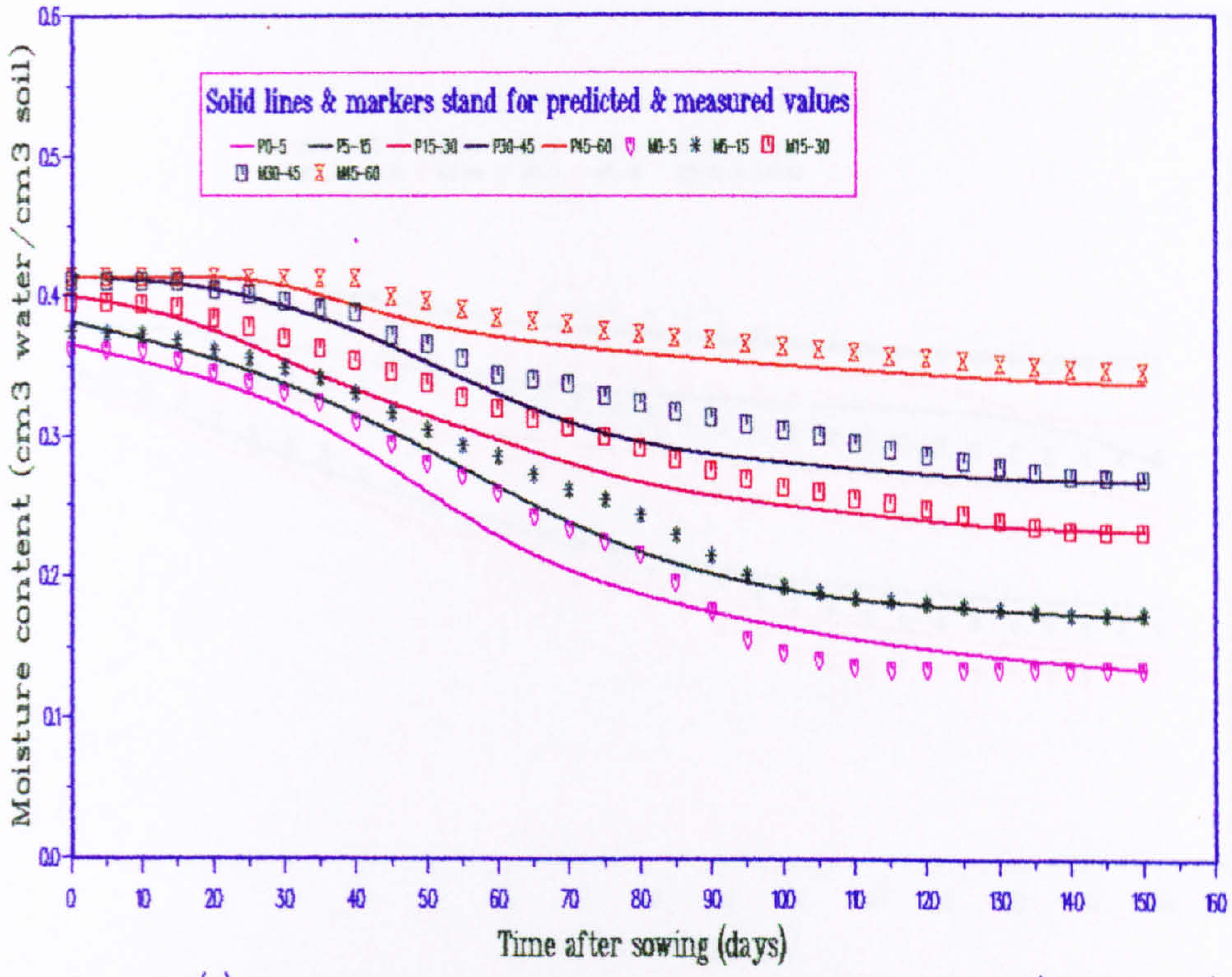


Fig. 6.2.2(a): Predicted & mesured moisture content for WT-60 (ryegrass'92)

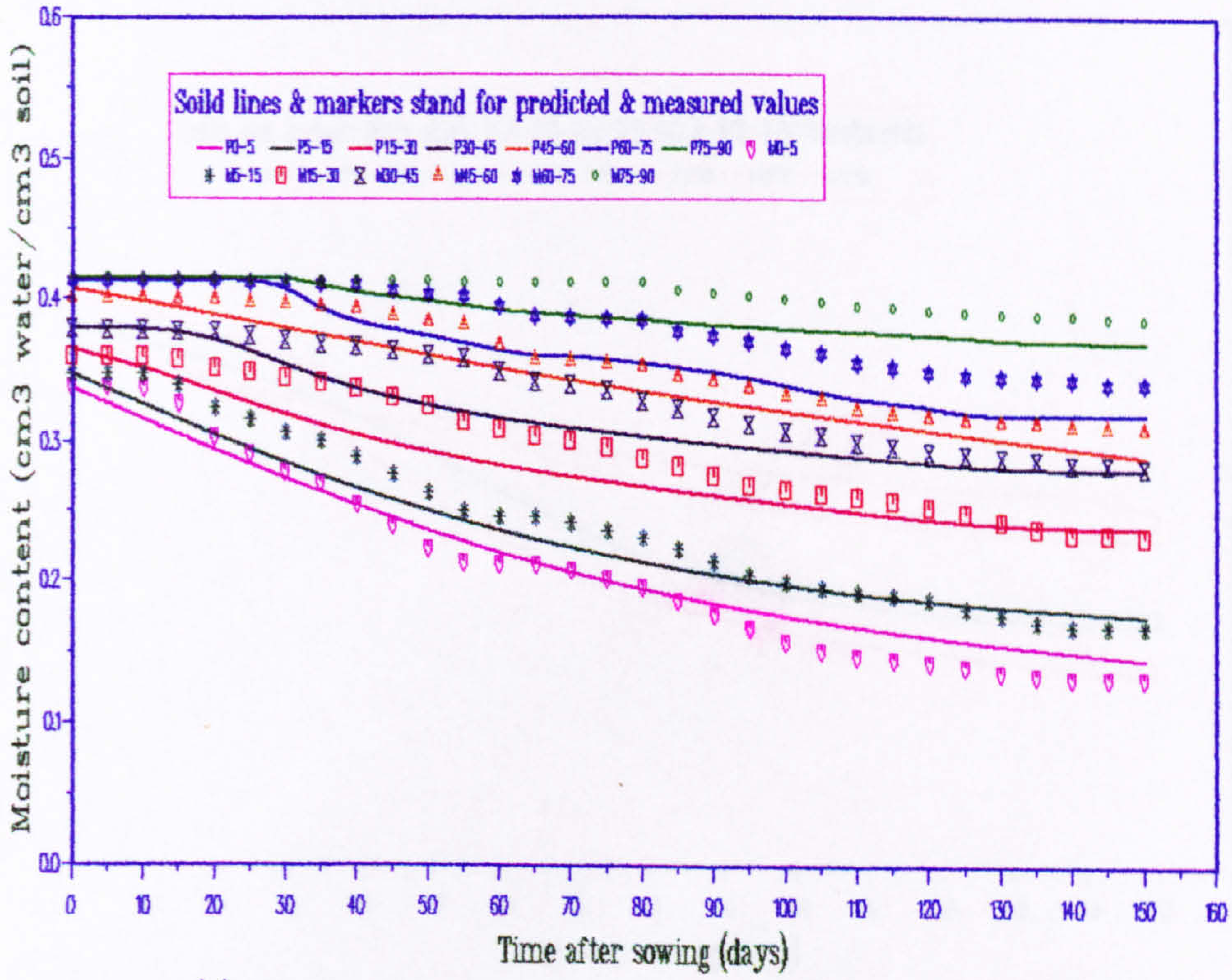


Fig. 6.2.2(b): Predicted & measured moisture content for WT-90 (ryegrass'92)

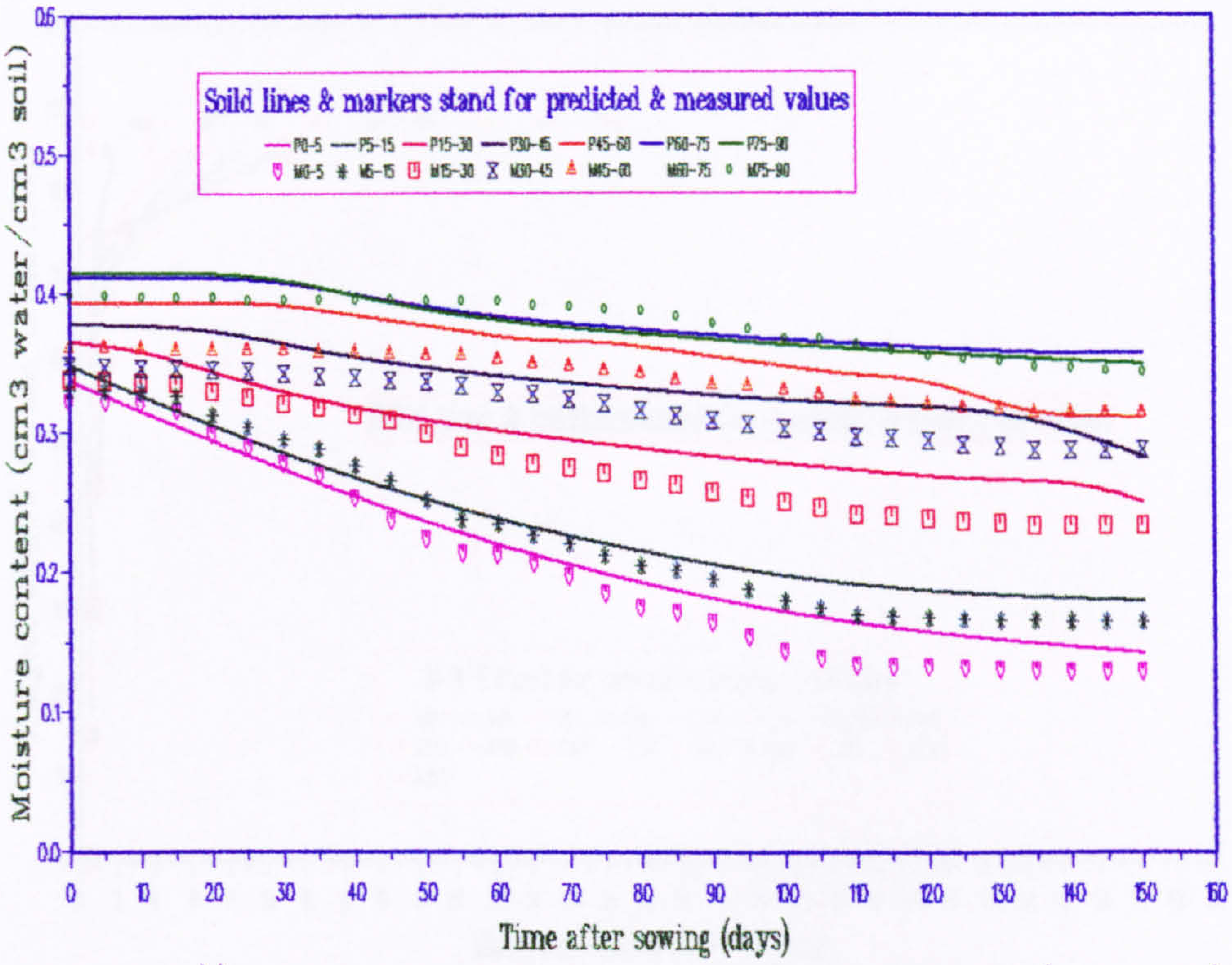


Fig. 6.2.2(c): Predicted & measured moisture content for WT-120 (ryegrass'92)

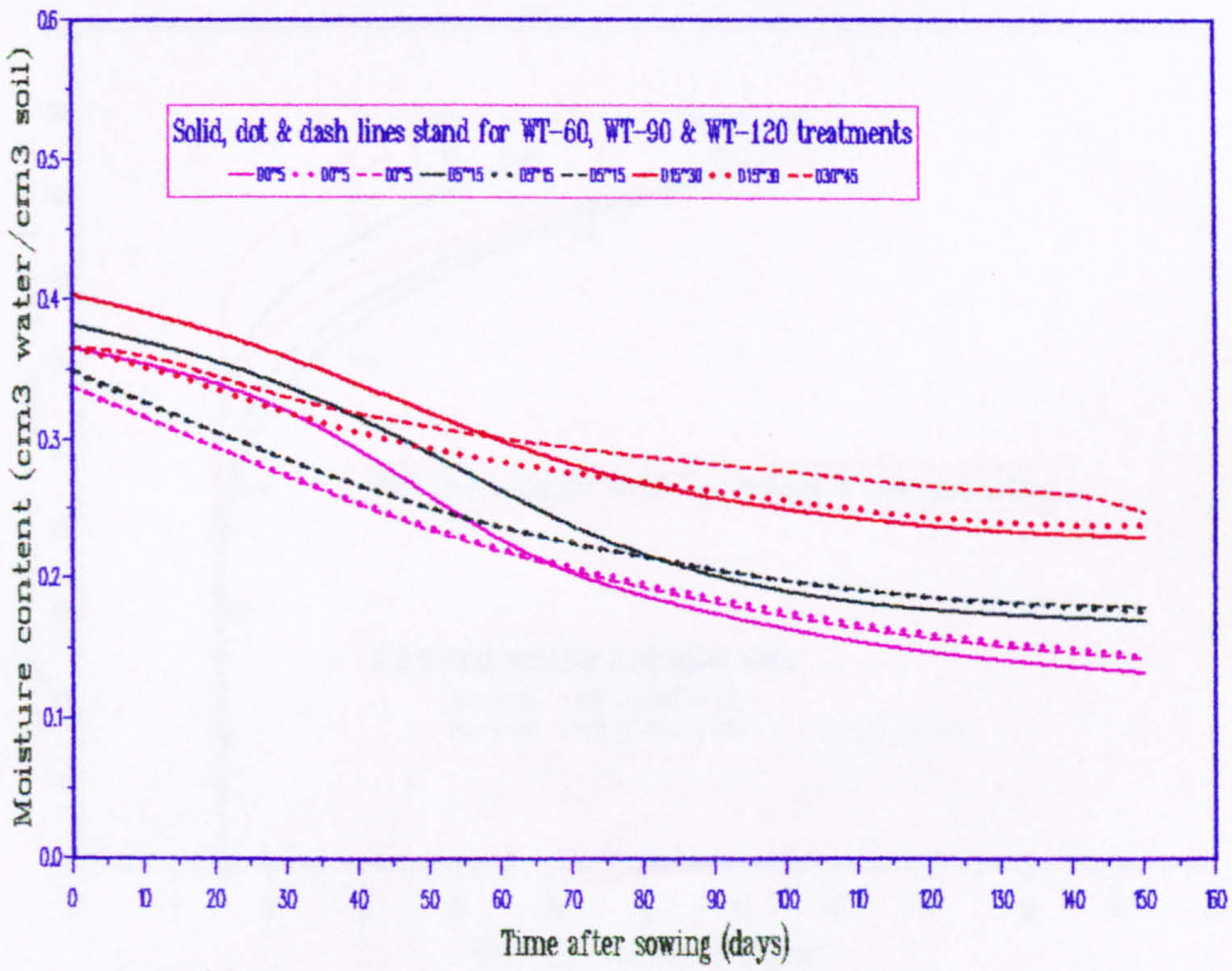


Fig. 6.2.2(d): Comparison of predicted moisture content among water tables

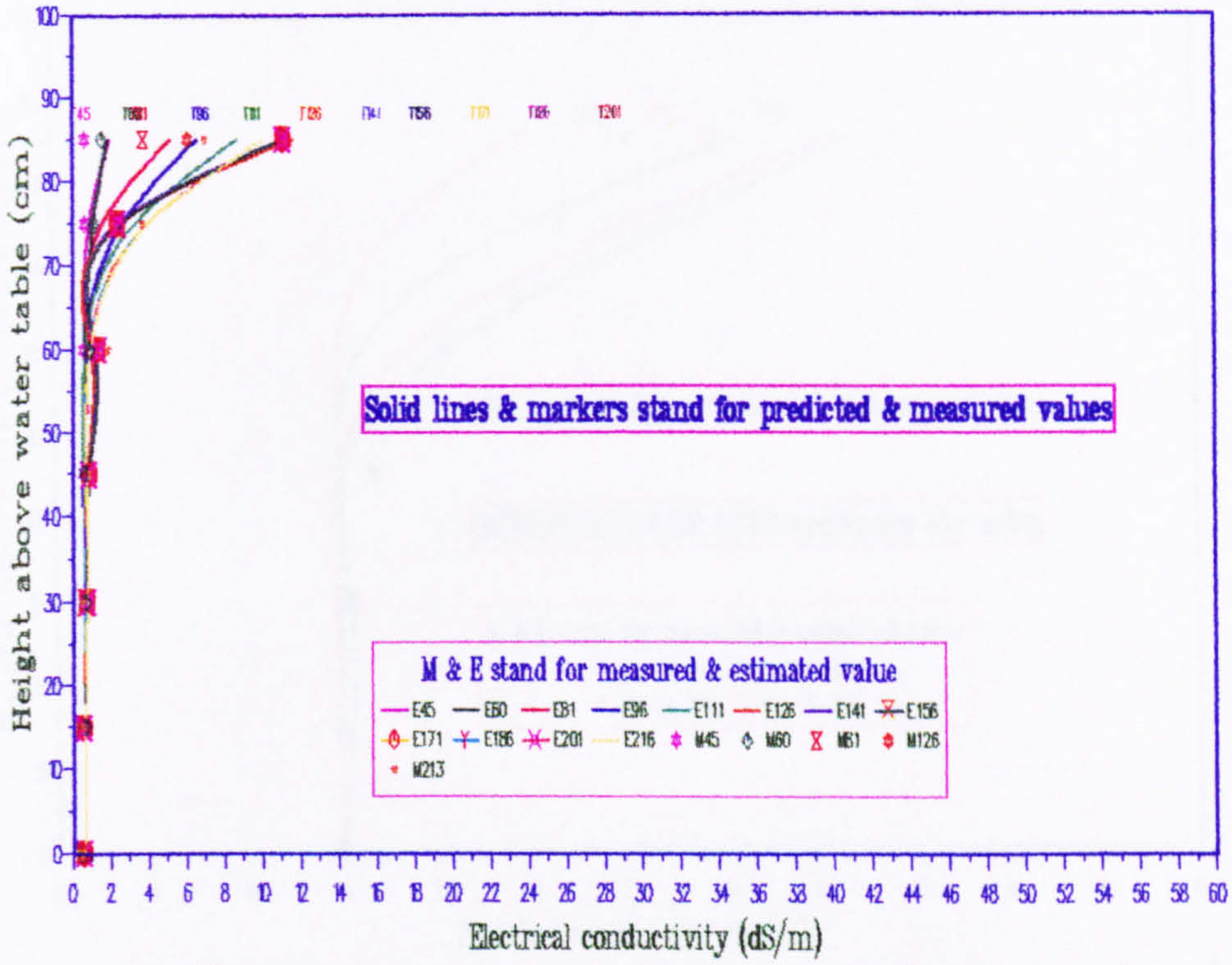


Fig. 6.3.1(a): Predicted & measured salt profile in 0.4 dS/m lysimeter (rye'93)

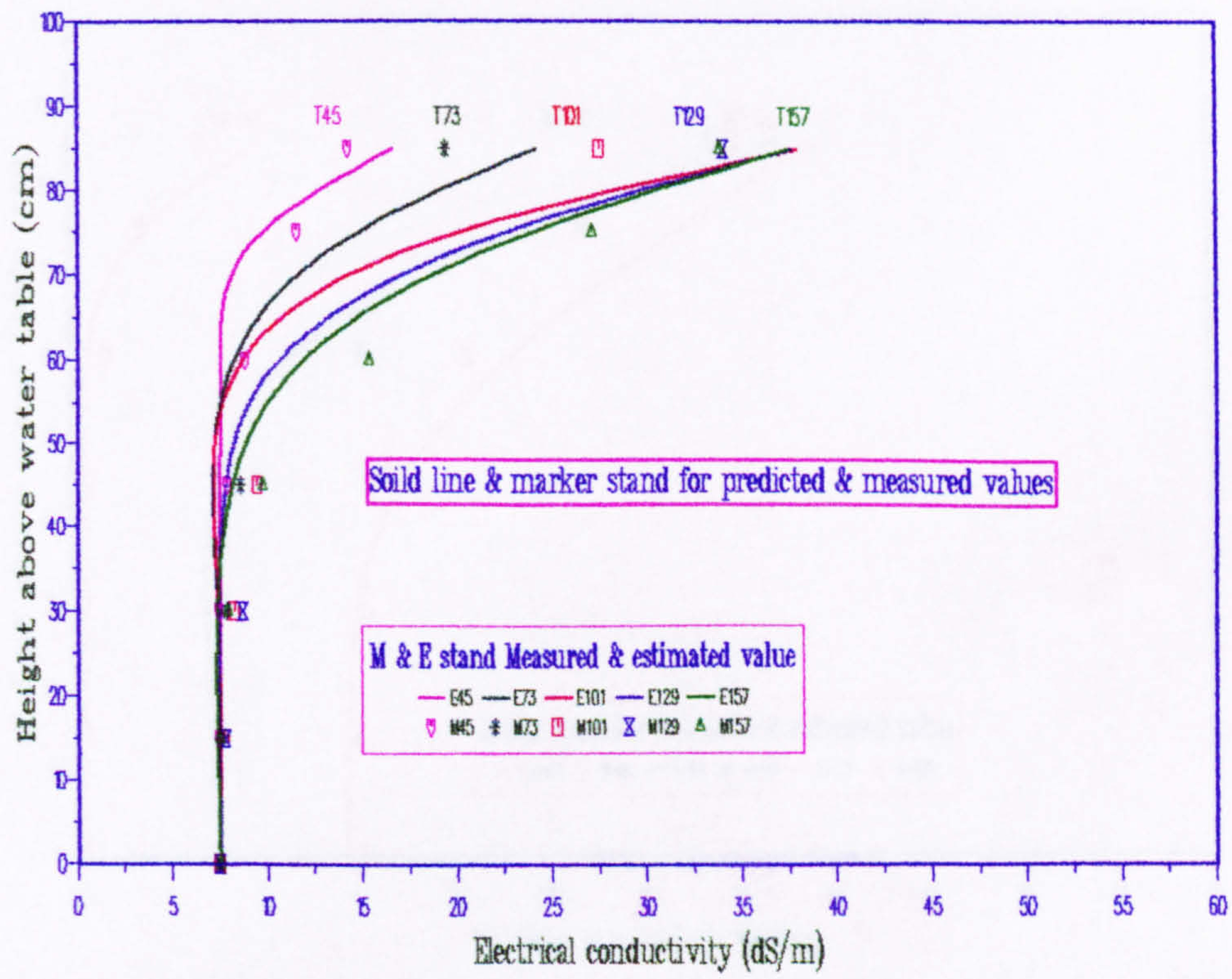


Fig. 6.3.1(b): Predicted & measured salt profile in 7.5 dS/m lysimeter (rye'93)

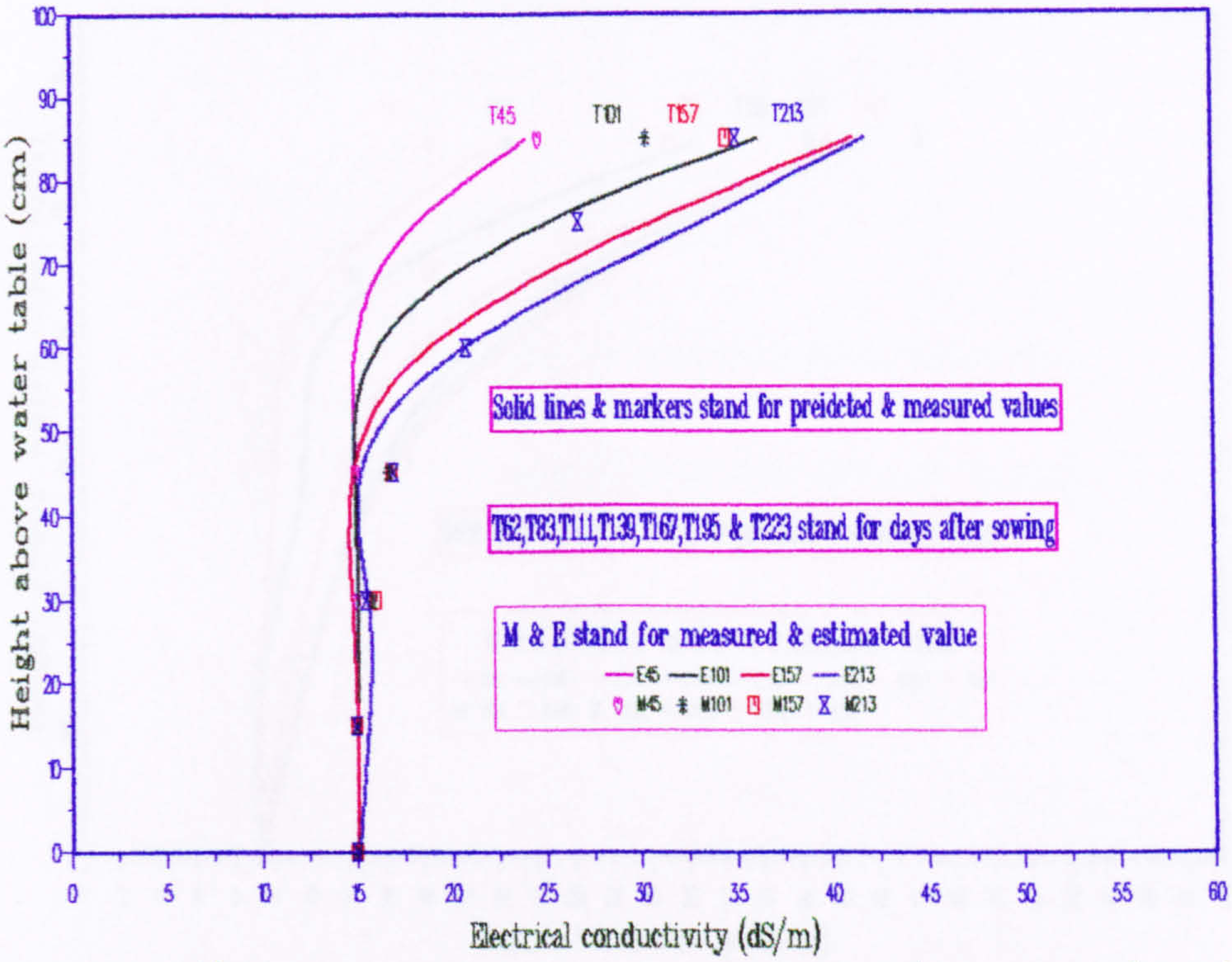


Fig. 6.3.1(c): Predicted & measured salt profile in 15.0 dS/m lysimeter (rye'93)

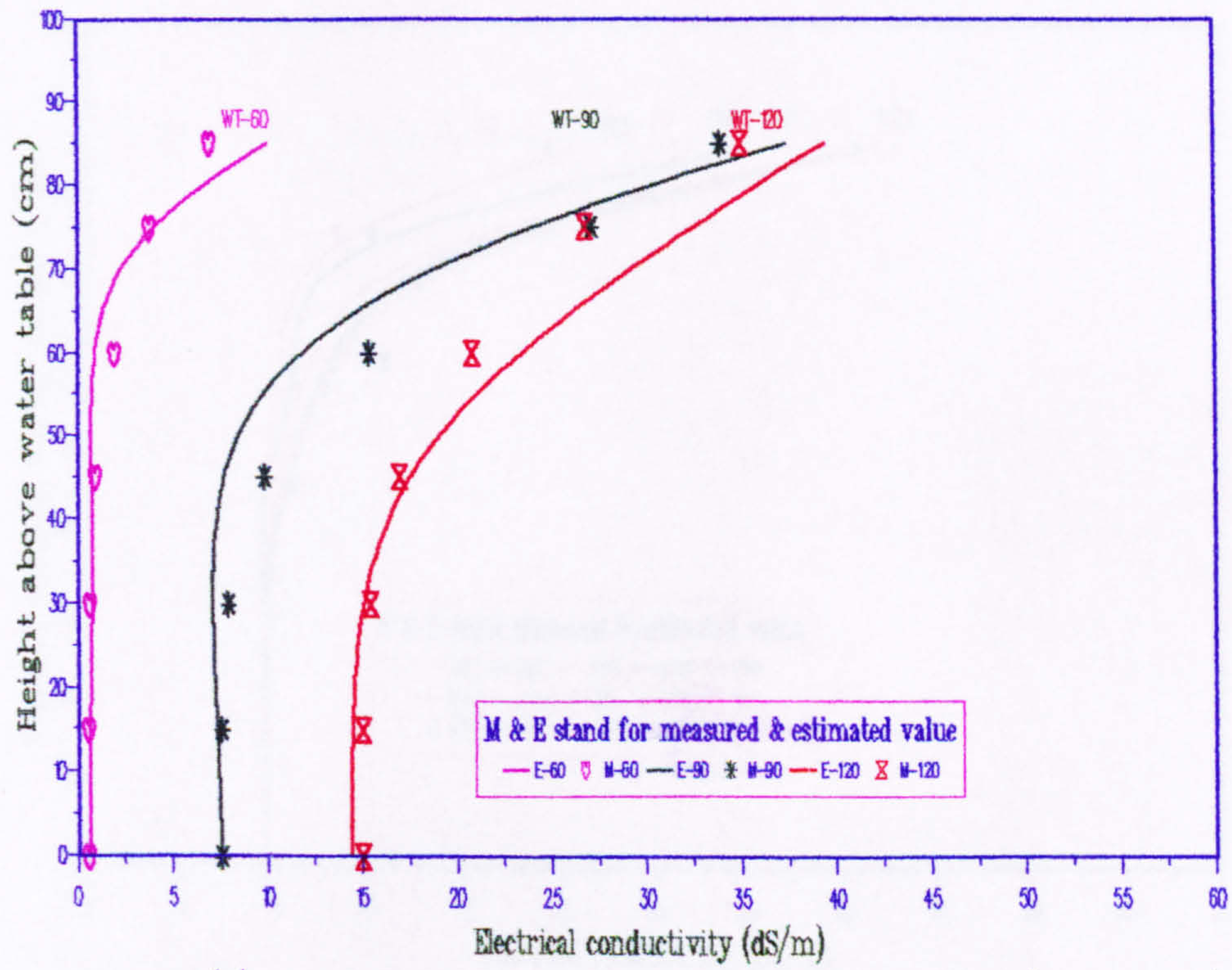


Fig. 6.3.1(d): Predicted & measured salt profile at harvest (ryegrass'93)

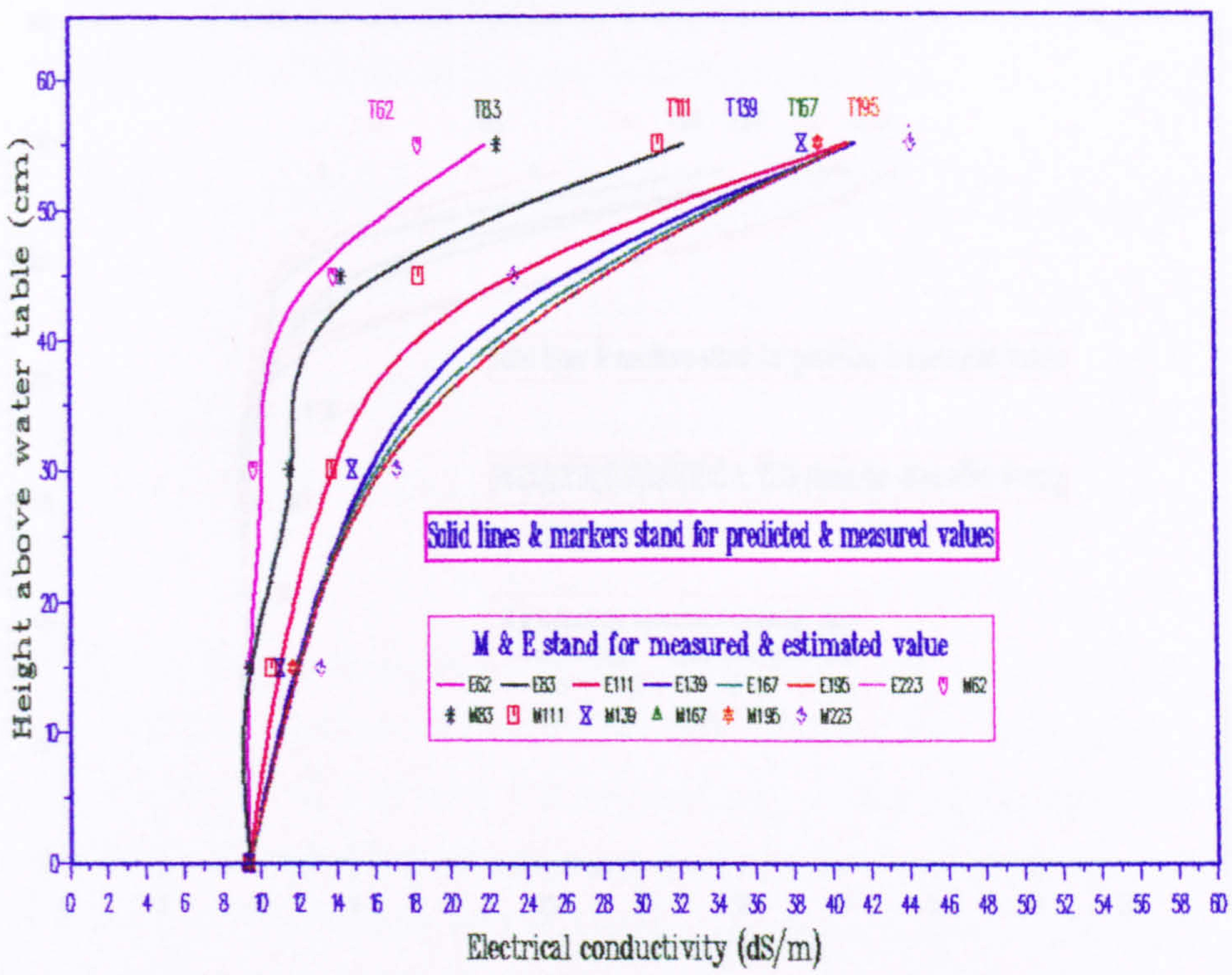


Fig. 6.3.2(a): Predicted & measured salt profile in WT-60 lysimeter (rye'92)

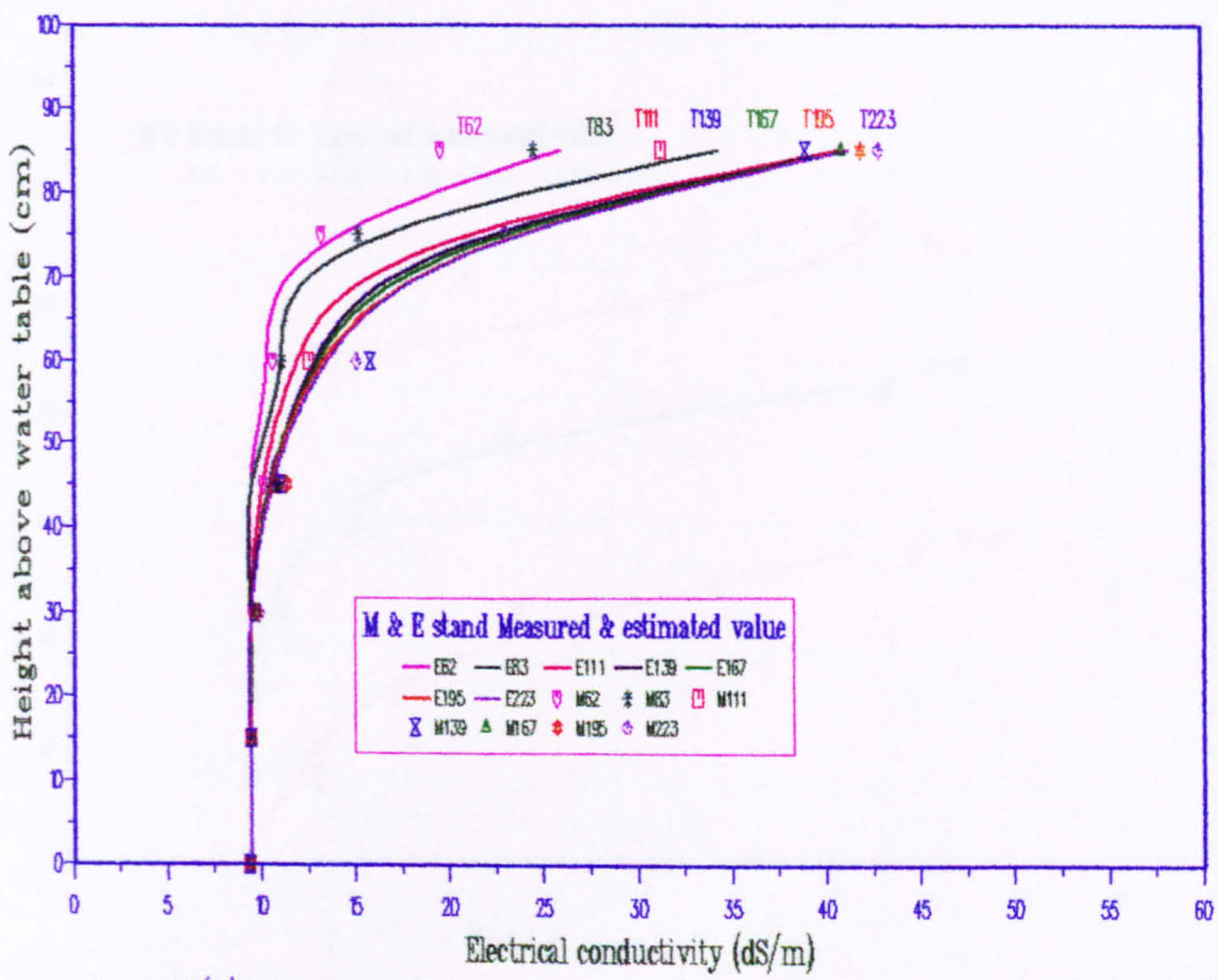


Fig. 6.3.2(b): Predicted & measured salt profile in WT-90 lysimeter (rye'92)

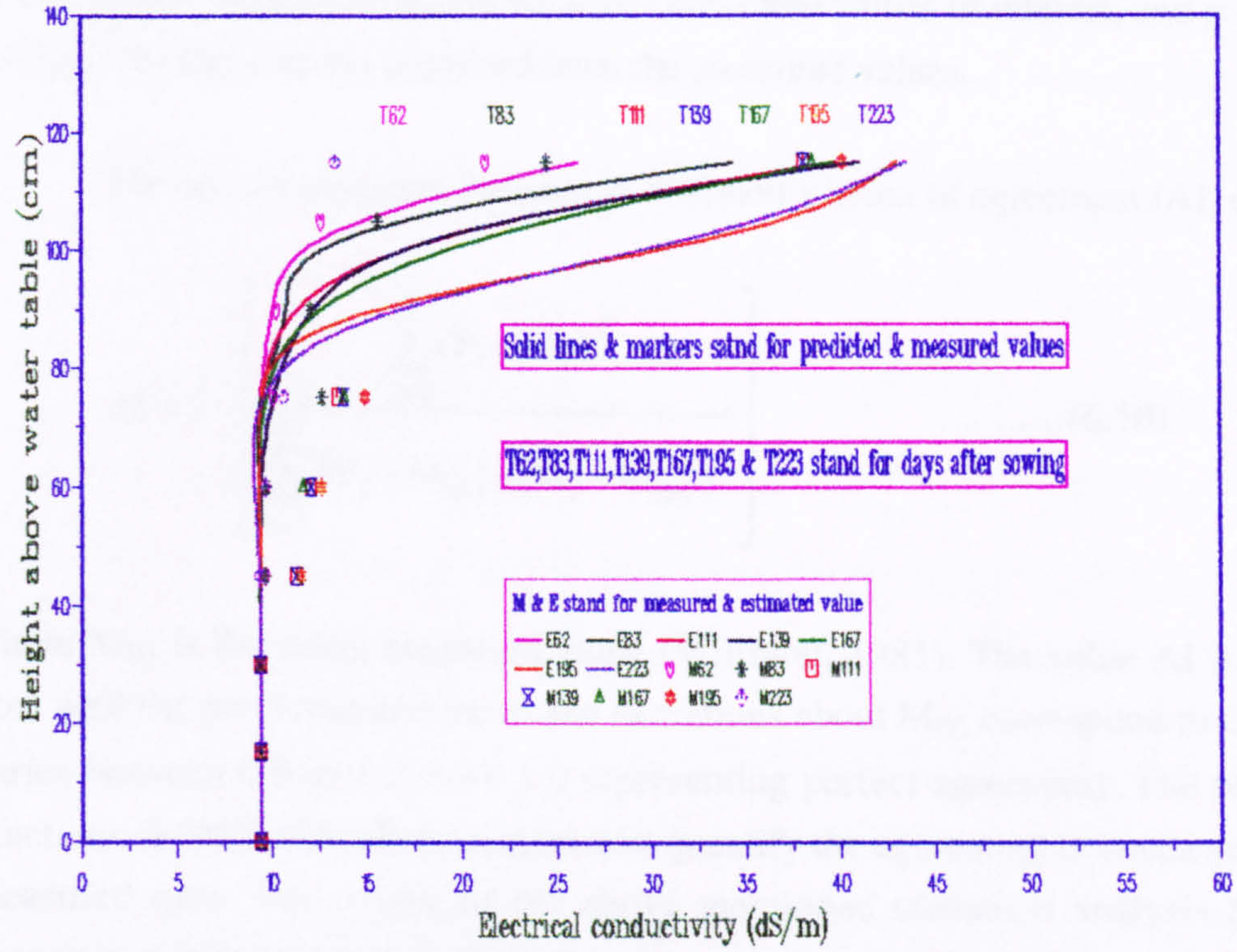


Fig. 6.3.2(c): Predicted & measured salt profile in WT-120 lysimeter (rye'92)

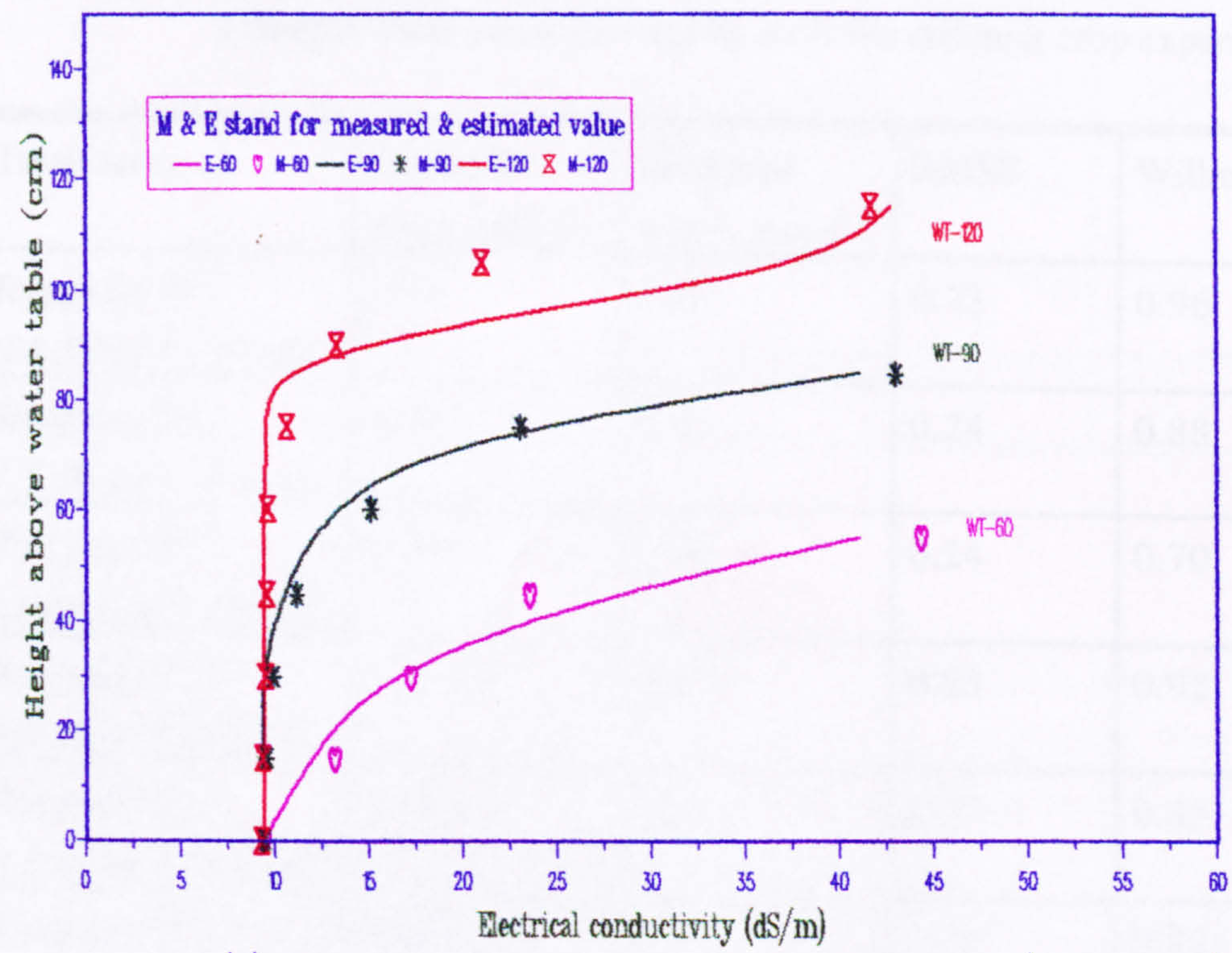


Fig. 6.3.2(d): Predicted & measured salt profile at harvest(ryegrass'92)

where P_i and M_i are the ith predicted and measured values of interest, and is a measure of average deviation of the predicted from the measured values.

The second objective function is Willmott's index of agreement (AI) expressed as:

$$AI = 1 - \left[\frac{\sum_{i=1}^{N} (P_i - M_i)^2}{\sum_{i=1}^{N} (|P_i - M_m| + |M_i - M_m|)^2} \right] \qquad \text{...........(6.10)}$$

where M_m is the mean measured value (Willmott, 1981). The value AI is an index of how well the predicted and measured deviations about M_m correspond to each other. It varies between 0.0 to 1.0, with 1.0 representing perfect agreement. The two objective functions (RMSE and AI) in conjunction quantify the agreement between simulated and measured data. The results of the above mentioned statistical analysis for different experiments are presented in Table 6.1.

Table 6.1: The results of agreement between simulated and measured (seasonal average) water extraction rate by roots for different crop experiments.

Treatments	Predicted mean, mm d^{-1}	Measured mean, mm d^{-1}	RMSE	Willmott's AI
Ryegrass'93 / 0.4 dS m^{-1} / 90 cm	1.09	1.03	0.23	0.96
Ryegrass'93 / 7.5 dS m^{-1} / 90 cm	0.92	0.81	0.24	0.88
Ryegrass'93 / 15.0 dS m^{-1} / 90 cm	0.39	0.30	0.24	0.70
Ryegrass'92 / 9.4 dS m^{-1} / 60 cm	0.63	0.59	0.25	0.92
Ryegrass'92 / 9.4 dS m^{-1} / 90 cm	0.59	0.61	0.27	0.83
Ryegrass'92 / 9.4 dS m^{-1} / 120 cm	0.56	0.55	0.26	0.89

The trend of variations of predicted water uptake from the measured values at 15 day intervals during the growing season is also shown in Table 6.2.

Table 6.2: Comparison of predicted & measured values water use (mm d^{-1}) for different treatments over different time periods in the cropping seasons.

Day	Ryegrass/0.4		Ryegrass/7.5		Ryegrass/15		Ryegrass/60		Ryegrass/90		Ryegrass/120	
	Meas.	Var.	Meas.	Var.	Meas.	Var.	Meas.	Var.	Meas.	Var.	Meas.	Var.
1-15	1.06	-0.11	0.22	-0.04	0.00	0.00	0.22	-0.05	0.22	-0.03	0.22	-0.06
16-30	1.23	0.02	0.86	0.05	0.73	-0.25	0.56	-0.08	0.61	-0.04	0.61	-0.04
31-45	1.61	-0.13	1.25	-0.39	0.42	-0.13	1.04	-0.35	0.80	-0.22	0.80	-0.23
46-60	1.55	-0.22	1.46	-0.41	0.60	-0.11	1.52	-0.26	1.42	-0.40	1.02	-0.17
61-75	1.69	0.03	1.07	0.06	0.67	-0.20	0.99	-0.04	0.90	0.00	0.88	-0.13
76-90	2.13	0.20	1.49	-0.21	0.72	-0.26	0.92	0.10	0.86	0.04	0.76	0.04
91-105	1.28	-0.16	0.86	0.16	0.57	-0.17	0.91	0.19	0.86	0.27	0.72	0.23
106-120	1.24	0.04	0.83	-0.03	0.31	0.11	0.72	-0.19	0.65	0.09	0.56	0.16
121-135	0.97	0.09	0.63	-0.13	0.23	0.23	0.62	-0.06	0.52	0.15	0.54	0.00
136-150	0.64	-0.17	0.48	-0.11	0.45	-0.10	0.33	0.17	0.36	0.23	0.42	0.03
151-165	0.59	-0.15	0.00		0.28	-0.12	0.24	0.08	0.27	0.18	0.21	0.20
165-180	0.52	-0.17	0.00		0.19	-0.04	0.30	0.01	0.37	-0.01	0.58	-0.13
180-195	0.43	-0.06	0.00		0.16	-0.02	0.22	0.00	0.29	-0.05	0.30	-0.03
195-210	0.32	-0.07	0.00		0.15	-0.04	0.17	0.02	0.19	0.03	0.19	0.05

Meas. = measured water uptake.

Var. = difference (Predicted-measured values).

From Table 6.1, the indices of agreements show that the higher the salinity, the higher the deviations, except the ryegrass'92 with 60 cm water table. This means that actual transpiration is less than the predicted. One reason may be that transpiration is limited by reducing stomatal aperture due to high stress. The other possible reason may be due to the accumulation of salts in the cells either by dehydration or some salt uptake from the soil and, thus, the added solute concentration might reduce the vapour pressure of cell water which in turn reduced the transpiration rate to some extent. The smaller root radius visually observed in the salinity treatments might also cause an increase in the hydraulic resistance of root water flow which was not considered in the model. Though the consideration of osmotic adjustment improved the root water uptake simulation in the later stages of the crop growing season, the exact degree of osmotic adjustment with the degree of salinity is not known. Apart from this, there could be many other micro-scale variations in fluid dynamics and biological function. However, the assumption of the plant wilting point as -1.5 MPa and the additive effect of matric & osmotic potential with respect to soil water can be considered acceptable for water management.

The present model may be better than the reported salinity models from the following point of view:

i) it generalizes the situation, not requiring different water uptake factors for different stages of crop growth (crop coefficient);

ii) it simplies the treatment of the water uptake reduction factor, to account for the matric and osmotic potentials differently, as a single stress factor;

iii) it is simple because the minimum number of variables have been used in this model.

Part 3: ANCILLARY EXPERIMENTS

Chapter: 7

HYDRAULIC PROPERTIES OF SOIL

7.1 Soil properties

7.1.1 Bulk density

The same three lysimeters (Hassan, 1990) with the same soils were used for this investigation. The bulk density of soils at various depths in each lysimeter were measured after harvesting the crops in each year. There were no appreciable difference found. Only the bulk density measured after harvesting the last experiment (ryegrass'93) was presented for comparison because the top 15 cm in each lysimeter was replaced (see section 4.2.1 for details). The bulk density was measured from undisturbed soil core samples and presented in Table 7.1.1.

Table 7.1.1: Bulk density of soil for three lysimeter treatments.

Lysimeters	Initial soil salinity (dS m^{-1}) / Water table depth (cm	Hassan expt. Bulk density (g cm^{-3})	Ryegrass'93 expt. Bulk density (g cm^{-3})
LYS-1	0.6 / 90	1.55 ± 0.07	1.547 ± 0.006
LYS-2	7.5 / 90	1.56 ± 0.05	1.540 ± 0.056
LYS-3	15.0/ 90	1.56 ± 0.05	1.537 ± 0.042

Table 7.1.1 shows that the compaction of soil in each lysimeter was fairly uniform. Assuming that the particle density of the soil is 2.65 g cm^{-3}, the proportion of space occupied by soil particles, from the relation of bulk density and particle density, is 0.582 ± 0.0021. In other words, the saturation water content should be 0.418 cm^3 (H_2O) cm^{-3}, in good agreement with the value of 0.415 cm^3 (H_2O) cm^{-3} reported by Hassan (1990) from the bulk density of 1.560 ± 0.006 g cm^{-3}.

7.1.2 Soil moisture characteristic

Hassan (1990) used a semilog model to express the soil moisture characteristic equations (see section 3.1.4) for his experimental soil and in the present study the same equations have been used. However, it was necessary to evaluate how far the newly replaced soils and salinity have affected soil physical properties. The average volumetric moisture content, θ, of the undisturbed but resampled (sampling core from lysimeters was 7.5 cm in diameter and 5 cm in height and then resampling was done with a core size of 4 cm in diameter and 2.5 cm in height to fit into the Haine's funnel) cores was determined. Each of the resampling size was 4.0 cm in diameter and 2.5 cm in height. As the collected samples were fairly dry, the appropriate resampling could not be done and consequently, the bulk density became somewhat lower. The bulk density of resampled soil used for measuring soil moisture characteristic are presented in Table 7.1.2.

Table 7.1.2: Bulk density of resampled soil for three lysimeter treatments.

Lysimeters	Water table depth (cm)	Initial soil salinity ($dS\ m^{-1}$)	Bulk density ($g\ cm^{-3}$)
LYS-1	90	0.6	1.425 ± 0.063
LYS-2	90	7.5	1.430 ± 0.007
LYS-3	90	15.0	1.434 ± 0.045

The moisture content, θ as a function of Ψ for three different treatment are presented in Fig. 7.1.1. The empirical two-segment semilog models fitted the relationship water content and suction. The Figure shows that there is no significant difference between the soil moisture characteristic between the present and the Hassan(1990) and also between the treatments. However, the slight variation between Hassan and the present is due to the bulk density difference due to resampling. When tensiometers failed to record the soil moisture suction, moisture content was inferred from the calibration of resistance data recorded by gypsum blocks (Fig. 7.1.2).The parameter values of soil moisture characteristic equations (Equation 3.9 and 3.10 in chapter 3) are also presented in Table 7.1.3.

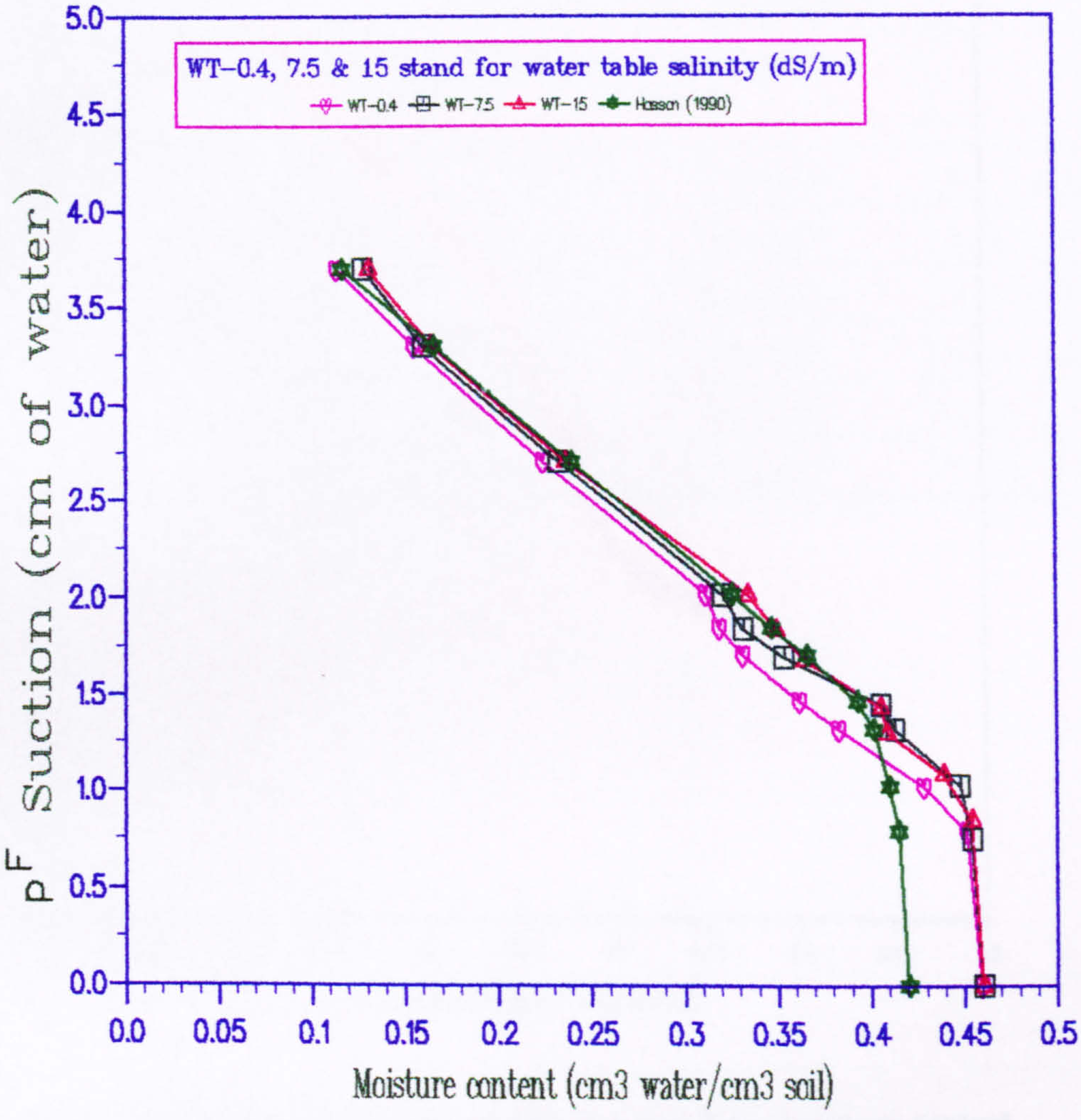


Fig. 7.1.1: Soil moisture characteristic curves

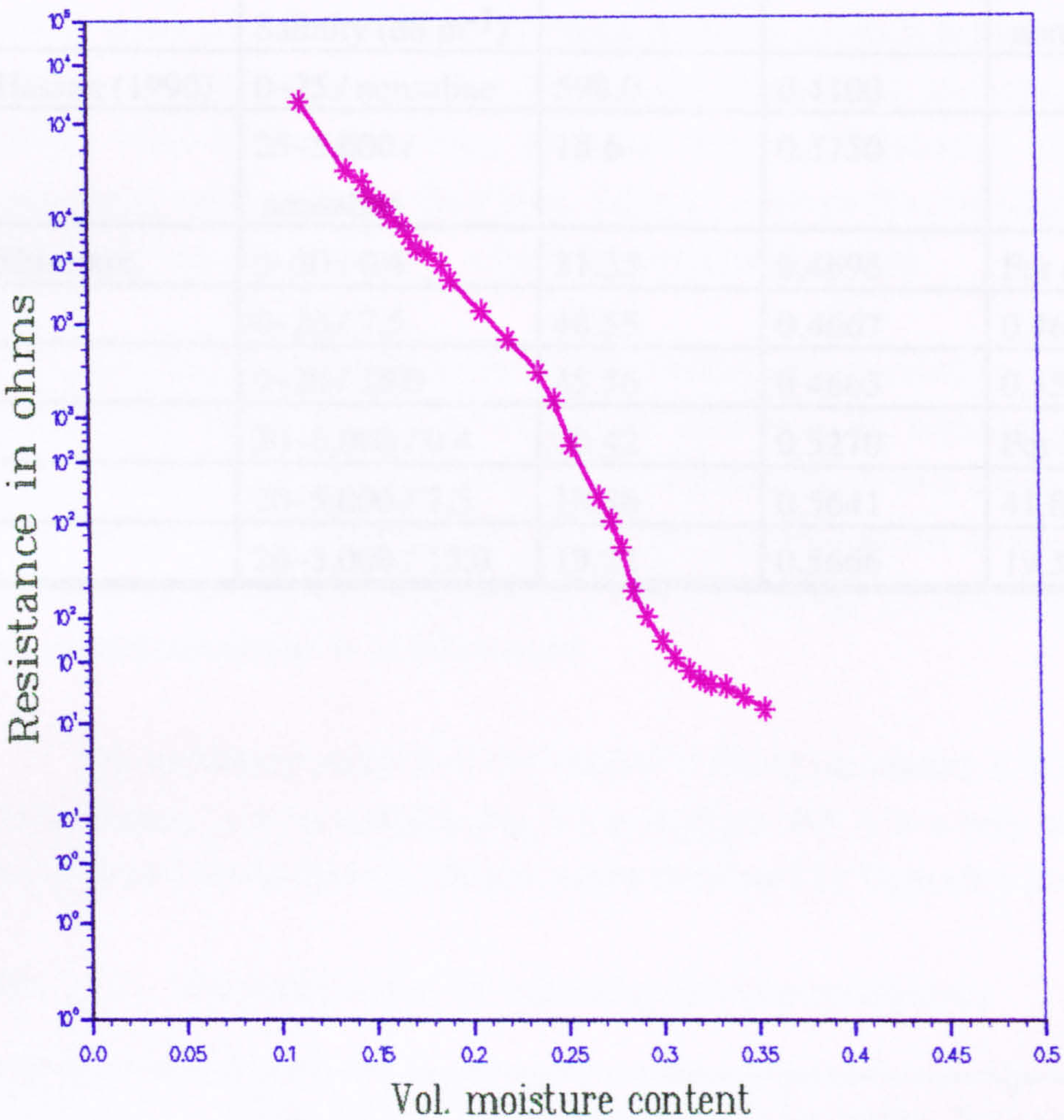


Fig. 7.12: Calibration of resistance block (gypsum) for moisture content

Table 7.1.3: Parameter values of soil moisture characteristic equations

Parameters determined by	Suction range (cm) / Salinity (dS m^{-1})	a value	b value	Std.dev.betn. saline & nonsaline
Hassan (1990)	0~25 / nonsaline	598.0	0.4100	
	25~5,600 / nonsaline	18.6	0.5750	
This expt.	0~20 / 0.4	31.35	0.4696	For a values:
	0~20 / 7.5	48.55	0.4667	0.4675 ± 0.018
	0~20 / 15.0	45.56	0.4663	0.5526 ± 0.022
	20~5,000 / 0.4	20.42	0.5270	For b values
	20~5,000 / 7.5	18.86	0.5641	41.82 ± 9.20
	20~5,000 / 15.0	19.22	0.5666	19.5± 0.817

7.1.3 Hydraulic conductivity

Hydraulic conductivity was determined by two field methods, (i) Soil Water Depletion and (ii) Drainage Flux. Derived values from the measured crop water use data and the fitted results by Gardner's equation are presented as a function of suction for different water table treatments by soil water depletion methods in Figs. 7.1.3(a),(b),(c) & (d). Fig. 7.1.3(a) & (b) represent the hydraulic conductivity at constant salinity and different water table depths. Figs. show that there is a reasonable matching of the experimental values with the fitted lines. Table 7.1.4 shows that the saturated hydraulic conductivity was almost same for saline and nonsaline treatments.

Hydraulic conductivity determined by Drainage flux method are presented in Fig. 7.1.4(a),(b),(c) & (d). Figures 7.1.4(a) & (b) show that, the higher the height above water table, the higher the rate of hydraulic conductivity up to a certain range of low suction which then dropped sharply. Fig. 7.1.4(c) shows that the difference in salinity did not affect the hydraulic conductivity. Note that, within 15 to 25 days draining of water ceased completely in all experiments.

The sensitivity analysis of the Gardner's fitting parameters a, b & n (Equation 3.14 in chapter 3) is presented in Fig. 7.1.5. It shows that 'n' is a very sensitive and 'b' is a very insensitive parameter. Similar results was found by Vereecken et al. (1990).

Table 7.1.4: Parameter values for estimating hydraulic conductivity

Lysimeters / salinity, dS m^{-1} / WT depth (cm)	K_{sat} (mm d^{-1}) for Soil water depletion	K_{sat} (mm d^{-1}) Drainage flux	n value for Soil water depletion	n value for Drainage flux
L1 / 0.4 / 90	17.61	18.25	1.12	2.10
L2 / 7.5 / 90	17.84	18.04	1.19	2.11
L3 / 15.0 / 90	17.67	17.36	1.43	2.14
L1 / 9.4 / 60	17.48	21.36	1.25	2.16
L2 / 9.4 / 90	17.60	19.53	1.21	2.13
L3 / 9.4 / 120	17.56	21.15	1.23	2.02
L1 / 9.4 / 60	17.83	18.81	1.32	2.25
L2 / 9.4 / 90	18.25	20.05	1.22	2.12
L3 / 9.4 / 120	17.53	19.25	1.20	2.01

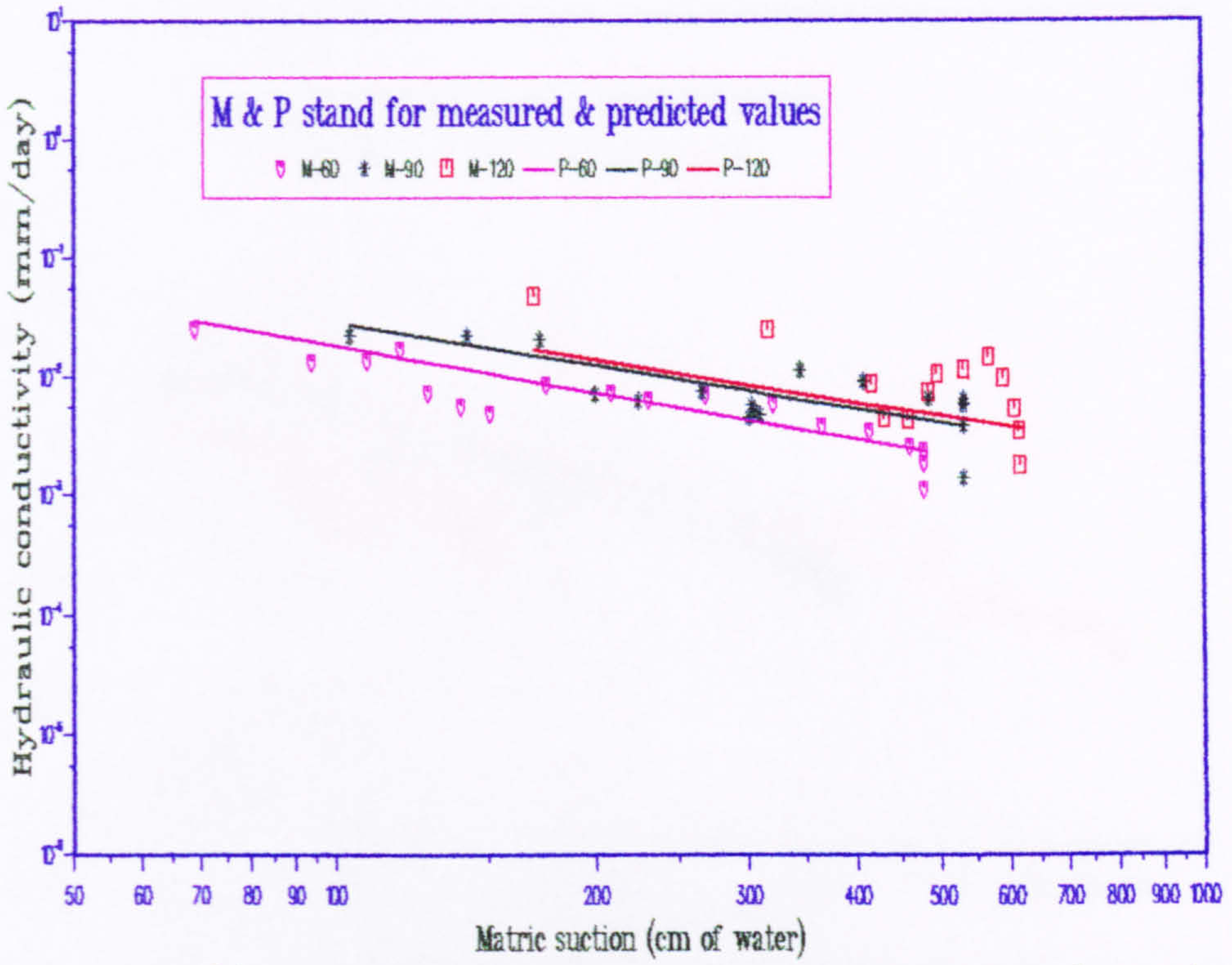


Fig. 7.1.3(a): Hydraulic conductivity by soil water depletion (lettuce)

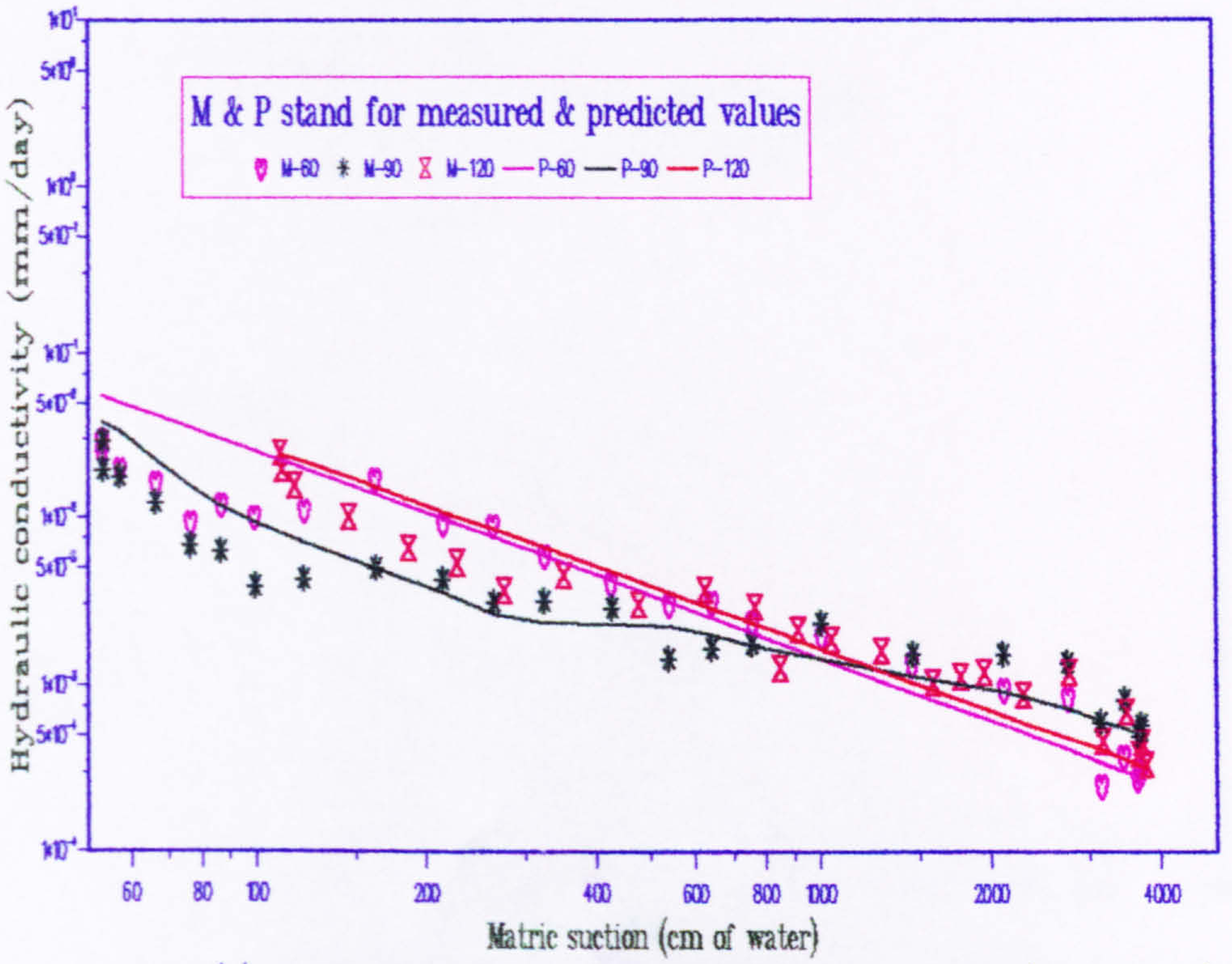


Fig. 7.1.3(b): Hydraulic conductivity by soil water depletion (ryegrass'92)

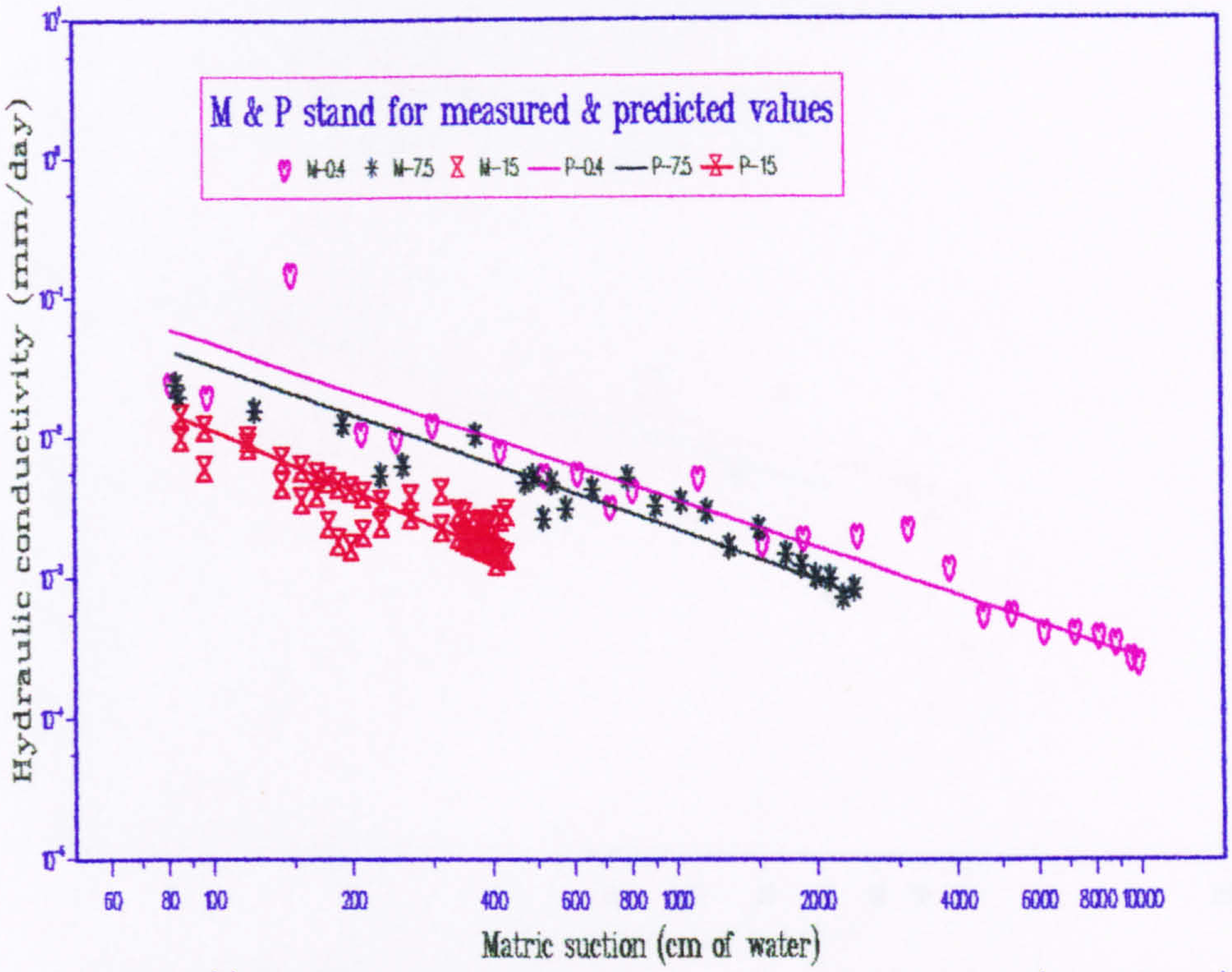


Fig. 7.13(c): Hydraulic conductivity by soil water depletion (ryegrass'93)

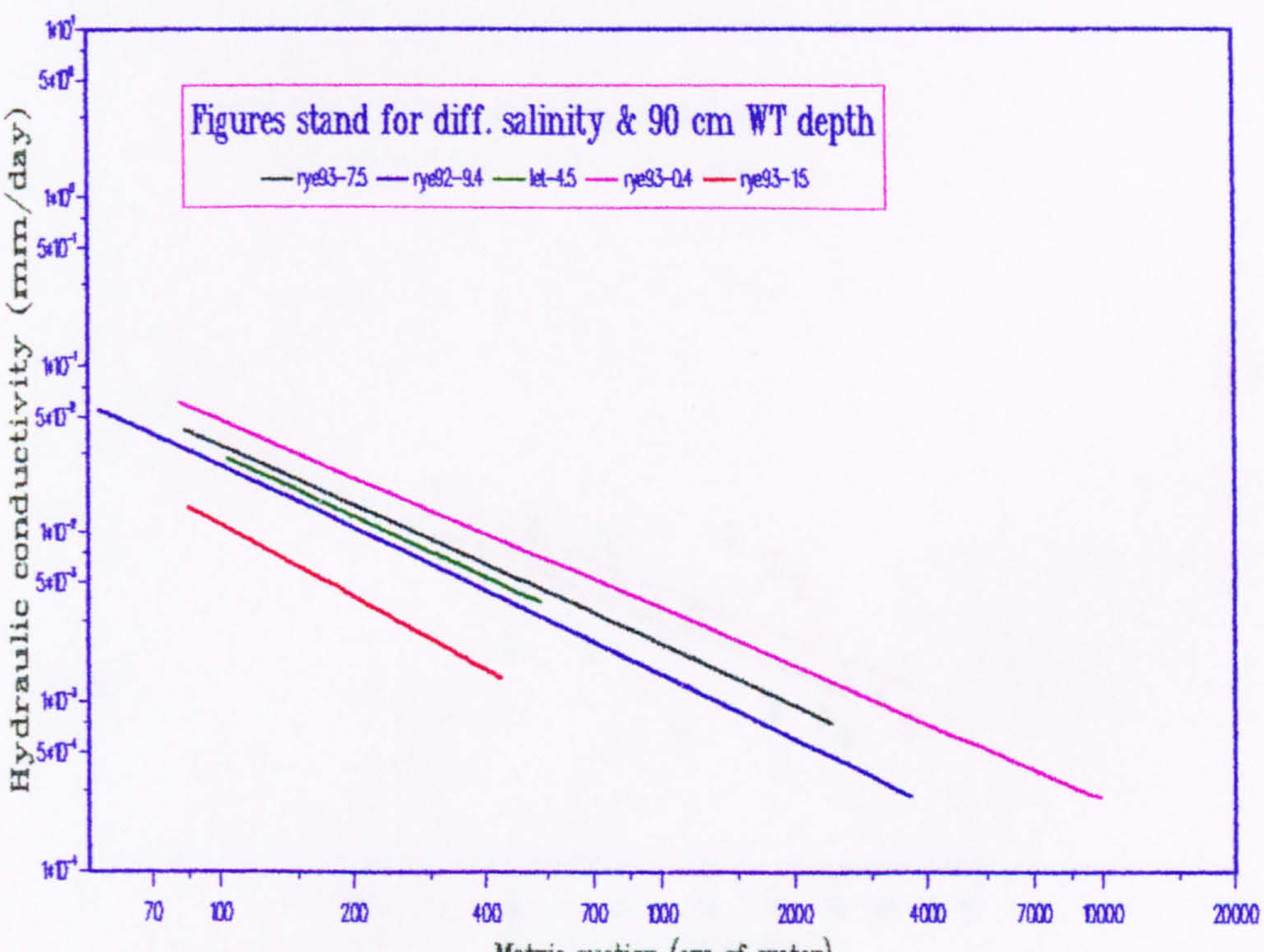


Fig. 7.13(d): Comparison of hydraulic conductivity for diff. treatments

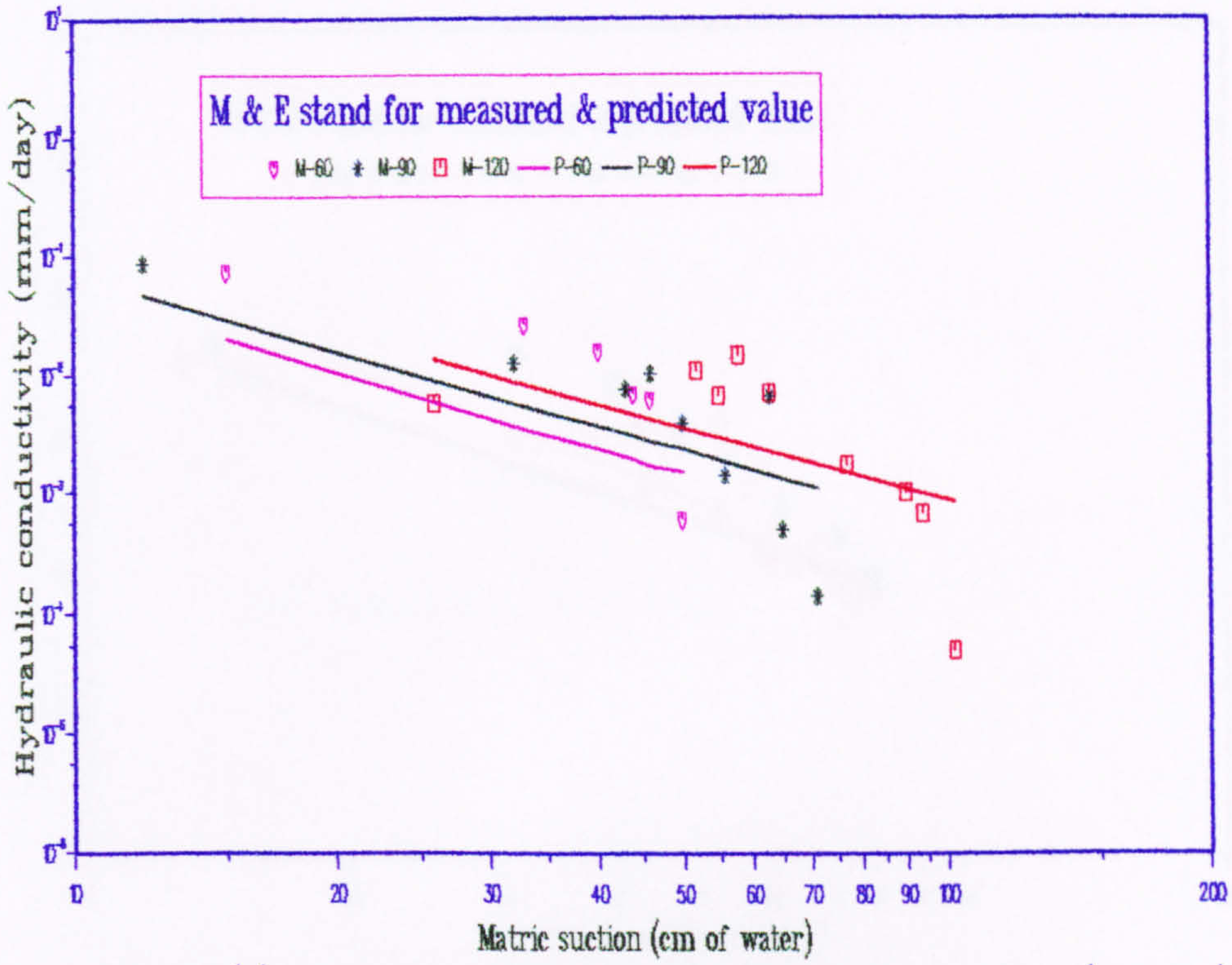


Fig. 7.1.4(a): Hydraulic conductivity by drainage flux (lettuce)

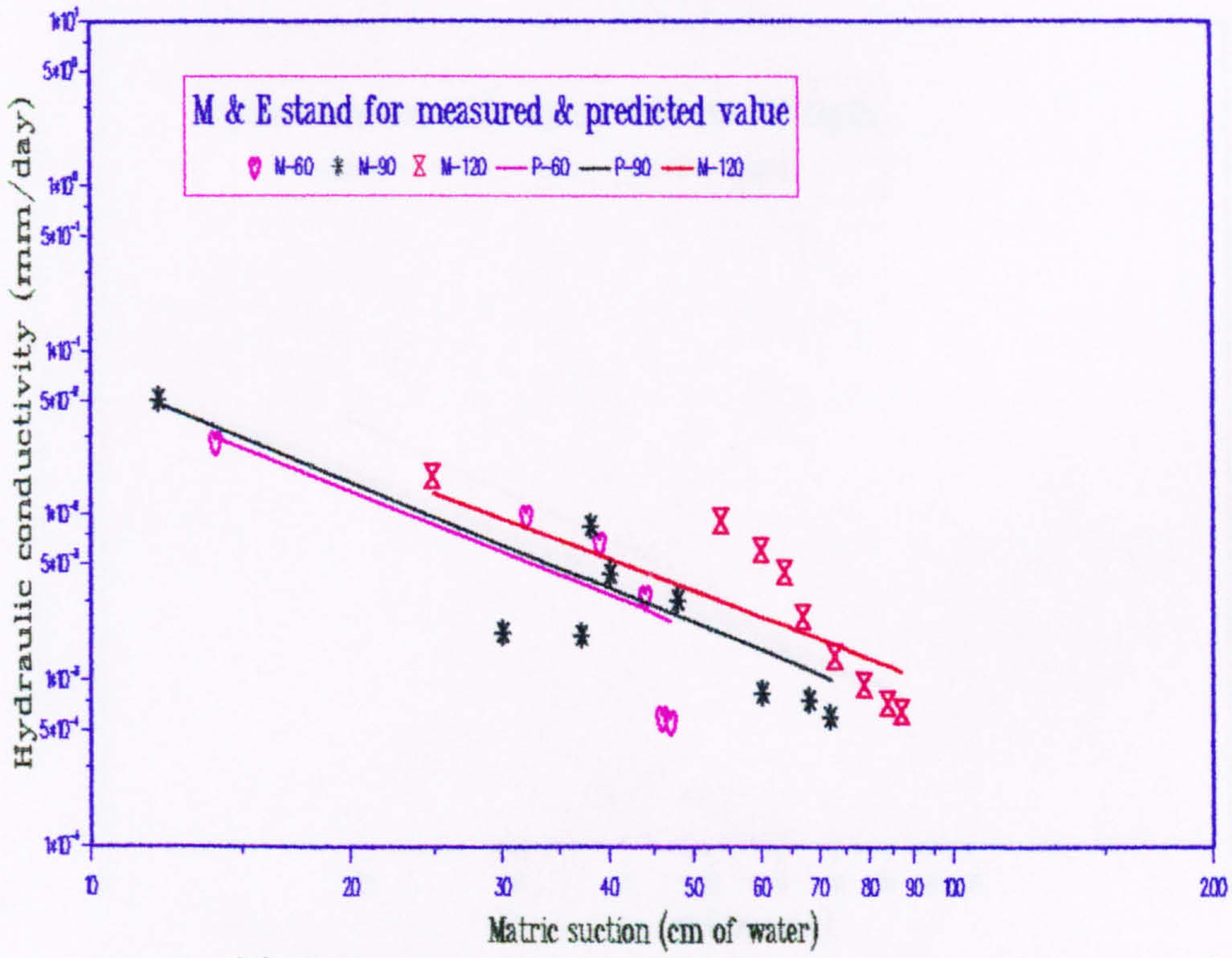


Fig. 7.1.4(b): Hydraulic conductivity by drainage flux (ryegrass'92)

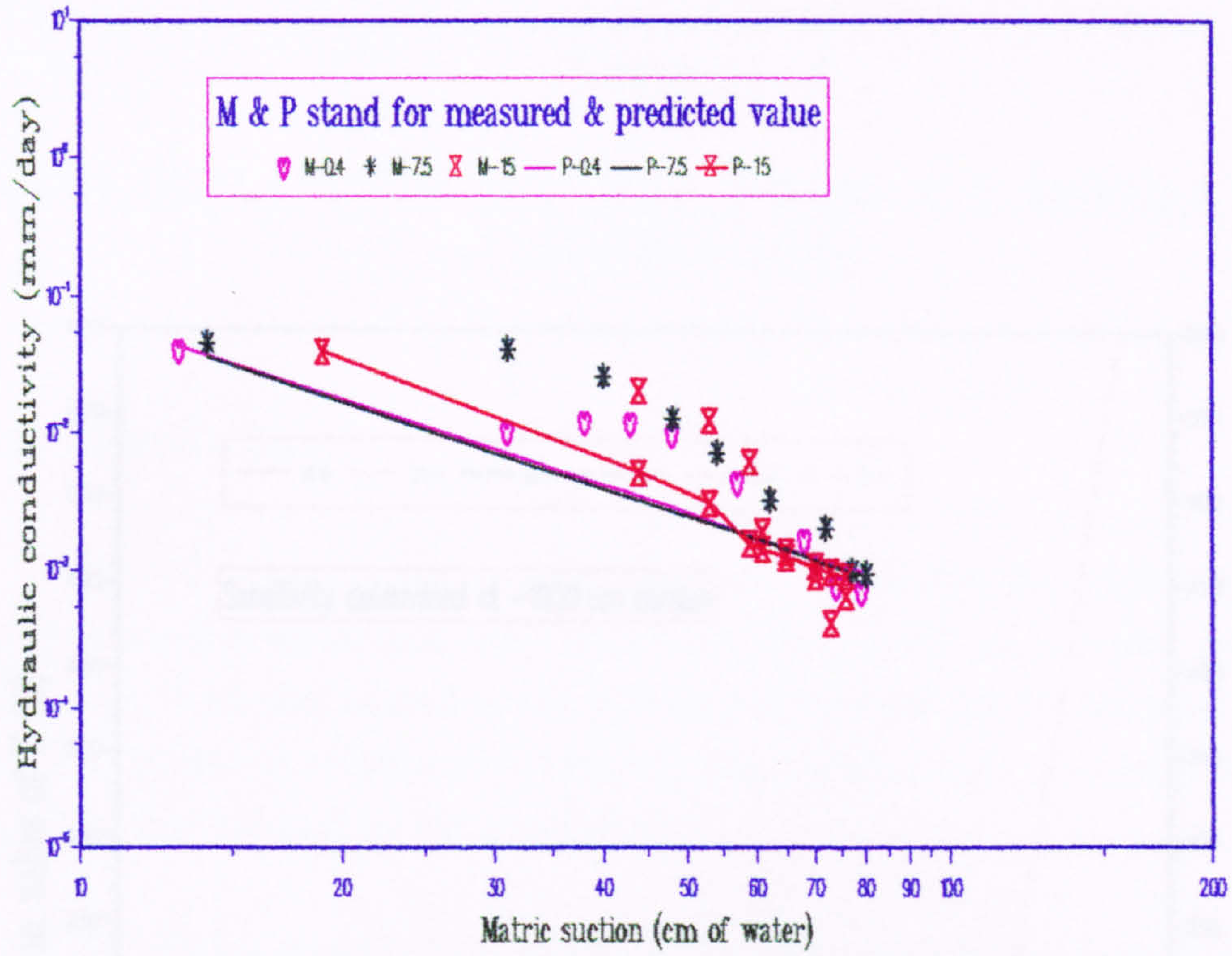


Fig. 7.1.4(c): Hydraulic conductivity by drainage flux (ryegrass'93)

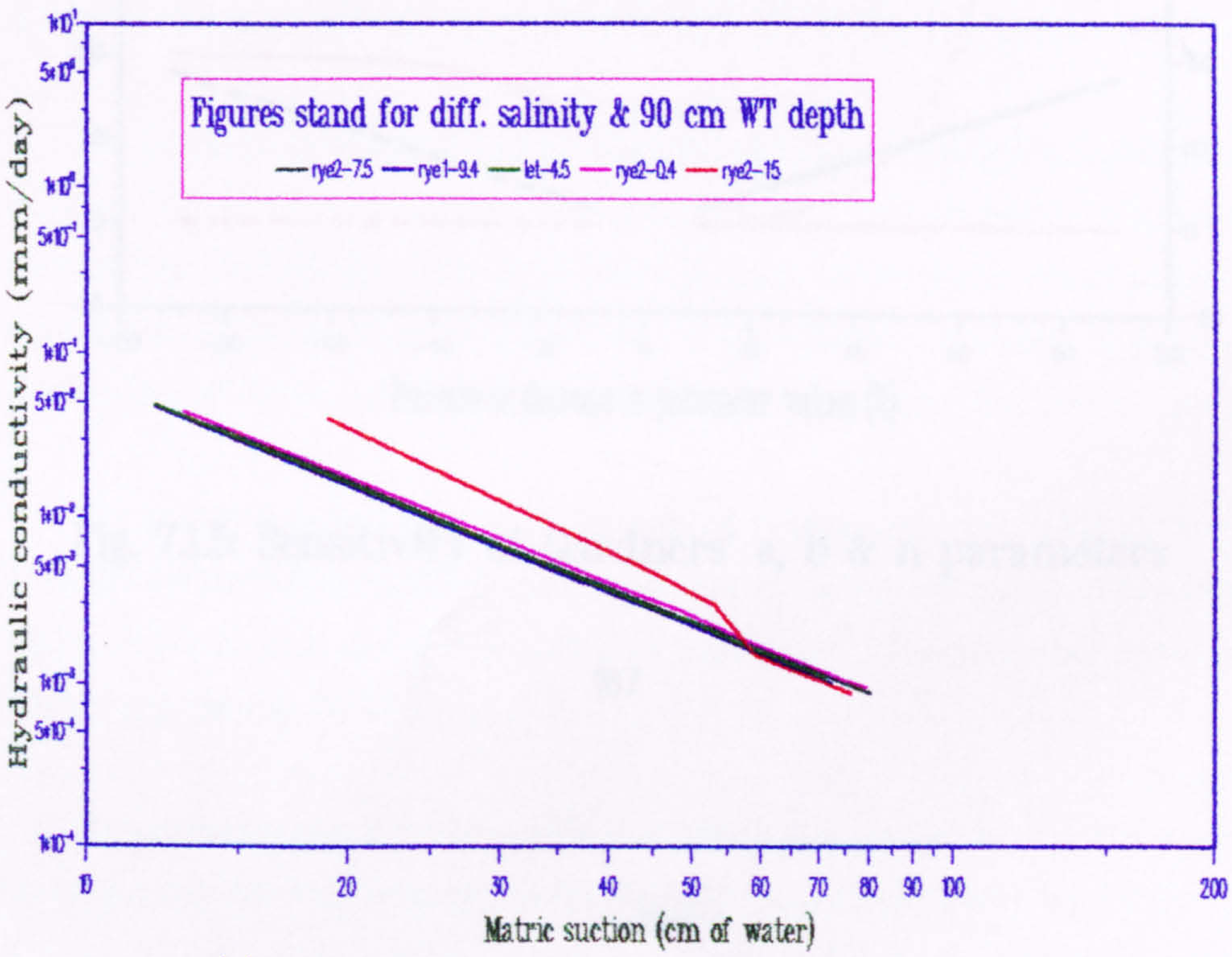


Fig. 7.1.4(d): Comparison of hydraulic conductivity for diff. treatments

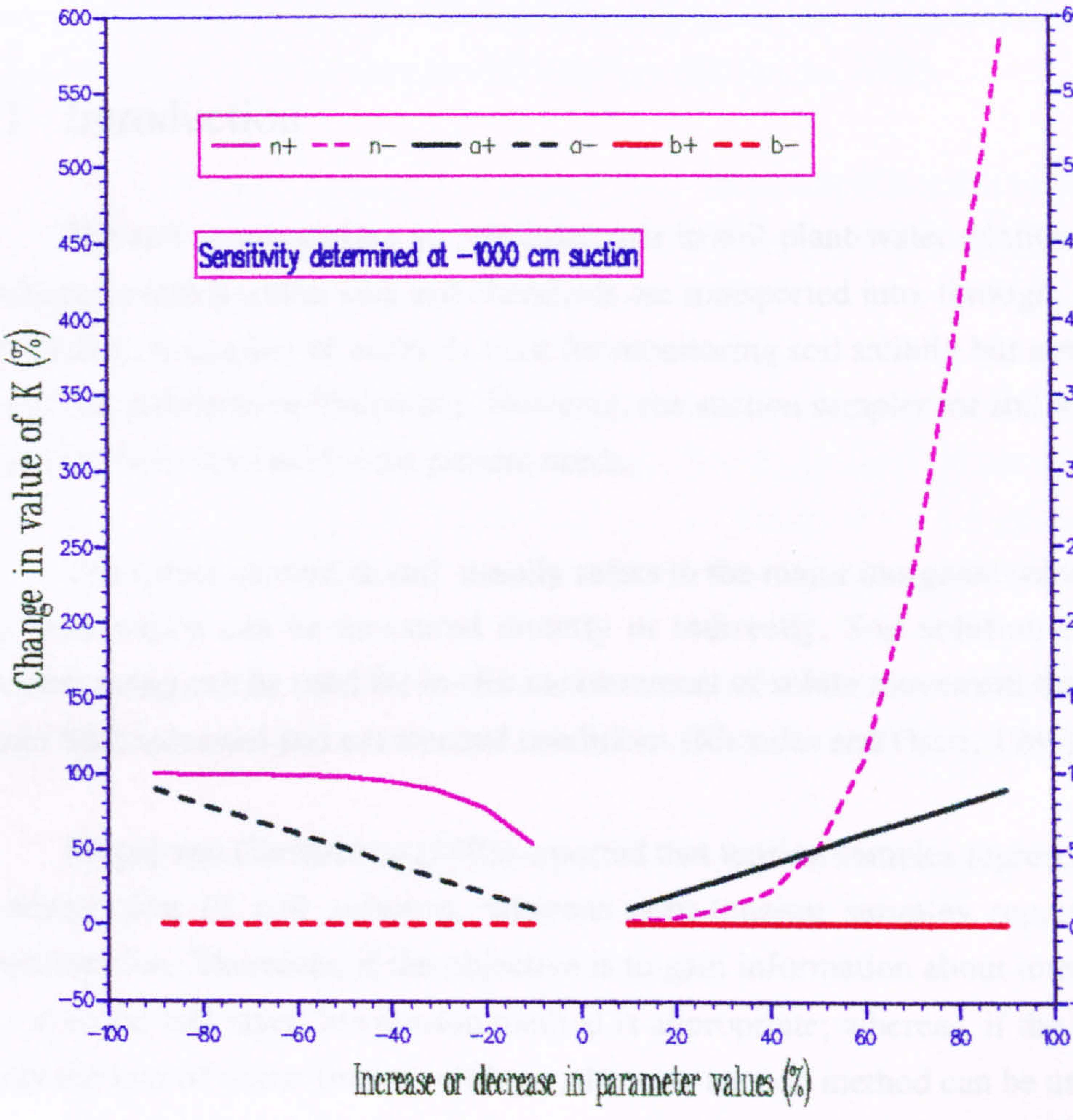


Fig. 7.1.5: Sensitivity of Gardners' a, b & n parameters

Chapter: 8

SOIL SOLUTION EXTRACTION: A LABORATORY APPROACH.

8.1 Introduction

The soil solution plays an important role in soil-plant-water relationships. It is the medium in which solute ions and chemicals are transported into, through, and out of the soil matrix. A number of methods exist for monitoring soil salinity but none of them yet is without problems or limitations. However, the suction sampler for soil water sampling insitu has been selected for the present needs.

The solute content in soil usually refers to the major inorganic solutes retained in the soils which can be measured directly or indirectly. Soil solution extracted by a vacuum pump can be used for in-situ measurement of solute movement through the soils under both saturated and unsaturated conditions (Rhoades and Oster, 1986).

Magid and Christensen (1993) reported that tension samples represent the resident concentration of soil solution, whereas zero-tension samples represent the flux concentration. Therefore, if the objective is to gain information about internal processes in a specific soil layer, the tension method is appropriate; whereas, if the objective is to study the loss of solute from a soil layer, the zero-tension method can be used.

The possible errors related to the use of porous ceramic extractor are: the rate of soil water movement; the chemical composition of the cup; sampler intake rate, plugging, sampler depth & size and the problem of adsorption of ions by the ceramic cup itself (Hansen and Harris, 1975).

To overcome some of the possible errors during extraction of soil solution through the ceramic cup the following precautions can be taken (Debyle et al., 1988) :

i) to flush new ceramic cup with at least 1 N HCl followed by distilled water;
ii) to maintain a uniform suction, vacuum pressure should not much exceed the tension at which the soil solution is being held;
iii) to check all new samplers for intake variation; and,

iv) to check periodically older samplers in the field for changes of intake rates and, if plugging occurs, the sampler should be removed and the cups flushed.

The sampler is a finite volume of space in which air pressure changes cause the water level to rise within. This change in air pressure can be a dominant factor that controls the transient flow of water into the sampler i.e. the volume of the sample and the radius of influence of the sampler (Narasimhan and Dreiss, 1986).

The radius of influence of a suction unit, which extracts water continuously, can be several tens of cm and care should be taken in order to obtain reliable data (Van Der Ploeg and Beese, 1977).

Grossmann and Udluft (1991) reported that the advantages of suction probes are their relative simple installation and the negligible disturbance of the soil profile. The water flux and gas exchange in the soil are not hindered by the probes. Moreover, continuous sampling is possible, if necessary, at different depths within the same profile.

The above citations imply that the objective functions can be maximized by adopting some operational procedures and precautions needed for vacuum extraction.

The objective of this experiment was to determine the radius of influence of unsaturated soil solution flow during extraction by vacuum pump and also, to test the suitability of the technique for use in the lysimeters.

8.2 Materials and methods

8.2.1 Extraction device

The extraction device assembly consisted of four functional units : the suction cup assembly, the sampling bottle, the suction manifold and the vacuum pump (shown in Fig. 8.1). The ceramic cup was rigidly fixed to one end of a 10 mm i.d. PVC tube and a rubber stopper was fitted in the other end of the tube. Another piece of 1 mm i.d. tube was inserted through the rubber stopper, so that one end of this tube reached near the bottom of the ceramic cup while the other end was fitted to a flow control valve and from where another piece of 1 mm i.d. tube extended to a sampling bottle. A 4 mm I.D transparent PVC tube was fixed to the sampling bottle with the aid of a reducer. One stopcock assembly was equipped in the flow line between the reducer and the T-connector. The T-connector was provided for connecting a series of secondary flow lines

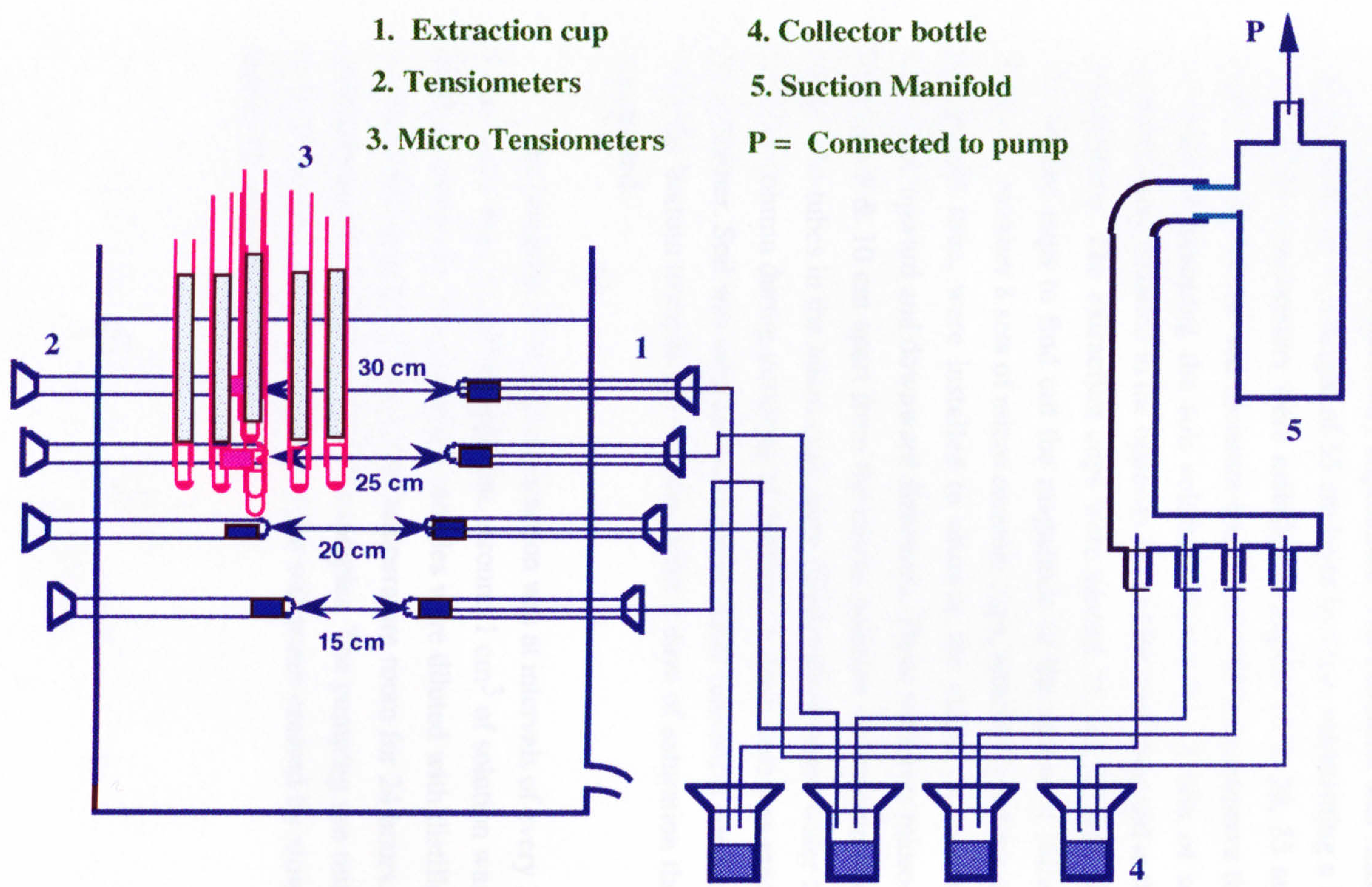


FIG. 8.1: SCHEMATIC OF SOIL-SOLUTION EXTRACTION DEVICE.

to a primary flow line to operate different suction points from a single pump. A suction manifold was attached between the primary flow line and the vacuum pump.

8.2.2 Experiment on soil solution extraction

A PVC container, 50 in cm diameter and 100 cm long, and closed at the bottom was selected for this laboratory experiment. The container was filled with Rivington soil (sandy loam) up to a height of 55 cm layer by layer maintaining a bulk density of 1.5 g cm^{-3}. Four tensiometers were installed at depths of 5, 20, 35 and 50 cm in the soil profile to monitor the soil moisture content as well as to observe the effect on radius of influence of pumping the soil solution. Similarly , 4 sets of soil water extraction assembly were installed in the opposite wall of the container and at the same depths as the tensiometers. The extraction cups were placed 15, 20, 25 and 30 cm apart from the tensiometer cups to find out the magnitude of the radius of influence in the forward direction. Another 8 sets of micro ceramic cups, attached to a long open-ended 2 mm i.d. transparent tube, were installed to observe the radius of influence in the forward, backward, upward and downward directions. There were two micro-tensiometers in each direction 5 & 10 cm apart from the centre position of the extraction cup at 20 cm soil depth. The tubes in the micro cups were filled with de-aired water to see the draw-down of water column during pumping of solution. A drain hole was made near the bottom of the container. Soil was saturated with saline water (electrical conductivity of 9.4 dS m^{-1}) from the bottom towards the surface. After 3 days of saturation the soil-water draining was started.

The pumping for solution extraction was at intervals of every 30 days and a total of 6 runs were done for this experiment. Around 1 cm^3 of solution was collected from each depth in every run. The collected samples were diluted with distilled water at a ratio of 1:10 and then kept for in a constant temperature room for 24 hours. Hence the electrical conductivity was measured from the samples. The pumping run time ranged from 2 min to 60 h depending on the tension of the soil water created by slow rate of evaporation during the period around 150 days.

8.3 Results and discussion

The radius of influence of soil solution flow, determined by observing the pressure drop or water level in the tensiometers placed around and away from the extraction cups during pumping, is presented in Table 8.1. The radius of influence was found to be 10 cm in the forward direction and 5 cm in the both upward & downward directions. No radius of influence was found in the backward direction. However, the radius of influence depends on the applied vacuum pressure.

It was also observed that the pressure drop in the vicinity of the extraction cups came to equilibrium with the surrounding soil-water pressure 48 hours after stopping pumping.

The measured and estimated (based on convective transport) values of salt development in the soil profiles are presented in Fig. 8.2. It shows that there is good agreement between the measured and the estimated results, except below 50 cm suction developed in the top 0~5 cm depth. Fig. 8.2 indicates that some salt might have been carried downward with the drainage flux, but after equilibrium of soil water no diffusion of salt was found.

			Pressure/water level drop in observation tensiometers.					
Soil tension (cm)	Applied vacuum (cm)	Pumping time (hr)	Separation distance (cm)					
			5	10	15	20	25	30
0-30	300	0.00-0.08	No	No	No	No	No	No
1-46	300	0.03-0.25	Yes	No	No	No	No	No
20-59	400	0.13-0.34	Yes	No	No	No	No	No
45-90	500	6.0--36.0	Yes	No	No	No	No	No
70-115	600	18.0--48.0	Yes	Yes	No	No	No	No
90-130	600	60.0	Yes	Yes	No	No	No	No

Table 8.1: Radius of influence of soil solution flow during extraction.

			Pressure/water level drop in the tensiometers around the extraction cups.					
Soil tension (cm)	Applied vacuum (cm)	Pumping time (hr)	Horizontally backward dist. (cm)		Vertically upward distance (cm).		Vertically downward distance (cm).	
			5	10	5	10	5	10
0-30	300	0.00-0.08	No	No	No	No	No	No
1-46	300	0.03-0.25	No	No	No	No	No	No
20-59	400	0.13-0.34	No	No	Yes	No	No	N0
45-90	500	6.0--36.0	No	No	Yes	No	No	Yes
70-115	600	18.0--48.0	No	No	Yes	No	No	Yes
90-130	600	60.0	No	No	Yes	No	No	Yes

Table: 8.1 (continued)

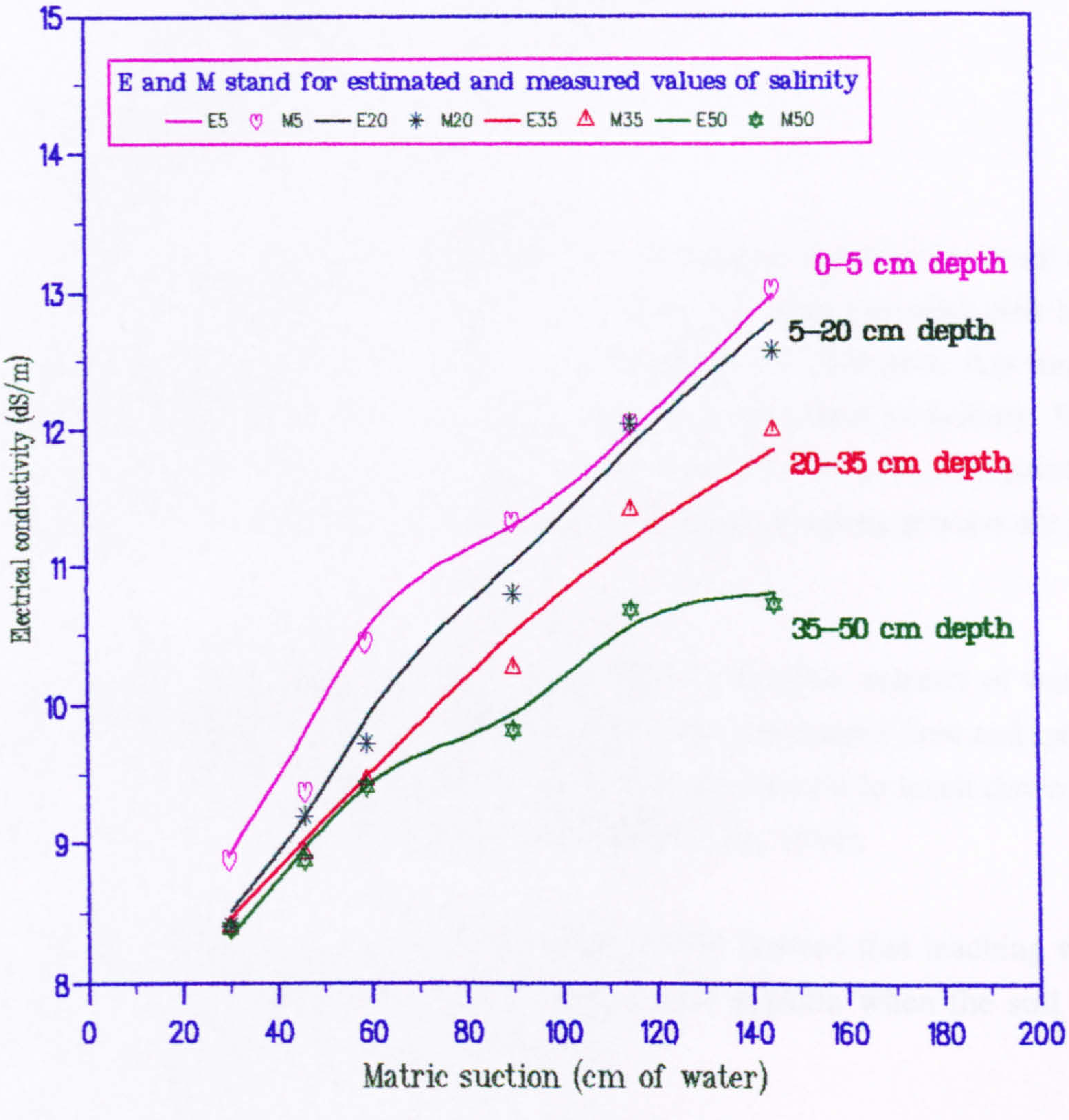


Fig. 8.2: Estimated and measured salinities at different soil depths.

Chapter: 9

SALT LEACHING AND RESALINIZATION OF SOILS.

9.1 Introduction

Leaching of salts is often necessary when soil salinity exceeds the permissible limit for crop initiation, or crop development. This situation is more prevalent after harvesting a crop with saline water or saline water table management. Therefore, it is important to know how much water is necessary to leach down what extent of salinity. Moreover, when there is a business of crop production with saline water, there is no point or scope of leaching all the salts from the soils. Some views of the leaching process are presented below.

The efficiency of salt leaching is controlled by the flow velocity of water during leaching. Intermittent application of water is better than continuous flow and sprinkling is better than ponding in terms of total volume of water needed to leach down the same degree of salinity (Nielsen and Biggar, 1961; Miller et al., 1965).

On the other hand, Abrol and Bhumbla (1973) showed that leaching was better accomplished by continuous ponding coupled with gypsum when the soil had poor permeability, because of structural deterioration.

Heilman et al (1968) showed that soil salinities of 17.5 and 19.7 dS m^{-1} leached down to 14.3 and 14.7 dS m^{-1} from bare fallow soils and bare fallow soils with gypsum treatments, respectively, with a rainfall of 386 mm. The soil type was Raymondville (Texas) clay loam above highly saline water table. They concluded that gypsum application did not significantly contribute to the leaching of salts from soils.

Tanton et al. (1990) demonstrated that soil salinity was leached down from 13.2 to 2.2 dS m^{-1} with a total flow of 1530 mm water over a period of 28 days in a clay soil restructured by subsoiling. They noted that gypsum greatly enhanced the leaching of salts and intermittent ponding had no practical advantage over continuous ponding.

Ismael (1993) showed that soil salinity of 24 dS m^{-1} leached down to 3 dS m^{-1} with a total distilled-water flow of around 230 mm in laboratory experiments.

Jaynes and Rice (1993) reported that the leaching behaviour of salts is affected by methods of irrigation. Flood irrigation resulted in greater variability of tracer velocity than drip irrigation.

A breakthrough curve is usually produced to present the changes in solute concentrations in the effluent and it is a plot of 'the ratio of the concentration of effluent (C) to the concentration of displacing fluid (C_o) versus the number of pore volumes of effluent. The displacing fluid is the fluid added to replace the fluid already in a soil and the pore volume is the volume of the porous medium occupied by fluid (Kirkham and Powers, 1971).

The above citations show that the salt leaching behaviour varies with the soil characteristics, methods of water application, chemical being used, etc., i.e. the leaching behaviour is situation specific. However, the present experiment is dealing with highly salinized (40-50 dS m^{-1}) soil profiles which developed after growing a crop under saline water table management. It is hoped that the present experiment will be able to prescribe economic use of water for leaching purpose.

The objective of this experiment was to determine how much water is needed to reduce a given degree of salinity and the specific objective was to obtain a favourable condition in the soil for the next season's experiments.

9.2 Materials and methods

9.2.1 Leaching experiment

The leaching experiments were conducted in lysimeters (see sections 4.1.1 and 4.2.3) after harvesting the ryegrass crop. The diameter of each lysimeter was 106 cm with a height of soil column of 130 cm. The initial soil salinities at different depths are given in Table 9.1.

Leaching continued for 60 days. During the first 12 days, 20 mm of tap water was sprayed onto the soil surface each day and in the following 48 days, alternately 40 mm of water was ponded for a period of 24 h and then drained for another 24 h. Total depth of water applied was 1200 mm. Around 450 mm was required to restore the equilibrium soil moisture profile in each lysimeter and at the beginning of leaching experiment around 360 mm water already was in the soils. The effluents were collected each time and salinity was determined in laboratory. The first effluent was collected after 80 mm of water application. During leaching, the lysimeters were covered to prevent evaporation loss.

Table 9.1: Soil salinitity profiles in the lysimeters at the beginning of leaching experiment.

Lysim-eters	Soil salinities at different depth of lysimeters (dS m^{-1}) Depth of soil profiles (cm)						
	0~15	15~30	30~45	45~60	60~75	75~90	90~130
LYS1	30.5	17.7	14.6	11.3	9.4	9.4	9.4
LYS2	30.4	16.1	12.6	12.4	11.7	10.2	9.4
LYS3	27.7	16.4	13.4	11.5	10.7	9.7	9.4

9.2.2 Salinization of soils

During salinization, 7.5 and 15.0 dS m^{-1} NaCl solutions were used in two different lysimeters. The soils were saturated with saline solutions from the bottom of the lysimeters & kept ponding 40 mm onto the surface for 24 h and then drained for another 24 h. The alternate process of resaturation, ponding and draining continued until the desired salinity was attained. Each time soil solutions from different soil depths were collected and electrical conductivity of the samples were determined. The water content in the lysimeters soils at the beginning of salinization was around 0.8 pore volumes and the salinities at different depths are presented in Table 9.2.

Table 9.2: Soil-water salinities in the lysimeters soils at the beginning of salinization experiment.

Lysimeter	Soil-water salinities at different soil depths (dS m^{-1}). Depth of soil profiles (cm)						
	0~15	15~30	30~45	45~60	60~75	75~90	90~130
L-7.5	0.40	0.40	0.40	0.50	0.56	1.00	0.85
L-15	0.40	0.40	0.40	0.40	0.45	0.63	0.63

9.3 Results and discussion

Fig. 9.1 shows that after application of 1200 mm water, soil salinity leached down from 15.5 dS m^{-1} to 1.0 dS m^{-1}, while with only 80 mm water concentration fell to 9.0 dS m^{-1}. This 80 mm water mainly diluted the solute already present in the soils as it was retained in the soil. Fig. 9.1 also shows that 400 mm water was enough to leach down two-third of the salts, whereas another 800 mm of water was needed to leach down the remaining one-third. Figs. 9.2 and 9.3 represent the breakthrough curves with respect to resident time and number of pore volumes respectively. Fig. 9.3 shows that three-quarters of one pore volume of water flow was enough to leach down the salts to an acceptable range for crop establishment. Overall results indicate that complete removal of salt requires a much bigger throughput of water. This can be avoided if some salinity is allowed for crop establishment.

Trends of salinization of soil profiles are presented in Figs. 9.4 and 9.5. They show that one pore volume of solution was needed to displace and equilibrate with the soil water (0.8 pore volume) in the lysimeters.

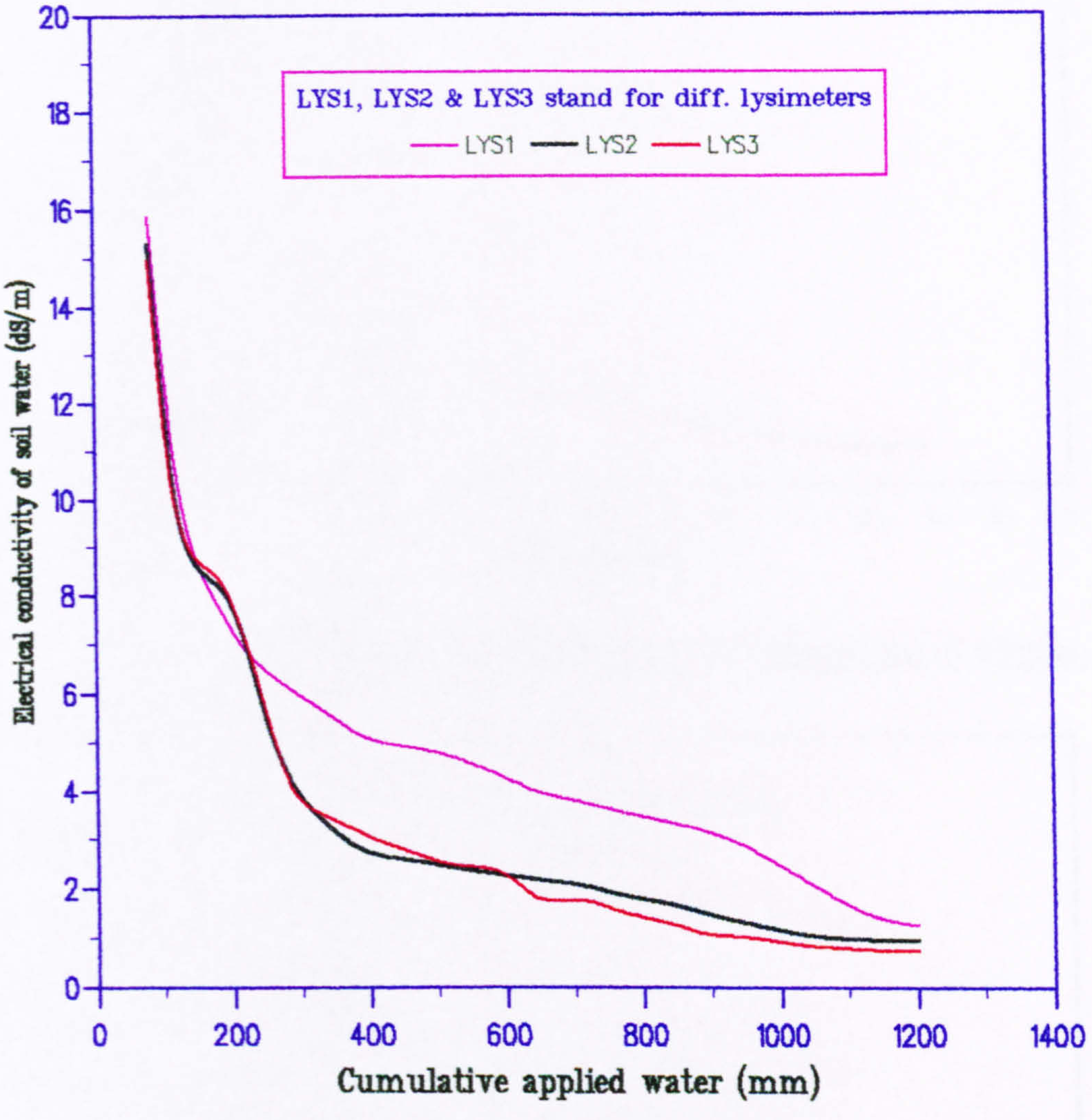


Fig. 9.1: Leaching of salts versus volume flow of water.

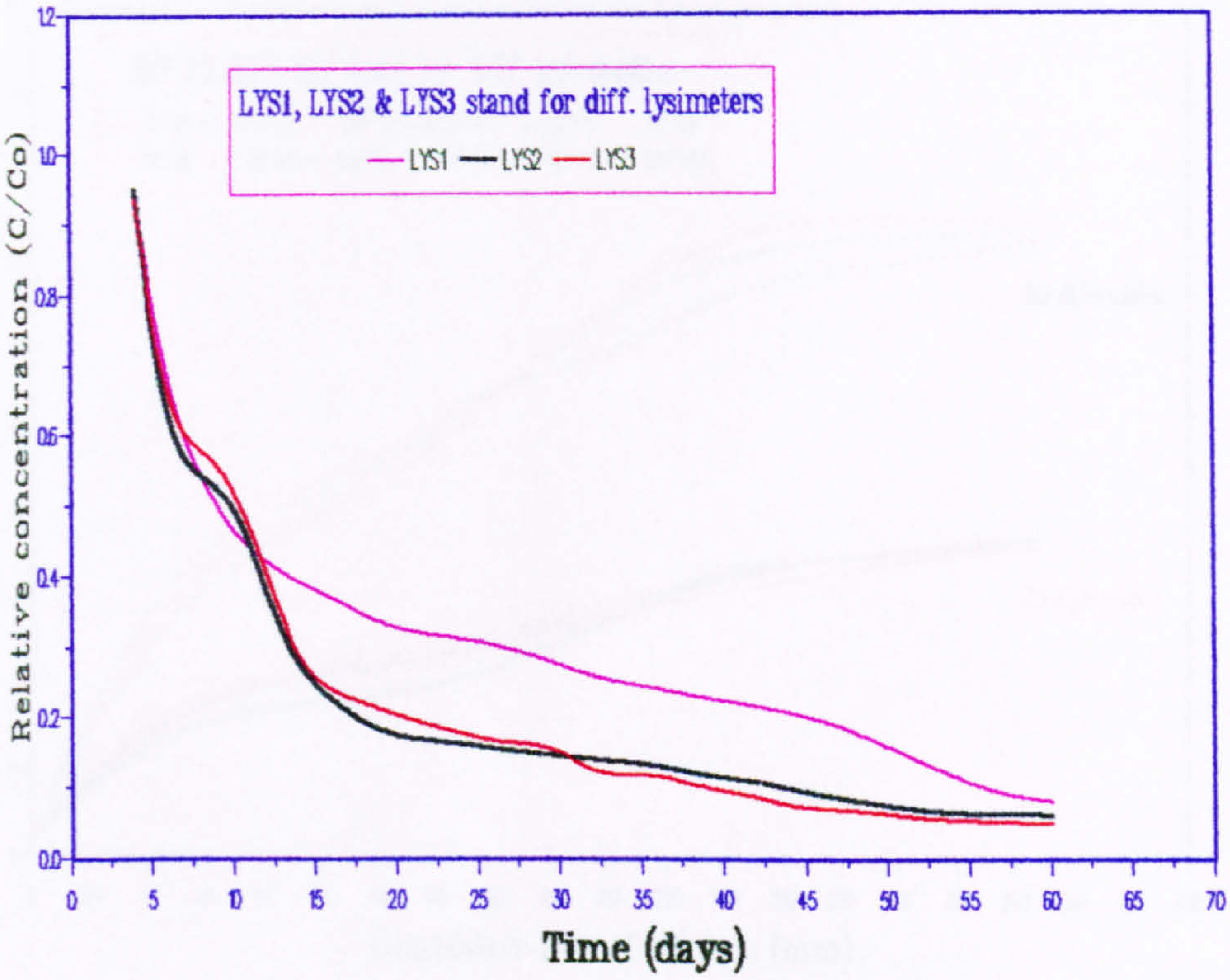


Fig. 9.2: Breakthrough curves with respect to resident time of solution.

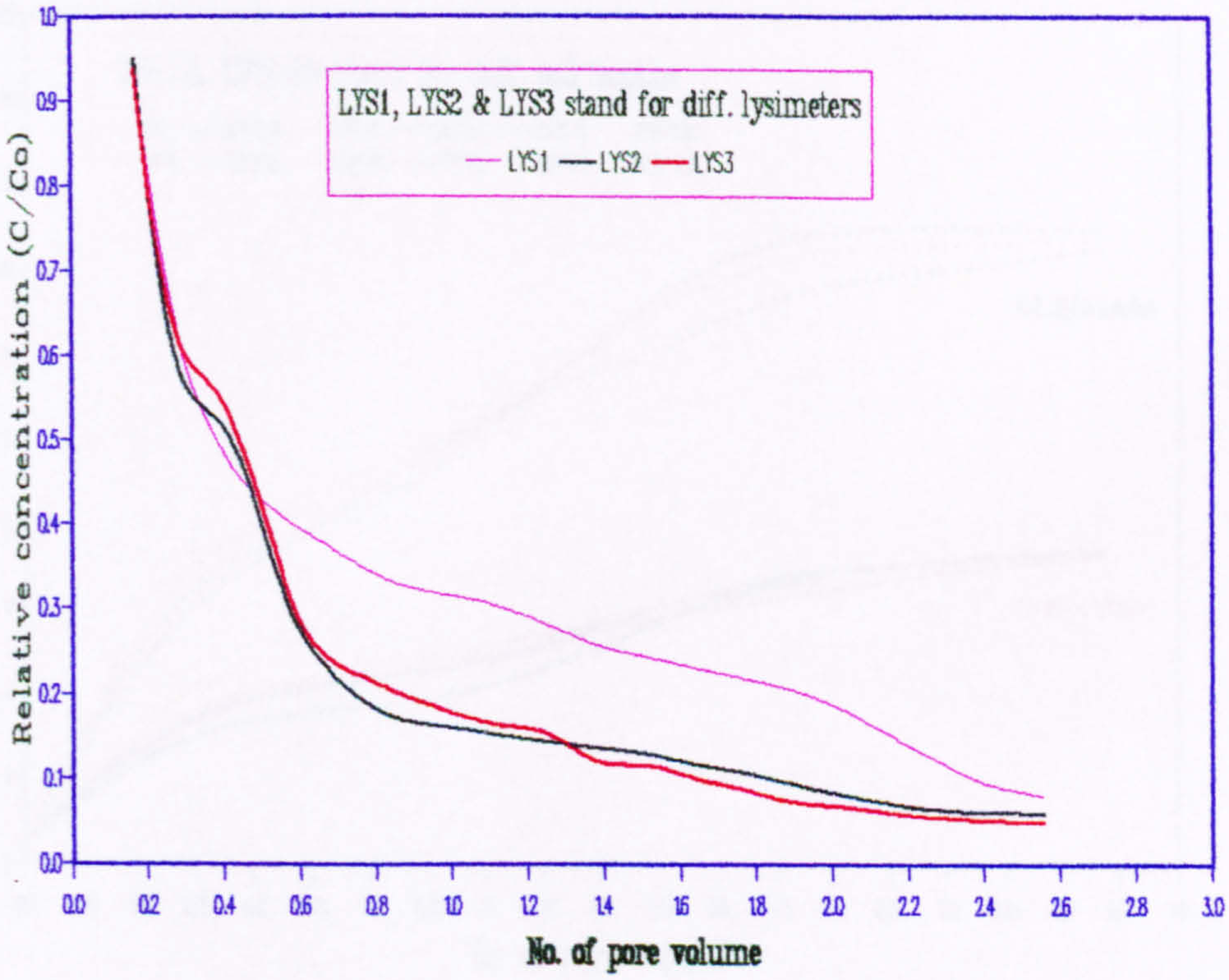


Fig. 9.3: Breakthrough curves with respect to No. of pore volumes.

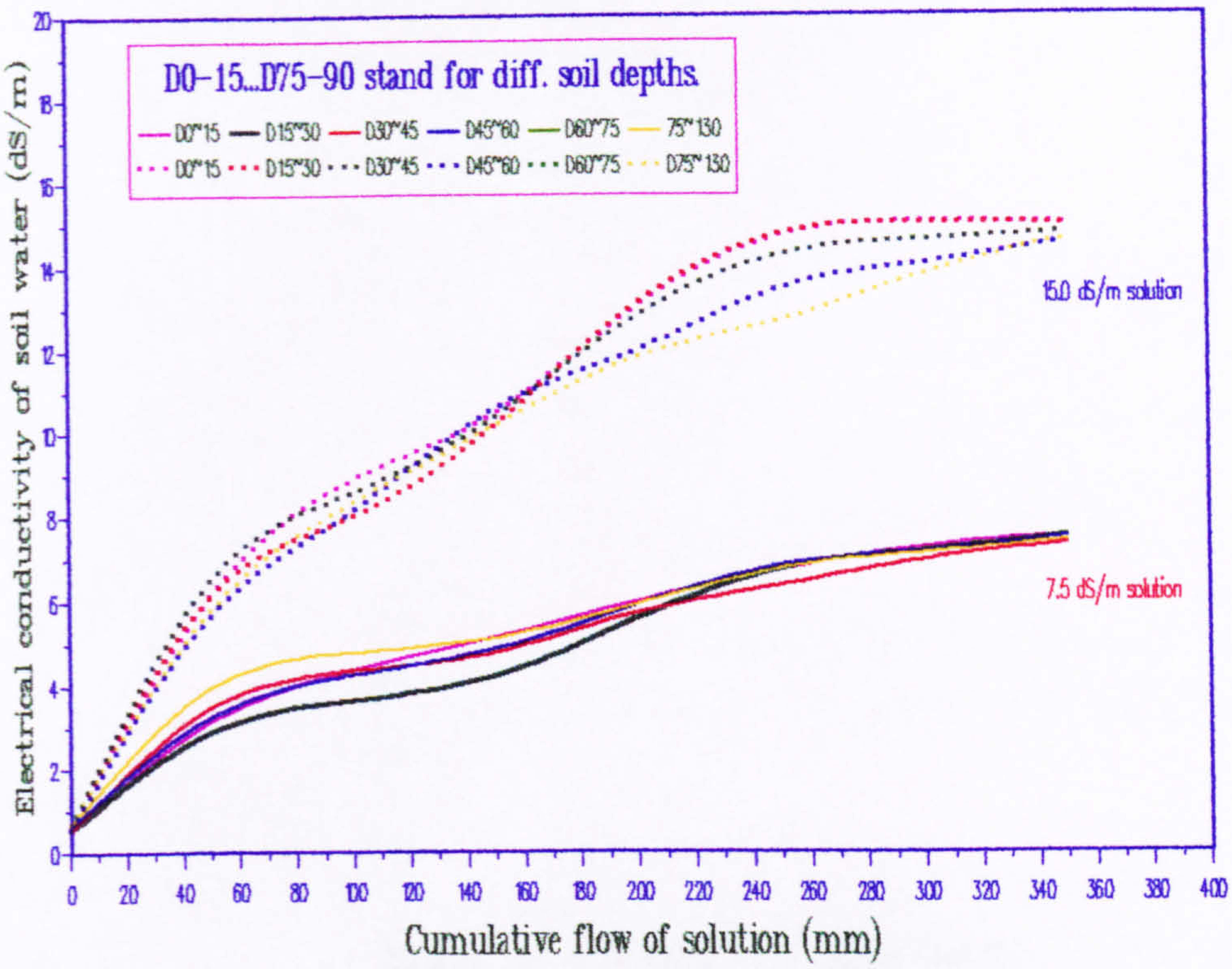


Fig. 9.4: Trend of soil salinization versus cumulative solution flow

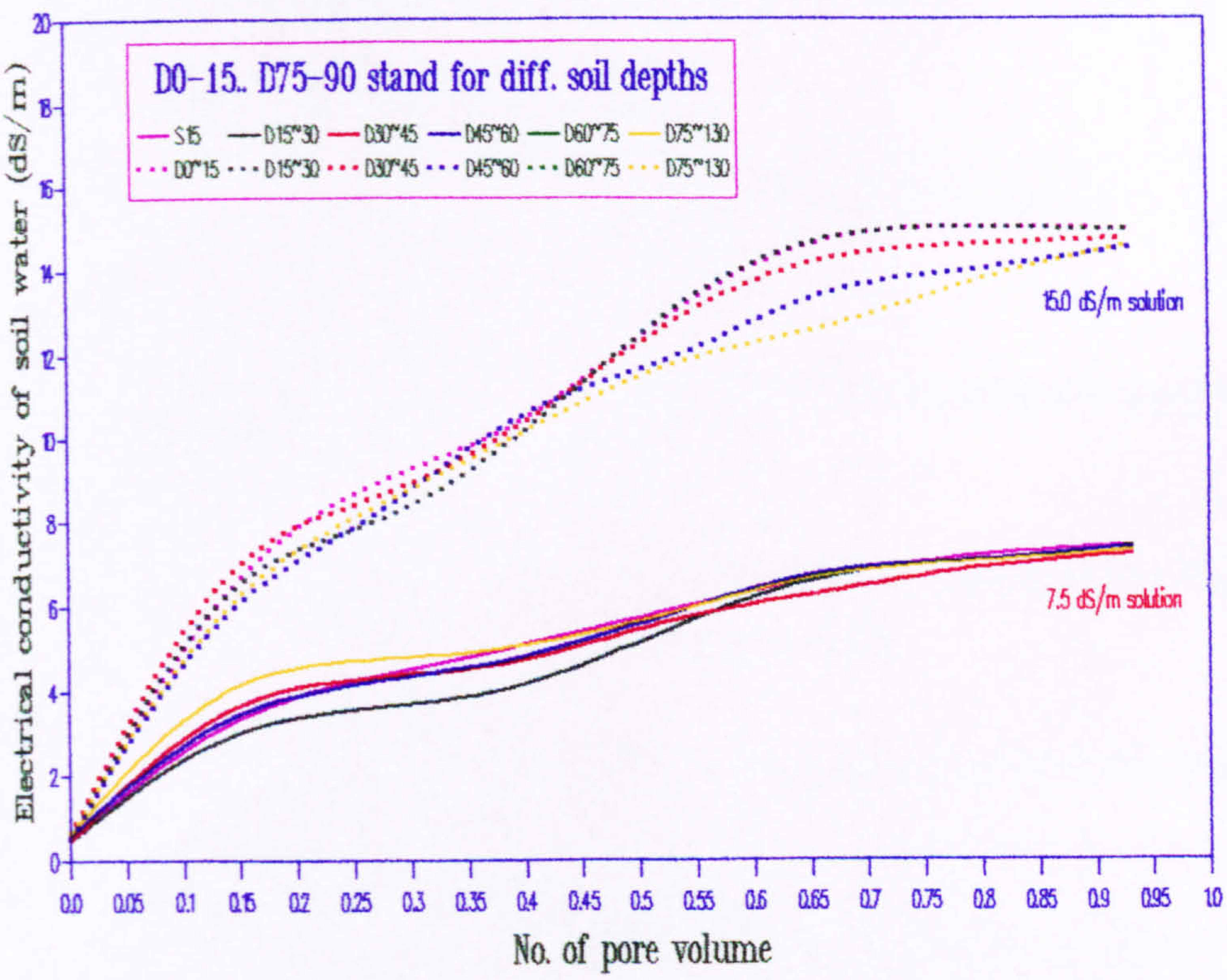


Fig. 9.5: Trend of soil salinization versus No. of pore volumes of solution flow

Part 4: CONCLUSIONS

Chapter: 10

CONCLUSIONS

10.1 Achieving the objectives

The following summarizes the main achievements of the work:

* How does ryegrass respond to prolonged & high salinity stress as well as matric stress in a climate of low atmospheric demand ? The answer to this question can be found in this investigation.

* Crop yield potential with the reported saline water management (water use from equilibrium saline soil water profile and water table) has shown that, though rootzone salinity during the crop growing period was much greater than the initial soil water salinity, the yield reduction was almost at the reported margin of yield reduction with the initial salinity level.

10.1.1 Crop water use and water table contribution

* Most of the water supply (70 to 75%) was from the soil water reserves andthe rest (25 to 30%) from the saline water tables.

* The difference in water table depth (60 to 120 cm) caused no significant difference in contribution even with same maximum rooting depth.

* Water table contribution proved to be useful as a supplementary resource of total crop water demand.

* With the same water depth (90 cm), saline water table contributed more to crop water use in the first half of the seven months cropping period, though total water table contribution was around 5% more from the non-saline (0.4 dS m^{-1}) water table.

10.1.2 Crop yield potential & root function

* The yield decreased proportionately with increased salinity of initial soil water and water table.

* The root elongation and root mass production were found to be smaller in the saline treatments than in the non-saline, and the more the salinity, the more the reduction.

* The drymatter proportion of crop yield was higher in the saline treatments than non-saline.

* The lower root zone had a potential role for continuing crop growth when the upper root zone became unable to extract water due to osmotic and matric stresses. Growth continued even when salinity in surface soil exceeded 40 dS m^{-1}.

10.1.3 Root zone salinization

* The salt accumulation was mainly confined to the top 15 cm root depth. Salinity increased up to 40-45 dS m^{-1} (4 fold of the initial salinity) in the top 5 cm depth, and within 15 to 45 cm rooting depth was less than 2 fold.

* Estimating the root zone salinity based on convective transport, ignoring hydrodynamic dispersion, agreed well with measurements.

10.1.4 Root water uptake model

* The model was successful in assessing root uptake behaviour under both osmotic and matric stress conditions and thus enabled us to forecast appropriate irrigation management decisions.

* The model considered the osmotic and matric stress to be additive function and require an osmotic adjustment effect for crop water management under saline situations.

10.2 Suggestions for future research

i) The present investigation showed that two-thirds of the total water use was supplied from soil water reserves but, it is likely that, in a climate of high atmospheric demand, the soil moisture reserves from equilibrium soil water profile will not be sufficient to fulfil the crop water demand. Therefore, the present saline water management method needs to be tested under more realistic field situations to assess the crop production potential. More water supply may be obtained by allowing shallower water table (i.e. close to the rooting depth), and also allowing the fluctuation of water table down to the growing rooting depth. Considering the rooting depth found in the salinity context, a shallow water table, e.g. 30-45 cm deep, may be considered for the future investigation. Prathapar and Meyer (1992) suggest that, plants can abstract around 20 % of total water table contribution directly from the water table, if water table remains close to the rooting depth.

ii) The model needs to be tested with field experimental data to make it useful in realistic situations.

iii) More careful experiments might show the need for osmotic adjustment in the model.

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