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AN INVESTIGATION OF SALINITY FLUCTUATIONS IN SOILS OF A NORTHERN UTAH MARSHLAND

by

Gary Roy Newman

A thesis submitted in partial fulfillment
of the requirements for the degree

of

MASTER OF SCIENCE

in

Engineering
(Environmental)

Approved:

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Logan, Utah

1979

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My gratitude is extended to all those who provided technical or consulting assistance to this effort. I would particularly like to thank my parents who provided moral support through their advice and encouragement. Most of all I'd like to thank my wife whose work, encouragement, and love were valuable to the completion of this paper.

Gary Roy Newman

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ABSTRACT

An Investigation of Salinity Fluctuations In Soils of a Northern Utah Marshland

by

Gary Roy Newman, Master of Science

Utah State University, 1979

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Department: Civil and Environmental Engineering

Ogden Bay Waterfowl Management Area is one of many marshlands bordering the Great Salt Lake that are extremely important as breeding habitats for North American waterfowl. Salinity, both of surface and interstitial waters, is an important factor in maintaining a suitable habitat within the marsh. For this reason, a study was undertaken to determine the factors affecting the salinity of surface and interstitial waters. A small (approx. 100 acre) subunit was studied extensively to determine; 1) variations in surface water salinity with flowrate and water level; and 2) variation in interstitial water salinity with depth and location within the subunit. Field data indicated a substantial increase in salinity (as measured by electrical conductance) with depth in the interstitial waters. Wide variation in interstitial water conductivities was also observed within the study unit. Of major significance were large increases in interstitial water salinities in areas where the soil had dried, due to a low water level, and again re-wetted. This was in contrast to relatively constant soil water salinily observed in areas that were perenially flooded.

In addition to the field study, a laboratory study, using three 20 cm diameter undisturbed soil cores, was performed to determine the factors affecting the movement of salts through the marsh soils. The surface and bottom of the cores were fed with fresh water and brine, respectively, for a period of three months, while interstitial water samples were taken to monitor changes in soil water conductivities with depth of soil. A computer model was developed to simulate the experiment, as well as to help interpret the experimental results. The comparison between the observed data and data predicted by the model, as well as the results of the field data, indicated that the major mechanisms affecting fluctuations in soil water salinity in the Ogden Bay marsh system is the movement of water through the soil.

(132 pages)

INTRODUCTION

The marshes bordering the Great Salt Lake are important breeding grounds for the migrating waterfowl of the Western North American continent. For this reason, waterfowl management areas have been established to help insure that a suitable habitat is maintained for waterfowl production. Adequate supplies of high quality water are essential for the maintenance of these marsh ecosystems. However, rapid urban and agricultural development in the Salt Lake Valley have increased the demand for the limited fresh water supplies in the area. Concern has arisen over the possibility that further development will reduce the quality and quantity of fresh water available to these marshes. This concern has prompted a study to determine the effects of reduced fresh water flow through the marshes on the quality of the marsh water themselves.

The study area involves the Ogden Bay Waterfowl Management Area, located adjacent to the Great Salt Lake, approximately 12 miles west of Ogden, Utah (Figure 1). The marsh system consists of three units, of which Unit 1 is the most convenient for study as it is completely surrounded by dikes and flow is controlled through culverts and concrete structures. For this reason, Unit 1 was chosen for study. It consists of approximately 3000 acres, within which cross-dikes have been constructed forming smaller units. Within Unit 1, a smaller unit of approximately 100 acres was chosen for intensive study. Virtually all the water entering the marsh comes from the middle run of the Weber River and enters through a gate structure in the

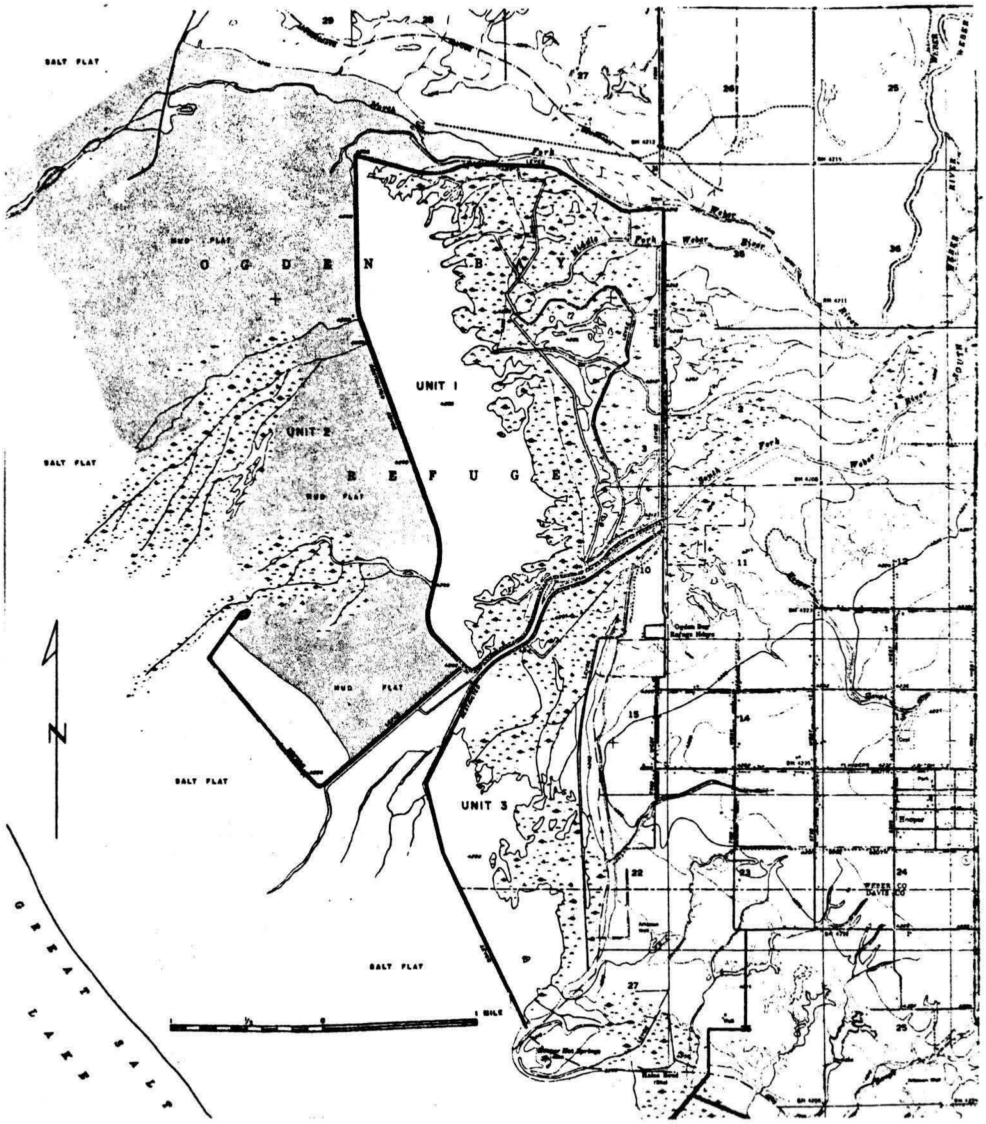


Figure 1. Map of Ogden Bay Waterfowl Management Area.

northeast corner of the refuge. Water entering the area flows into a channel along the eastern boundary of the marsh. From there water is distributed throughout the marsh through a series of culverts and concrete control gates placed at intervals in the cross-dikes. Water depths within the marsh range from a few centimeters to 1.5 meters. Areas in shallow water are covered with various forms of vegetation; some of the most prominent being cattail (*Typha latifola*), bullrush (*Scirpus acutus*), and saltgrass (*Distichlis stricta*).

For this study the primary water quality parameter of importance is salinity. Many studies have shown the variation in sensitivity of marsh plants to salinity. Conceivably, a reduction of fresh water flow through the marsh could cause a build-up of salts in the surface waters to a level detrimental to the marsh vegetation. This potentially could occur through two major mechanisms. The first, evapotranspiration, which has been extensively studied on Northern Utah marshlands, acts to concentrate the salts brought into the marsh in the inflow water. The second potential mechanism is the movement of salts between the surface waters and the interstitial waters of the marsh soils. The presence of high salinity water just a few centimeters below the soil surface has been well documented. However, the movement of salts within the soil has been studied very little to date. Thus, a determination of the factors affecting salinity fluctuations within the marsh soil and the subsequent effects on surface water is necessary if a comprehensive assessment of water quality and water flow is to be made for the Ogden Bay marsh system. For this reason, this research effort has been devoted to the study of the transport of salts within the soil system, with particular emphasis on the movement of salts with soil water.

The data collected and used in this study were generated in three separate procedures. First, field data consisting of electrical conductivity measurements of surface and interstitial waters were collected over a period of approximately 18 months. In addition, water movement in terms of flow, was measured at various points within the marsh system. The second procedure was a laboratory experiment using 20.3 cm (8 inch) diameter undisturbed soil cores, which were taken from the marsh itself. This experiment was designed to monitor interstitial water salinity over a period of time under controlled conditions of static hydraulic head. Finally, the third procedure involved laboratory analysis of the soil used in the soil cores themselves. The soil was characterized in terms of hydraulic conductivity, soil texture, and bulk density. With this data, the soil-water system can be characterized as to its ability to transmit water and salts under different hydraulic conditions.

LITERATURE REVIEW

The literature dealing with the relationship between surface and interstitial water salinity can be divided into two general categories: (1) studies on marshes in general, including specific studies on salinity fluctuations, as well as general information on marshes and marsh soils; and (2) literature dealing with the movement of water and solutes through soil. A review of the significant literature in these two categories follows.

Marshes and Marsh Soils

In a report that deals specifically with Northern Utah marshlands, *Christiansen and Low* [1970] made extensive studies of surface water salinities, evapotranspiration, and salt tolerances of marsh plants. The authors developed a relationship between salinity of incoming and outgoing marsh water and the total water requirement for the marsh. Although they did not consider the effect of soil water salinity on surface water salinity directly, their work is important in that they provided background information with which to compare present results.

In the *Christiansen and Low* report, electrical conductivity (EC) data is reported over a five year period from 1959 to 1963 at Ogden Bay Waterfowl Management Area. Average EC values for spring, summer, and fall consistently show at least a two fold increase in EC from inflow to outflow through Unit 1. Average inflow EC over the five year period ranged from 0.7 mmhos/cm in the spring and fall seasons to 1.0 mmhos/cm for the summer. Average outflow EC values ranged from 1.6 mmhos/cm in the fall to 1.8 mmhos/cm in the spring and 2.4 mmhos/cm in the summer. A general classification of water quality in waterfowl marshes, based on studies of salinity tolerances of marsh plants, shows that water flowing out of Ogden Bay Waterfowl Management Area ranged from good to fair (Table 1).

Another important aspect of the work done by *Christiansen and Low* was the development of relationships which predicted evapotranspiration from Utah marshlands based on climatic data. Evapotranspiration is important in salinity fluctuations as it acts to concentrate salts by reducing the volume of water moving through the marsh. The authors used established evapotranspiration equations, such as the Christiansen method, the Blaney-Criddle method, and the Grassi method, and developed empirical constants to fit data from Howard Slough Waterfowl Management Area to predict water losses from a marsh.

Based on the evapotranspiration relationships and extensive studies on the tolerance of different species of marsh plants to salinity, *Christiansen and Low* developed an equation to estimate the water requirements of a marsh. In inches of depth, the annual water requirement for a marsh (WR) is given by:

$$\mathrm{WR} = Fs(Et-Pn)$$

where Et is the seasonal evapotranspiration requirement for the wetted area in inches, Pn is the mean annual precipitation in inches, and Fs is a salinity factor.

The value of Fs is given by:

$$Fs = \frac{(Co-k)}{(Co-Cw)}$$

Co is the maximum desirable average annual conductance of the outflow from the marshes (6 mmhos/cm); Cw is the average annual conductance

Table 1. Tentative classification of water quality for waterfowl management areas in Utah.

Class	Rating	Conductance	Salinity (ppm)
1	Excellent	Less than 1	0 - 640
2	Good	1 - 2	640 - 1280
3	Fair	2 - 4	1280 - 2560
4	Poor	4 - 8	2560 - 5120
5	Restrictive	More than 8	More than 5120

Christiansen and Low, 1970. p. 85

of available water; and k is the conductance of a saturated calcium bicarbonate solution (approximately 0.3 mmhos/cm).

Generally, Fs values range from 1.0 to 2.5 depending on the specific conditions. Thus, from these equations, it can be seen that salinity plays a major role in determining water requirements for marsh systems.

Nelson [1955] completed a general survey of soil salinity changes over a 6 year period (1946 to 1952). The author maintained records on plots representing intermittently flooded areas, areas that would have 0-18 cm (0-7 inches) of water flowing over them, and an area that would constantly be flooded with 25-30 cm (10-12 inches) of standing water. Salinity was monitored at soil depths of 0-7.6 cm (0-3 inches), 7.6-15 cm (3-6 inches), 15-30.5 cm (6-12 inches), and 30.5-61 cm (12-24 inches).

The results of these studies showed that over a period of six years, intermittent flooding had little effect on the salinity of soil at all depths studied. The areas that were covered with flowing and standing water were within an irrigation system that was constructed in 1947, prior to the study. Before this time the areas were dry, the water table being 15-30.5 (6-12 inches) below the surface. Monitoring of soil salinity began when the area was initially flooded. Over the six year study period, salinity was markedly reduced, particularly in the first 8 cm of soil. Most of this decrease in soil salinity occurred in the first one to two years after flooding. Of particular importance is the fact that prior to flooding, soil salinity decreased with depth of soil. After a period of flooding, this trend reversed itself, with salinity increasing with depth.

Also of interest is the variation of soil salinity along a transect. Figure 2, taken from the *Nelson* paper, shows typical variation in soil salinity, soil pH, and vegetation types along a 427 meter (1400 foot) transect. The important features to note are the very low soil salinities (as indicated by the EC of a 1:5 soil-water suspension) at all depths, in areas covered by 15-30.5 cm (6-12 inches) of water. In contrast, the areas covered by shallow water show much higher soil salinities at depths from 8-61 cm (3-24 inches). Another interesting feature of this figure is the very high salinity in the surface 7.6 cm (3 inches) of soil in the region of barren ground between 366-427 m (1200-1400 feet) along the transect. Also, the soil salinity was found to vary during the course of the year. At one station *Nelson* reported that a 1:5 soil suspension had an EC of 1.33 mmhos/cm during the spring months and 8.10 mmhos/cm in the late summer and fall. The author did not mention, however, whether or not this station was intermittently flooded or submerged yearly.

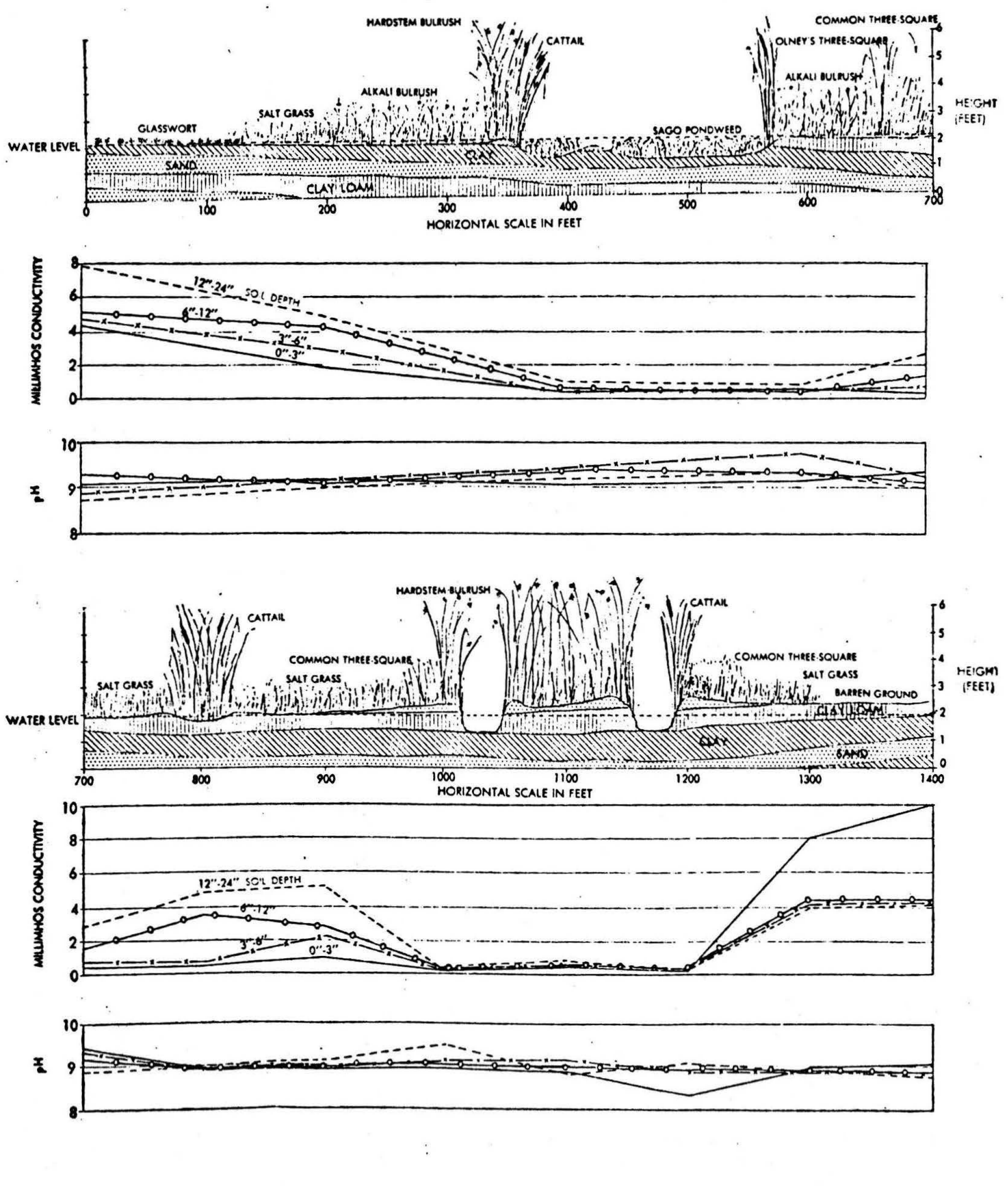


Figure 2. Variation in soil salinity, pH, and vegetation type along a 427 m (1400 ft) transect [*Nelson*, 1955].

Salt affected soils and marshes can be classified into three basic categories [*Chapman*, 1960]: (1) non-saline alkali soils, which are high in exchangeable sodium, but low in soluble salts; (2) saline alkali soils, which are high in both exchangeable sodium and soluble salts; and (3) saline soils, where soluble salts dominate. Though this is a purely arbitrary classification, it is useful as an indicator of the types of soil-salt systems that exist.

Chapman describes five evolution processes of alkali and saline soils. The first phase, salinization, refers to the actual accumulation of salt in the surface layers of soils. This could be due to capillary rise of saline water from subsurface deposits, or from evaporation of saline surface water. The second process, alkalinization, is the adsorption of sodium ions as a result of base exchange. Third, desalinization, refers to the removal of salts through leaching. This often acts in the presence of calcium, which will exchange with sodium, on the soil particle surface, allowing the sodium to become soluble and thus readily leached out. The fourth process, degradation, is the exchange of hydrogen ions for sodium ions; again allowing the sodium to come into solution and be leached from the soil. Finally, the fifth process, regradation, refers to the reaccumulation of soluble salts.

In addition to the chemical processes mentioned above, nine physical factors that affect the salinity of inland marsh soils have been listed [*Chapman*, 1960]:

1. Rainfall, which causes downward leaching of salts through soil.
2. Proximity of streams and creeks. Good drainage to a stream will act to flush the salts from the soil.

3. Nature of soil. Fine silty mud will retain more salt than sandier soil.
4. Presence or absence of vegetation. Bare soils always have higher salt concentrations than soils covered with vegetation. Vegetation acts to bring a continual rise of water to the surface of the soil and also reduces evaporation from the surface.
5. Slope of ground surface. Greater slopes will produce higher drainage rates and hence greater flushing.
6. Depth of soil water table. The nearer the water table is to the surface the more constant the soil salinity.
7. Depth of subsurface salt deposits. The greater the depth, the less saline the surface layers.
8. Inflow of streams into area. Streams will bring salt into the area or act to dilute or flush the salt from the marsh.
9. Climate and temperature. High temperatures, particularly in areas that are not continually submerged, may have profound effects on salinity. This is of great importance in inland marsh systems.

Specific factors will be more important at one location than at another.

The relationship between soil salinity and salinity of applied surface water was explored in a study done on Suisun Marsh in California [*Rollins*, 1973]. This study was conducted in two phases. The first was performed at four private hunting clubs, to determine the salinity response of soils to the normal applications of slough water as applied under routine management conditions. The second phase was undertaken to determine the interrelationship between soil salts and the salinity of applied water, using a control pond and high salinity infiltrometers. The results show a significant relationship between surface water salinity and soil salinity.

During Phase I, surface and soil salinities were monitored for a 12 month period. During this time the marsh areas were flooded for 4 months, drained, and allowed to remain relatively dry for the remainder of the year. The water used for flooding was relatively fresh, averaging 1000 to 8000 parts per million (ppm). The results of the study showed that significant reductions in soil salt concentrations, as well as amounts of salts (as determined by a 1:5 salinity extract) in the first 20 cm (8 inches) of soil accompanied the flooding of the marsh areas. Reductions of 72 to 85 percent in salt concentrations and 50 to 67 percent in salt amounts occurred during the flooded periods when compared to the dry season. A majority of this reduction occurred during the first month of flooding. Whereas salinity fluctuations in the top 20 cm (8 inches) of the soil were large, salinity in the 50-100 cm (20 to 40 inch) range was observed to be relatively constant during the year. Differences in response between the areas monitored were explained by the differences in soil types. The denser clay soils showed the smallest reduction in salinity.

The second phase of the Suisun Marsh study used infiltrometers located in a pond where water level could be controlled. Water level in the infiltrometers was controlled to simulate the flooding, draining, and dry periods in the actual marsh areas. Water of 20,000 ppm and 10,000 ppm salinity, as well as slough water of varying salinity, was used for flooding. The results of this study showed a statistically significant correlation between applied surface water salinity and soil water salinity at the one and two foot levels within the soil. Application of 20,000 ppm water was applied to dry, highly saline soils, and significant decreases in soil salinity in the systems studied was clearly a function of moisture and salinity of overlying water.

Of related interest is the relationship between interstitial water salinities and surface water salinities in a marine (tidal) marsh system. In one study [*Lindberg and Harriss*, 1973] soil cores 25 cm deep were taken at different times during a tidal cycle. Of primary interest were salinities of interstitial water at 2.5 cm and 23.5 cm below the mud water interface. Results showed that interstitial salinity at 2.5 cm decreased slightly in response to lower surface water salinity; but rose again after a period of exposure to more saline water. The authors suggest that mass exchange or convective interchange, due to a density gradient, is responsible for these fluctuations at shallow soil depth. At 23.5 cm, however, salinity of interstitial water decreases as depth of overlying water decreases. This occurred in spite of the fact that surface water salinities remained relatively constant. In this case, the authors believe that bulk flow of subsurface freshwater acted to decrease the salinity in response to the lower hydrologic head of salt water at low tide.

Movement of Water and Solutes Through Soils

The principles which affect the transport of solutes through soil must be addressed if the potential for salt movement from interstitial to surface waters is to be adequately described. These principles can be divided into three categories: (1) the physics of water flow through soils; (2) the effect of chemical composition of the soil and permeating solution on soil-water movement; and (3) the diffusion of solutes through the soil in response to a concentration gradient.

The basic equation which describes water movement through a porous media, such as soil, is Darcy's Law, given as:

$$J_w = -K \frac{dH}{dX}$$

For this equation, dH/dX is the hydraulic potential gradient across the soil (expressed in units of length per length when H is defined as an equivalent head of water), J_w is the water flux density (flow per unit area, $L^3/T \cdot L^2$), and K is a constant called the hydraulic conductivity (expressed in units of L/T). The hydraulic conductivity is probably the most important term in Darcy's Law with respect to characterization of a soil. The hydraulic conductivity is dependent on several soil properties [*Hillel*, 1971]. Total porosity and size distribution of soil pores are very important. A soil with low porosity but large pores, characteristic of a sandy soil, will have a larger hydraulic conductivity, under saturated conditions, than a clay soil, which is characterized by high porosity and very small pores. The value of K, however, is not dependent on the soil alone. Density and viscosity are important physical properties of the fluid which can also affect hydraulic conductivity. In addition, the chemical properties of the soil and the permeating solution can also have a profound affect on the hydraulic conductivity, as will be explained later.

Another factor that has a major effect on K is the water content of the soil. The transition from a saturated soil to an unsaturated soil entails a large drop in hydraulic conductivity, often several orders of magnitude. There are several factors that account for this. First, as soil pores empty of water, the cross-sectional area for water flow decreases. Related to this is the fact that the large pores, which are highly conductive at saturation, are the first to empty; thus, leaving the smaller, less conductive pores to carry the flow. These empty pores must be circumvented, thus increasing the tortuosity of the conducting media.

Though the hydraulic conductivity decreases dramatically with decreasing water content, flow will occur if a hydraulic potential gradient exists. Unlike saturated soil, where the water is under positive pressure, water in unsaturated soil is under a negative potential energy (suction), being held to the soil particles by matric forces (forces resulting from the interaction of soil water with the soil matrix). As the water content decreases, the potential energy holding this water to the particles increases. Thus, a hydraulic potential exists in unsaturated soil when a moisture gradient exists. Water will flow from a region of high moisture and higher potential energy, to a region of low moisture and more negative potential energy. Thus, the hydraulic potential gradient provides the necessary energy difference to cause flow.

The variation of hydraulic conductivity with water content and suction is important in the characterization of soils. Hydraulic conductivity varies differently with moisture content for soils of different textures (Figure 3). It is evident that hydraulic conductivity decreases much faster with decreasing potential energy (decreasing water content) for sandy soils than for clay soils. As desaturation occurs in sandy soils, the large pores empty, thus severely limiting flow. On the other hand, many of the small pores in clay soils remain filled even at high suction and continue to conduct fluid. These types of graphs can be useful in characterizing a particular soil.

The concept of the dependence of hydraulic conductivity on soil moisture is important when describing water flow through a soil system

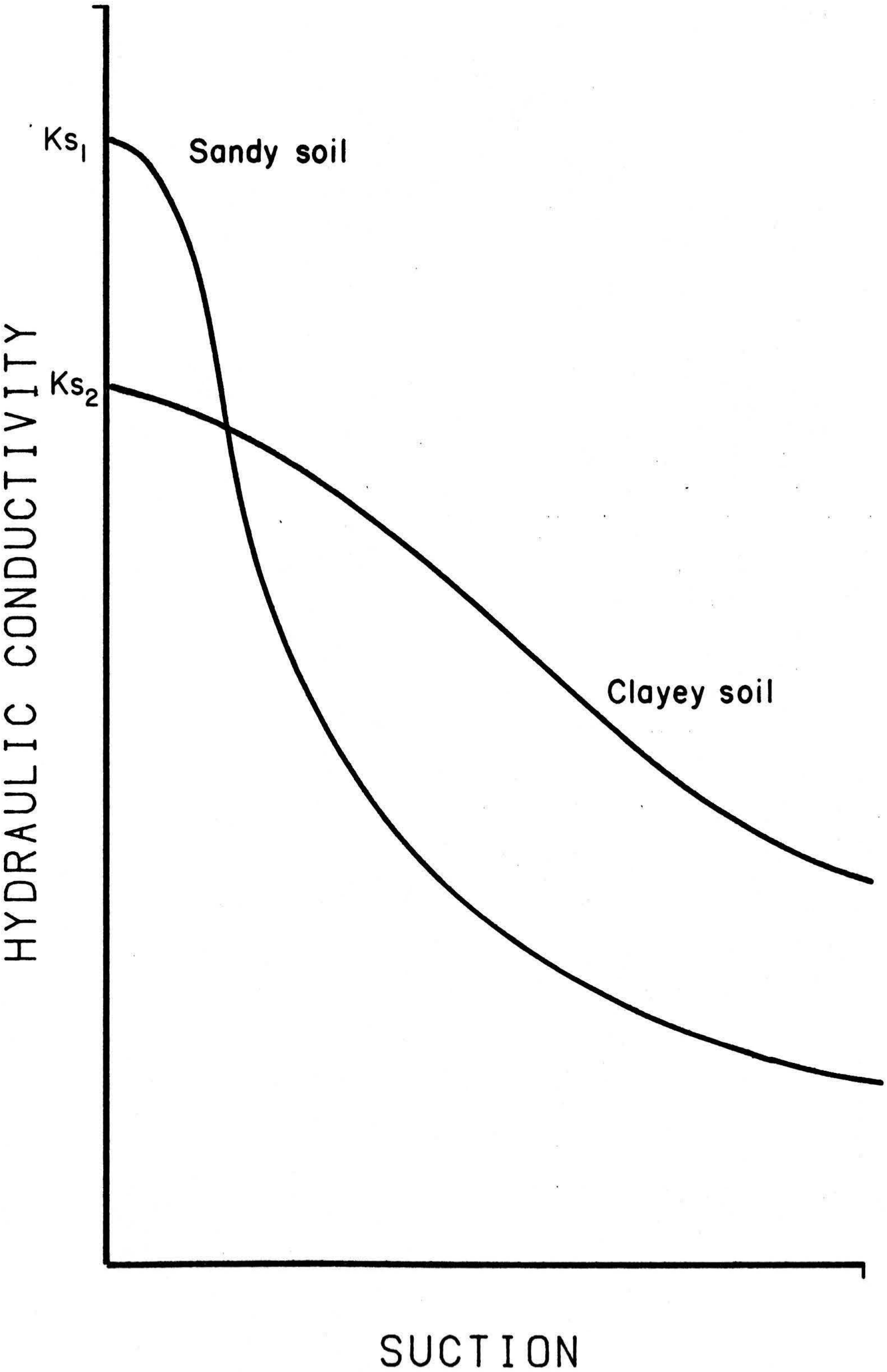


Figure 3. Variations of hydraulic conductivity with soil marshes suction in soils of different texture [*Hillel*, 1971].

consisting of two or more layers. *Srinilta, et al.*, [1969] studied such a system, in which a profile consisted of a topsoil with a smaller saturated hydraulic conductivity than its subsoil. In such a system, the subsoil may remain unsaturated even though water is ponded over the topsoil.

The experimental procedure utilized by *Srinilta, et al.*, involved measuring soil water pressure at given depths within the two layer soil column while the water flow was maintained at a steady-state. In general, the results showed that the observed steady-state flux rates for the two layer systems were greater than the saturated hydraulic conductivity of the topsoil, but less than the saturated hydraulic conductivity of the subsoil. Specifically, the authors found that by increasing ponded water depth from 5 cm to 25 cm, both soil water flux and soil water pressures were increased. Also, for a given depth of ponding, increasing topsoil thickness decreased soil water fluxes and soil water pressures. On the other hand, increasing subsoil thickness had an almost insignificant effect of measured soil water fluxes. In addition, the type of topsoil was shown to have a profound effect on the soil water pressure profiles. The authors demonstrated that for a more sandy topsoil (56 percent sand) the soil water pressures were approximately twice those found for the same depth of a clay soil (10 percent sand). Also, the soil water flux for the sandy topsoil was found to be about 20 times that for the clay topsoil. Outflow pressure also had a large effect on soil water pressures. At large outflow pressures, the soil-water pressures throughout the column were greater than zero. As the outflow pressure was decreased through increments the soil-water pressures were observed to decrease also.

Perhaps the most important result was the excellent agreement between observed and calculated soil-water flux rates and soil-water pressures. Thus, it is apparent that a two layer soil system can differ substantially from a monolayer system, but the flow and moisture regime can be accurately described by theoretical equations.

As was mentioned earlier, the physical properties of the soil are not the only factors affecting the movement of water through soil [*Quirk and Schofield*, 1955]. The chemistry of the soil and the permeating solution can have large effects on hydraulic conductivity; especially in soils that have a large percentage of clays. In their paper, *Quirk and Schofield* examined the effect of electrolyte concentration and ion species on the permeability of soil samples. The well accepted phenomenon of decreasing permeability with decreasing electrolyte concentration was studied by passing solutions of differing concentration of specific cations (Na^+, K^+, Ca^{++}, Mg^{++}), through soil pads 1 cm thick. The definition of a threshold concentration as that concentration of a particular cation that causes a 10-15 percent reduction in permeability was made to aid in the comparison of the effects of different ions on soil permeability. For the soil studied, the threshold concentration for a Na^+ solution was 2.5×10^{-1} Molar (M) and 2.0×10^{-4} M for a calcium solution: a difference of approximately three orders of magnitude. The threshold concentrations for a K^+ and Mg^{++} solutions were 6.7×10^{-2} M and 1×10^{-3} M respectively. The authors also studied the effects of mixed ion solutions on the permeability of soil samples at varying degrees of Na^+ - Ca^{++} saturation (exchangeable sodium percentage - ESP). The results of the experiment showed that as the ESP of the soil increases, the electrolyte concentration necessary to maintain a satisfactory permeability also increases. Also, the divalent cation concentration in the percolating solution seemed to be critical in affecting soil permeability, regardless of the ESP of the soil sample. In addition to these observations, the authors noted that at certain concentrations, a turbid percolate was obtained; indicating that deflocculation was occurring. The concentration at which turbidity occurred in the percolate was found to increase as ESP increased. Also, when turbidity was observed, high ESP samples yielded percolate solutions of much higher turbidity than low ESP soil samples.

In discussing their results, *Quirk and Schofield* explain that decreases in electrolyte concentration can cause decreases in permeability, as a result of swelling of clay particles resulting from repulsive forces acting within the particle itself. Also, as swelling increases, deflocculation occurs, which results in the breaking up of the clay particles. The "broken pieces" then flow with the percolating fluid before lodging in a pore and blocking flow. The authors also suggest that mechanical stress can add to this problem, causing the particles to deflocculate at a higher electrolyte concentration than if no stress were applied. The individual effect of swelling and dispersion on decreased permeability will be discussed later.

Further investigation of the effect of soil and water chemistry on the permeability of soil has led to many papers examining the different mechanisms involved. Some factors that influence the effect of saline water on soil hydraulic conductivity are given in a paper dealing with Texas soils [*Naghshineh-Pour, et al.,* 1970]. In this study soil characteristics and solution composition were related to hydraulic conductivity. The authors found that two soils high in montmorillonite clays exhibited pronounced decreases in hydraulic conductivity at high sodium adsorption ratios ($SAR = Na/(\frac{Ca + Mg}{2})^{\frac{1}{2}}$ concentrations in meq/ℓ) and low electrolyte concentrations. On the other hand, two soil samples low in montmorillonite clays were shown to be much less sensitive to solution composition. In addition, the authors found that for the high montmorillonite clays, higher soil ESP required substantially higher salt concentrations in the permeating solution to maintain a significant hydraulic conductivity in the soil. This finding agrees with the results of *Quirk and Schofield*. A related paper which also studied Texas soils [*Yoron and Thomas*, 1969] made similar conclusions. In this paper the authors found that as a high sodium water was leached through a soil sample, the sodium was adsorbed more rapidly in the top layers until equilibrium was reached. The authors also concluded that the hydraulic conductivity of the entire column was controlled by the mean ESP of the column rather than the ESP of the top layer.

Two papers by *B. L. McNeal, et al.*, [1966 and 1968] deal with factors affecting soil hydraulic conductivity. Several soils with differing mineralogies were studied by measuring hydraulic conductivity in the presence of solutions with varying SAR's and salt concentrations. All the soils reacted differently, yet all demonstrated decreasing hydraulic conductivity at high levels of sodium and low electrolyte concentration. On the other hand, solutions with SAR = 0 caused little variation in hydraulic conductivities even at very low salt concentrations. Also, as *Naghshineh-Pour, et al.*, reported, the presence of montmorillonite increased the sensitivity of a soil to low salt-high sodium solutions. On the other hand, soils containing kaolinite, amorphous materials or sesquioxides were found to be much more stable. As might be expected, the clay content was also found to affect the stability of a soil. A soil with a high clay content, when compared to one with a low clay content, showed a greater susceptibility to reductions in hydraulic conductivity in the presence of low salt-high sodium solutions.

In another paper by *McNeal, et al.*, [1966], the authors discuss clay particle swelling as a mechanism causing reductions in hydraulic conductivity. In this study, macroscopic swelling was measured as the amount of additional solution imbibed by a soil-sample after equilibration with a high salt solution. The results showed a strong correlation between observed macroscopic swelling and reductions in hydraulic conductivity. The authors attempted to explain the observed macroscopic swelling in terms of the theoretical interlayer swelling of the clay particle. Interlayer swelling values were based on a theory which assumes that the clay soil is divided into sodium saturated domains and calcium (or magnesium) saturated domains. The percentage of Na^+ saturated domains and Ca^{++} saturated domains depends on the saturation percentage of the ions in the soil-water system. When compared to the experimentally determined macroscopic swelling values, a good correlation (> .900) was observed for the four soils studied. Also, interlayer swelling regressed against relative hydraulic conductivities also showed a highly significant correlation. Based on these observations, the authors concluded that in situ mineral swelling resulting in the closing of water conducting pores is a plausible mechanism causing reductions in hydraulic conductivity in soils under the influence of high SAR and low salt solutions. The authors did, however, indicate that dispersion and the resulting blockage of water conducting pores could also act to decrease hydraulic conductivity; especially in loosely compacted soils where the movement of particle layers are not restrained by surrounding soil particles.

A paper which examines dispersion as the major mechanism resulting in hydraulic conductivity reductions was done by *Frenkel, et al.*, [1978]. The authors performed a study in which hydraulic conductivity was measured on montmorillonite, kaolinite, and vermiculite soils as they were leached with different electrolyte solutions. The ESP of the soils and the solution electrolyte concentrations were chosen to be representative of that found in the field. Hydraulic head was monitored at different depths throughout the soil column to determine what layers in the soil were limiting to water movement. In addition, clay content in the leachate was monitored to determine the amount of dispersed soil particles passing out of the soil. The results showed that for all soils the reduction in hydraulic conductivity was due to dispersion and subsequent plugging of soil pores. Hydraulic conductivity reductions were more pronounced at higher ESP values (20 to 30). Some plugging was observed at ESP of 10 and electrolyte concentration of 10 meq/ℓ for both kaolinite and montmorillonite; but vermiculite remained virtually unaffected. Leaching with distilled water, however, caused appreciable reductions in hydraulic conductivity at all levels of ESP. Also, the sensitivity of a soil to excessive exchangeable sodium and low electrolyte concentration increases with clay content and bulk density. Thus, the authors concluded that in the range of soil ESP's and solution salt concentrations normally found in the field, dispersion of clay particles and the resulting plugging of conductive pores is the dominant mechanism controlling reductions in hydraulic conductivity.

Powell, et al., [1969] also studied clay particle dispersion and its effect on soil permeability. In addition, the authors determined the effect of mechanical stress on the dispersion of soil particles. Their results demonstrated that turbidity, which indicates the amount of dispersed and displaced particles, increased with decreased salt concentration and increased ESP. This agrees with the results of *Frenkel, et al.* Also, *Powell, et al.*, concluded that the initial drop in hydraulic conductivity was due to clay particle swelling. Mechanical stress, however, can also have a profound effect on the degree of dispersion and the conditions under which dispersion will first occur. For example, the electrolyte concentration at which dispersion first occurs depends on the amount of mechanical stress applied. In the presence of mechanical stress, such as raindrop impact or actions of machinery, dispersion will occur at a higher salt concentration than if no stress were applied. When small amounts of stress are applied, the proportion of sodium saturates zones will have a large effect on the amount of dispersion that occurs. Finally, *Powell, et al.*, showed that under conditions of large mechanical stress, dispersion will occur even in the Ca^{++} saturated zones. Thus, significant reductions in hydraulic conductivity will result even at low ESP's when large dynamic stresses are applied to the soil.

When the flow of water through a soil is very small, and a concentration gradient exists within the soil water solution, then the movement of solutes may occur through diffusion. Mathematically, the quantity of solutes moving at any time can be described by Fick's Law given as:

$$q = -D \frac{dC}{dX}$$

In this equation q is the flux of solutes, expressed as moles per area per time; dC/dX is the concentration gradient where C is in moles/liter; and D is the diffusion coefficient, usually expressed as cm^2/day. *Peck* [1971] describes some of the factors important in the diffusion of solutes through soils. First, diffusion in soil will only occur through that portion of the soil cross-section occupied by water. Therefore, both the porosity and the moisture content will affect the rate of diffusion. The diffusion rate will also be affected by the tortuosity, or the diffusion path length. As the path length for a diffusing ion exceeds that of a straight line separation between two points, the diffusion coefficient, and hence the diffusion rate, will decrease.

Another paper on diffusion through a porous media [*Saxena, et al.*, 1974] related diffusion coefficients to the effective pore radius of a small matrix. The authors fit an exponential equation to observed data to obtain a relationship between the relative diffusion coefficient and the pore radius. The equation developed was given as:

$$K^1/K_o = 1 - e^{-0.13r^2}$$

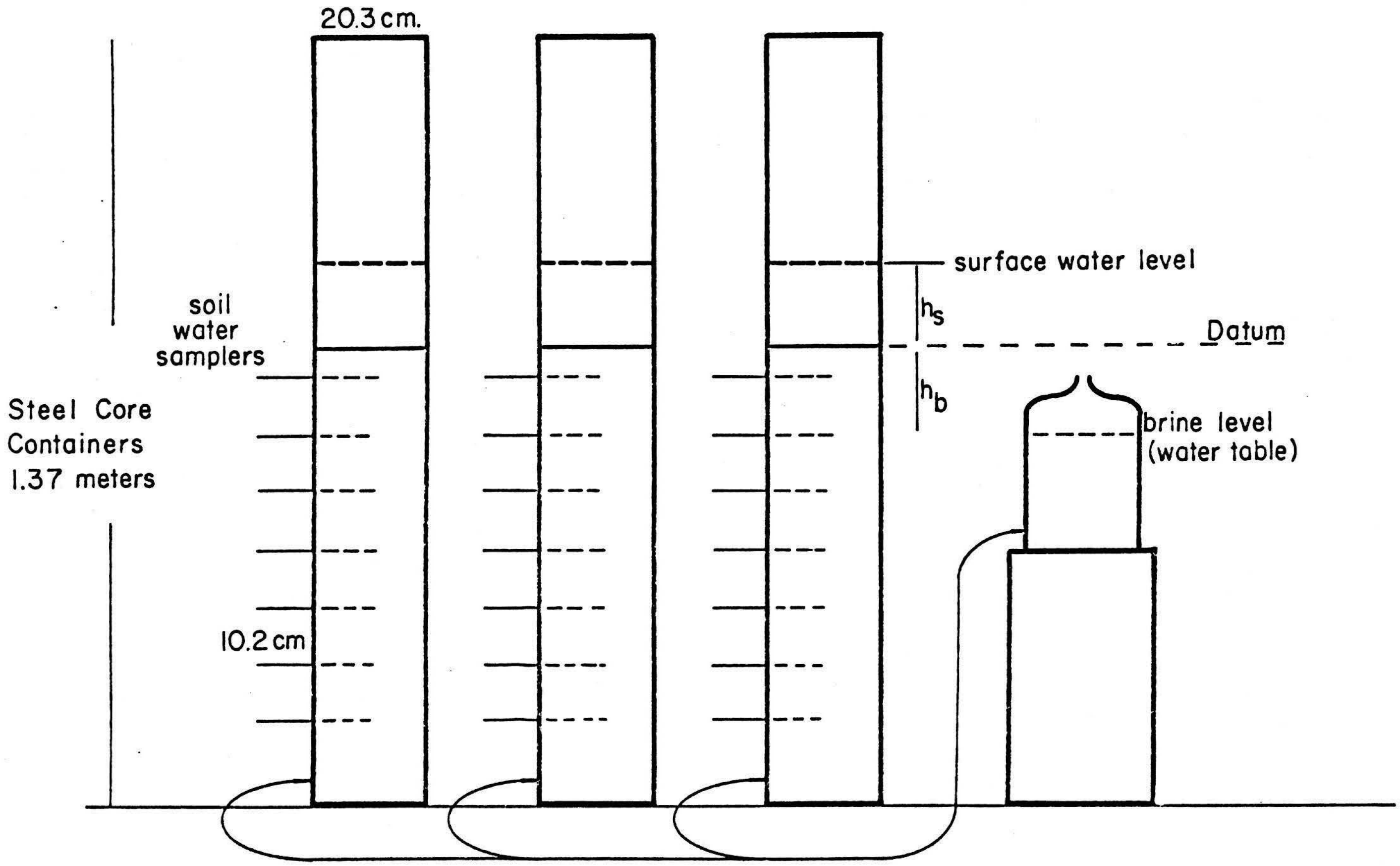
where K_o is the free solution diffusion coefficient, K^1 is the soil diffusion coefficient, and r is the pore radius.

Thus, from this equation, it can be seen that pore size can significantly affect the diffusion coefficient for the soil-water system.

EXPERIMENTAL THEORY

Fluctuations in salinity of soil water at a given soil depth can be produced by several mechanisms. Perhaps the major mechanism that could potentially act to move salt through the soil column is the movement of the soil water itself. Water, however, will only move through the soil media in response to a hydraulic potential difference. This potential difference could be created in the field under circumstances of high surface water level and low water table, or high soil water pressure and shallow surface water depth. As the soil water flows through soil pores it carries dissolved salts, thus causing changes in the interstitial water salinity. The magnitude of the salinity change depends on the flow rate of water and the salinity gradient existing within the soil.

To study the effects of surface and water table levels on soil water salinity, an experiment was designed using 20.3 cm (8 inch) diameter, undisturbed soil cores. The cores were placed in a hypothetical situation where surface water depth and water table level were maintained constant, while soil water salinity was monitored with depth, over a period of time. In theory, the observed salinity changes could be related to the rate of water movement through each core, which in turn could be related to the hydraulic potential difference between the surface water elevation and the water table elevation. A schematic of the experimental apparatus is shown in Figure 4. Water was ponded on the surface of each core to a depth h_s above the soil water interface. A water table was simulated by brine solution whose level was



SCHEMATIC OF EXPERIMENTAL APPARATUS

Figure 4. Schematic of experimental apparatus.

maintained at a level h_s below the soil water interface. Thus, a hydraulic potential was created, equal in feed of head to $h_s + h_b$, causing water to move through the columns. The amount of water that flows through the column is dependent on the magnitude of the hydraulic head and the properties of the soil. The direction of water flow depends, of course, on the elevation of the brine level relative to the surface water level. If the brine level is below the surface water level, then water will flow down through the column. The resulting change in salinity within the column is in turn related to the quantity and salinity of the water moving through the column. During the experiment, salinity was monitored by removing a small water sample from the soil through samplers placed at 10.2 cm (4 inch) increments along the soil column. By measuring EC on these samples, salinity fluctuations could be monitored in each layer over the course of the experiment.

To help interpret the results of the soil core experiment, a computer program was developed which would predict salinity changes in the soil column experiment, based solely on the mass transport of salts resulting from the movement of water. In short, the model was designed to predict the electrical conductivity (EC) at each sampling port in a column, on each sample day, based on the amount of water that flowed through the column between sampling days.

To develop the model the soil column was broken up into imaginary sections, as shown in the schematic diagram in Figure 5. For each section a water balance can be written as:

$$\text{Inflow} - \text{Outflow} - \text{Sample} = \text{Soil water storage} \qquad (1)$$

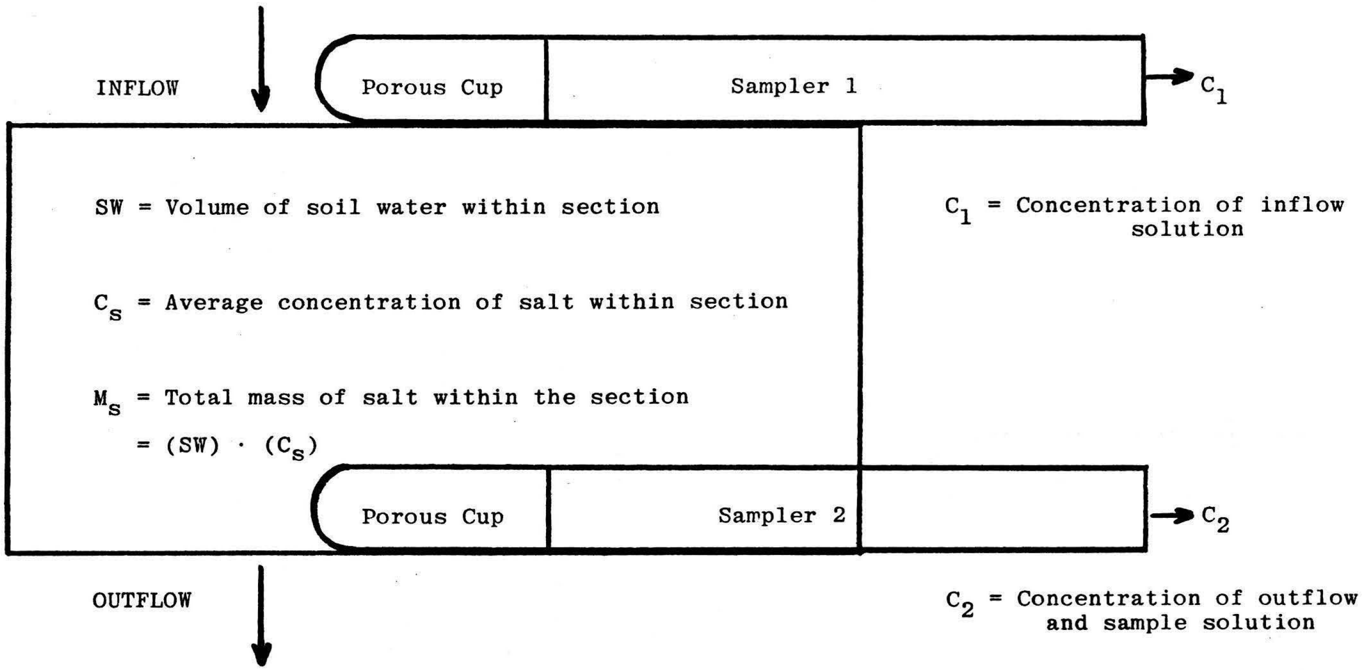


Figure 5. Schematic of soil core suction used to develop an equation for EC change.

where all terms are in units of volume. As shown in Figure 5, the sections were chosen so that a sampling port lay just above the section, and one lay at the bottom of the section. In this way, the salinity of the solution entering, as well as leaving, the section is known. Thus, if the salinity is expressed as a concentration, a mass balance of salt can be written around the section.

$$\text{Inflow } (C_1) - \text{Outflow } (C_2) - \text{Sample } (C_2) = \Delta M_s \qquad (2)$$

In this equation, C_1 and C_2 are the salt concentrations at samplers 1 and 2 respectively, while ΔM_s is the change in the mass of salt within the section. Because the salinity in this relationship is expressed as a concentration and the data taken in the soil core experiment is in terms of EC, a relationship was used which related EC to milliequivalents per liter [*U.S.D.A.*, Agriculture Handbook 60, 1954]. This was assumed to be valid over the range of EC encountered in the experiment. Also, because the cores were enclosed, changes in soil moisture were impossible to measure. Therefore, soil water storage was assumed to be constant over the course of the experiment. The above expression can now be written as:

$$F\ (C_1 - C_2) = SW \cdot (\Delta C_s) \qquad (3)$$

F, in this case, equals the total volume of water flowing through the section between sampling days (F = Inflow = Outflow + Sample). By defining the terms C_1, C_2, and ΔC_s in terms of average salt concentrations between sampling days (see Appendix II), Equation 3 was used to predict the salt concentration, and EC, at each sampling port over the course of the experiment, based on the flow of water through the column.

The major advantage to predicting salinity changes based on this simplified approach is that the input data is easily obtained. As explained in Appendix II, the data necessary for this model consisted of initial EC at each sampling depth, the surface water EC on each sampling day, the average flow of water per day through the column, the volume of sample taken from each sample port and the moisture content of each layer (measured at the end of the experiment). The limitations of this model include the fact that it assumes other mechanisms affecting soil water salinity have a negligible effect on the salinity fluctuations within the soil. These mechanisms include the cation exchange process, the precipitation and solubilization of salts in high concentration, and the diffusion and dispersion of salts in response to the salinity gradient within the soil column. Also, the model ignores changes in soil water storage which could have a major effect on the salt content in a particular section and the flow of water and salts into another section. A discussion of these assumptions, as well as a comparison of predicted and actual EC values, will be given later.

METHODS

Data were collected in three phases. In the first phase field data were collected over a seven month period, beginning in April 1978. The second phase involved monitoring salinity fluctuations in 20.3 cm (8 inch) diameter, 76.2 cm (30 inch) deep undisturbed soil cores taken from the field. The third phase consisted of determining the chemical and physical properties of the soil used in the soil core experiment.

Field Data

Collection of field data was begun in the Spring of 1977. Chemical data, electrical conductivity (EC), as well as flow data, was collected on a weekly basis to provide background information for further, more intensive study. Also, the data that was collected was used to determine what chemical and physical parameters were most important in terms of salinity and salinity fluctuations.

Beginning in the spring of 1978, field sampling was directed toward determining the factors affecting salinity fluctuations in the marsh system. A small sub-unit of the marsh (approximately 100 acres) was chosen as a study area; in which inflow, outflow, water level, and EC of the surface water was monitored on a weekly basis. Also, interstitial water sampling stations were established at nine locations within the sub-unit. At each station three wells were installed at 15, 38, and 61 cm (6, 15, and 24 inches) of depth, below the soil surface. Interstitial water samples, as well as surface water samples were taken and analyzed every other week for a period of 28 weeks. In addition, flow and EC were measured weekly along the western dike and at the inlet structure of Unit 1. It should be noted that all water samples were taken by technicians working through the Department of Wildlife Science at Utah State University. Figure 6 shows Unit 1 and the location of the study area (hereon referred to as the "study unit"). Also shown are the surface water sampling locations around the outer dike of Unit 1.

These sampling locations are also the locations of water flow control structures. Thus, all surface water EC data is associated with a flow of water for a specific date and station. Figure 7 is a detailed map of the study unit showing the locations of the interstitial water sampling stations. Interstitial water samples were collected on alternate weeks from surface samples. Also shown are the locations of water table monitoring wells established in October of 1978.

The following are descriptions of the specific procedures used for each step of the field sampling phase.

Flow Measurements

Water flow through most of the control structures was measured by use of a Marsh McBirney, Model 201, Portable Water Current Meter. All but two control structures consisted of circular steel culvert pipe 46 to 61 cm (1.5 to 2 feet) in diameter. Flow was determined by measuring the water velocity at three locations in the cross-section of the downstream end of the culvert pipe. The three velocity readings obtained were averaged and the resulting velocity taken as the average velocity through the pipe. The depth of water flowing through the pipe was measured and a cross-sectional area of water flow determined. Finally, the flow was calculated as the product of the cross-sectional

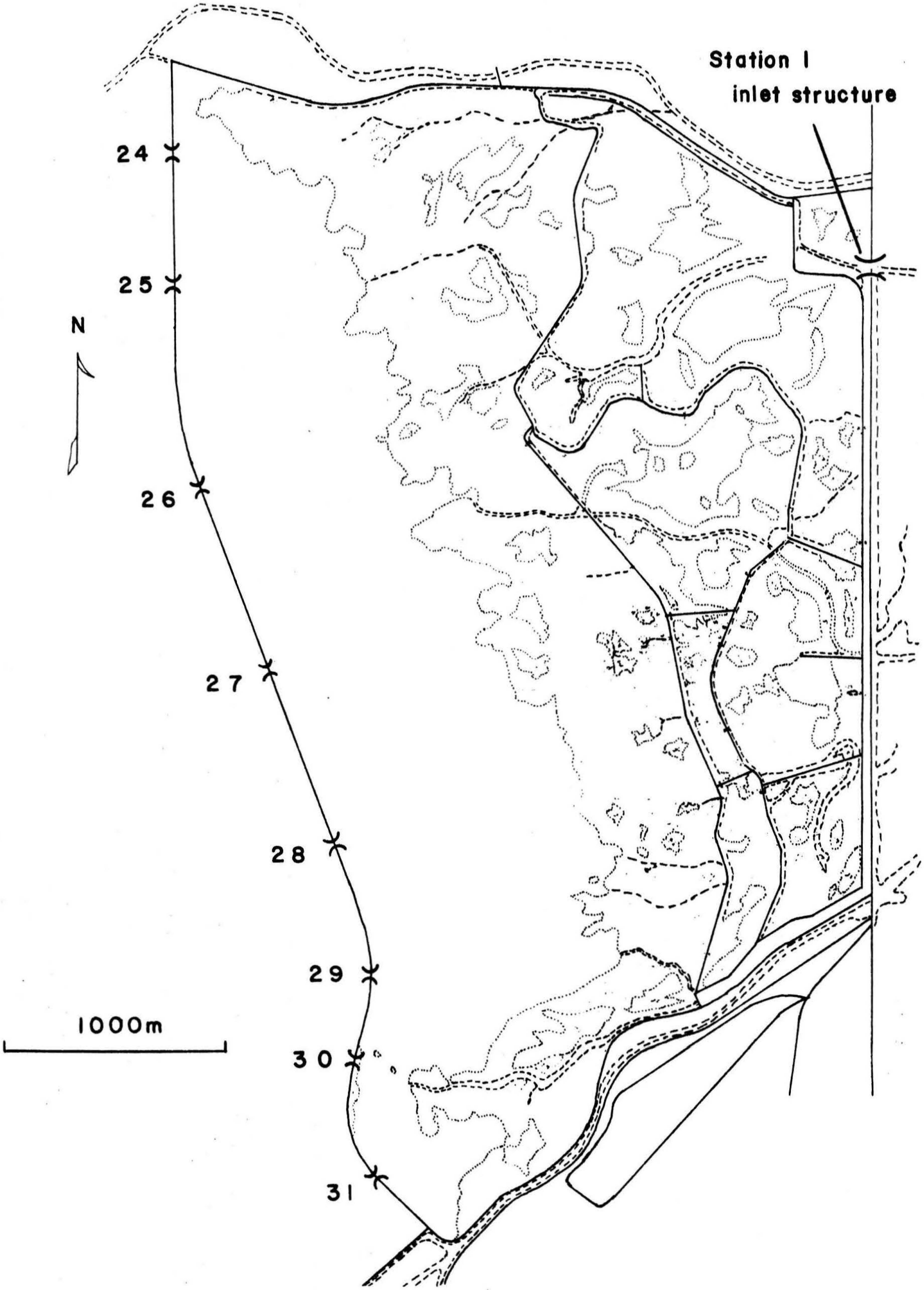


Figure 6. Map of Unit 1.

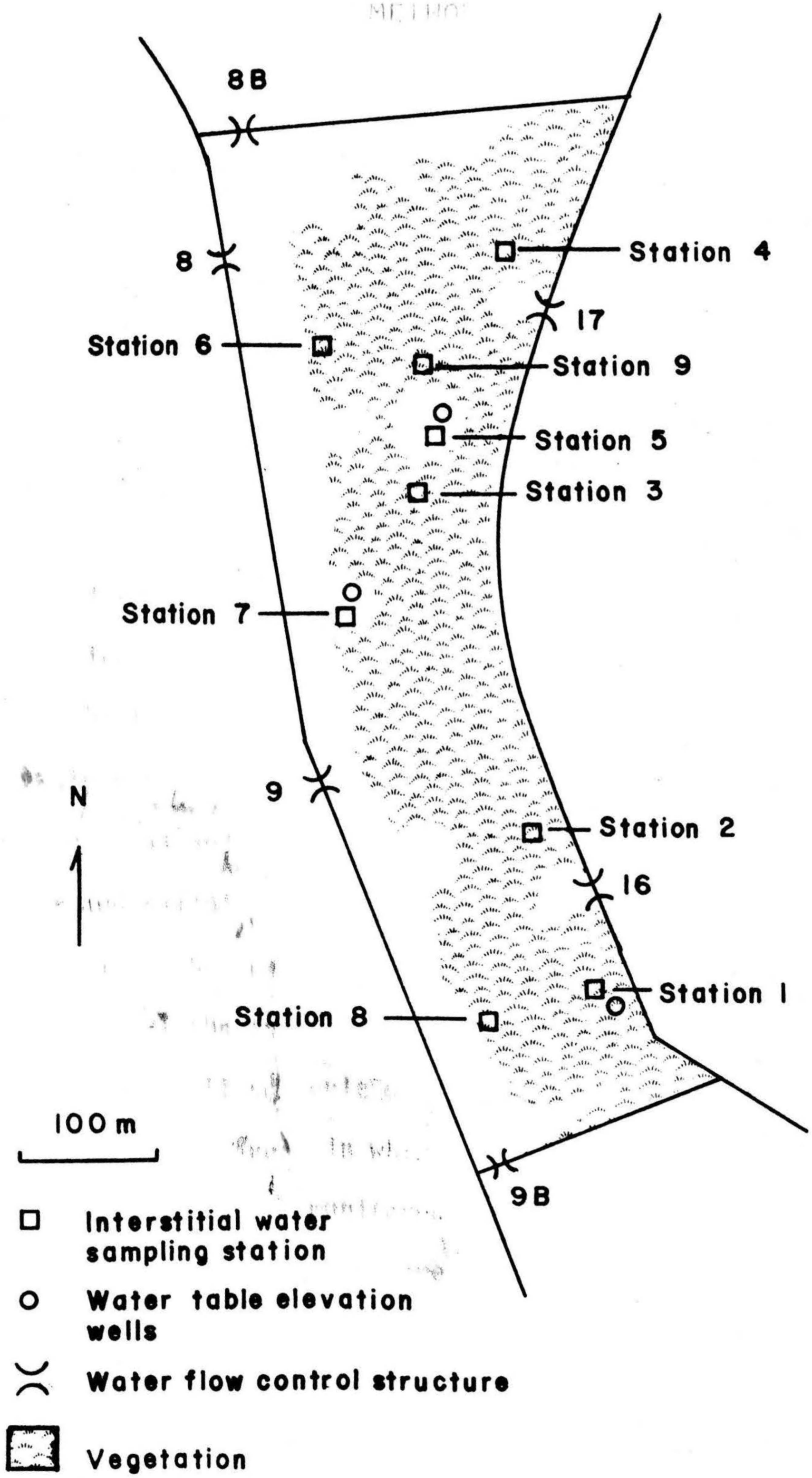


Figure 7. Map of study unit.

area and the average velocity.

The method described above was used to calculate flows at all stations except 25 and 26. Due to the type of structures at these locations, flow could not be measured with the velocity meter. These structures were constructed as weirs, with the amount of water flowing over the structure dependent on the elevation of water behind the weir. Thus, flow was measured using the equation for the flow of water over a sharpcrested weir.[1] This equation is believed to have yielded fairly accurate results at low flows. At high flows, however, the measured flows were probably low.

Flow was not measured at the inlet structure station. It was originally believed that flow at this point could be obtained from the water superintendent for the Weber River Basin. In mid-1978, however, it was discovered that this data was inadequate and could not be used.

Water Level Fluctuations

Surface water level in the study unit was measured by use of staff gauges set at each control structure. The purpose of the gauges was twofold. First, changes in water level from week to week could be monitored. Second, empirical relationships between discharge through the control structures and water level elevations could be developed to aid in future research efforts.

Water table monitoring stations were installed in late September

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[1] Equation for the flow of water over a sharpcrested weir: $Q = k \cdot 2g \cdot L(H)^{3/2}$, where Q = flow in cubic feet per second, L = length of weir (feet), H = height of water over top of weir, g = gravity = 32 feet/sec^2, $k = 0.40 + 0.05\ (H/P)$, P = height of weir above the bottom of the channel.

and monitoring was begun in October. The stations consisted of a 4 foot length of PVC pipe with the end set 60 to 76 cm (2 to 2.5 feet) below the surface of soil. A meter stick attached to a bottle was placed into tube and allowed to float on the free surface. Thus, by reading a meter stick at the top of the tube, the depth of the water table could be determined.

Electrical Conductivity Measurements

Electrical conductivity (EC) was measured in the field by use of a Yellow Springs Instrument, Model 33, conductivity meter. At all locations, including the interstitial water wells, temperature and EC were recorded. The field EC values were then corrected for temperature using the equation given in the 1975 edition of *Standard Methods*. Flow and EC data were then used to determine average inflow and outflow EC for the study unit and for all of Unit I.

Interstitial Water Samples

The interstitial water sampling wells consisted of PVC pipe fitted with a porous ceramic cup at the end. The pipe and cup were set at the desired depth and the open end covered with a rubber stopper. Every two weeks, EC was measured on the water that had drained into the cup. After EC had been measured the water was pumped from the pipe and the rubber stopper replaced.

Soil Core Experiment

As was discussed earlier, the soil core experiment was designed to determine how soil water salinity changed in response to different conditions of hydraulic head acting across the soil column. Thus, an apparatus was developed to allow the taking of large undisturbed soil cores from the field. These cores were brought back to the lab and fit with soil water samplers in order that soil water salinity could be measured. Finally, an experimental procedure was developed so that water movement, hydraulic head, soil water salinity and surface water salinity could be monitored over a period of several weeks.

The core containers themselves were 20.3 cm (8 inch) diameter steel well casing, cut to 1.37 meters (4.5 feet) long lengths. The wall thickness was approximately 0.3 cm thick. Holes of 2.5 cm (1 inch) diameter were drilled through the casing at 10.2 cm (4 inch) increments along the bottom 76 cm (2.5 feet) of the casing. These were used to insert the soil water samplers. While taking the soil samples in the field, these were covered with duct tape to avoid loss of soil or water. The inside of the core casing was sanded and then painted with an epoxy seal. This seal minimized the steel from rusting during the experiment.

One end of each core was beveled to create a sharp cutting edge for driving into the soil. The other end was fitted with steel brackets. The brackets contained holes through which a large bar could be inserted for lifting and driving the containers into the ground. Aluminum caps were made for each end. The top caps were designed to make an airtight seal so that suction could be maintained when removing the core from the surrounding soil. The bottom caps were designed so that a water tight seal would be created, thus allowing the cores to be transported and stored without leakage. These bottom caps consisted of a 1.3 cm (0.5 inch) aluminum plate with a groove into which the bottom edge of the core casing fit. This groove was filled with a rubber O-ring and caulking putty to create a seal. The bottom plate and the casing were bolted together using tie rods between brackets welded to the casing and to the plate.

The soil cores were taken from the field in the following manner. A site was chosen where only a few inches of water covered a bare soil surface. A bare surface was chosen so as to avoid changes in soil permeability during the experiment resulting from decaying vegetation. The core casing was placed upright on the soil (beveled edge down) and a bar inserted through the brackets. The casing was driven into the soil by four people applying pressure to the bar. When the casing had been pushed into the soil to the desired depth, the top cap was placed on the open end of the cylinder and the surrounding soil was dug away from the core. The casing and soil were then lifted out of the ground and placed on the bottom cap. The cap and casing were then bolted together. This procedure was followed for all three cores. It was not possible to get exactly the same depth of soil in each core. Core 1 contained 79 cm, Core 2 contained 66 cm, and Core 3 contained 74 cm. After sampling, the soil cores were brought back to the Utah Water Research Lab where the water samplers were installed.

Each soil water sampler consisted of a small porous cup, 2.2 cm in diameter and 5.6 cm long, connected to a 15.2 cm piece of PVC pipe, as shown in Figure 8. Plexiglass tubing connected to a sample bottle was then inserted through the PVC tubing and into the porous cup. The samplers were installed by first removing the duct tape covering each hole and then using a drill bit to remove 13 cm of soil from within the soil core. The sampler was then inserted into the core. A

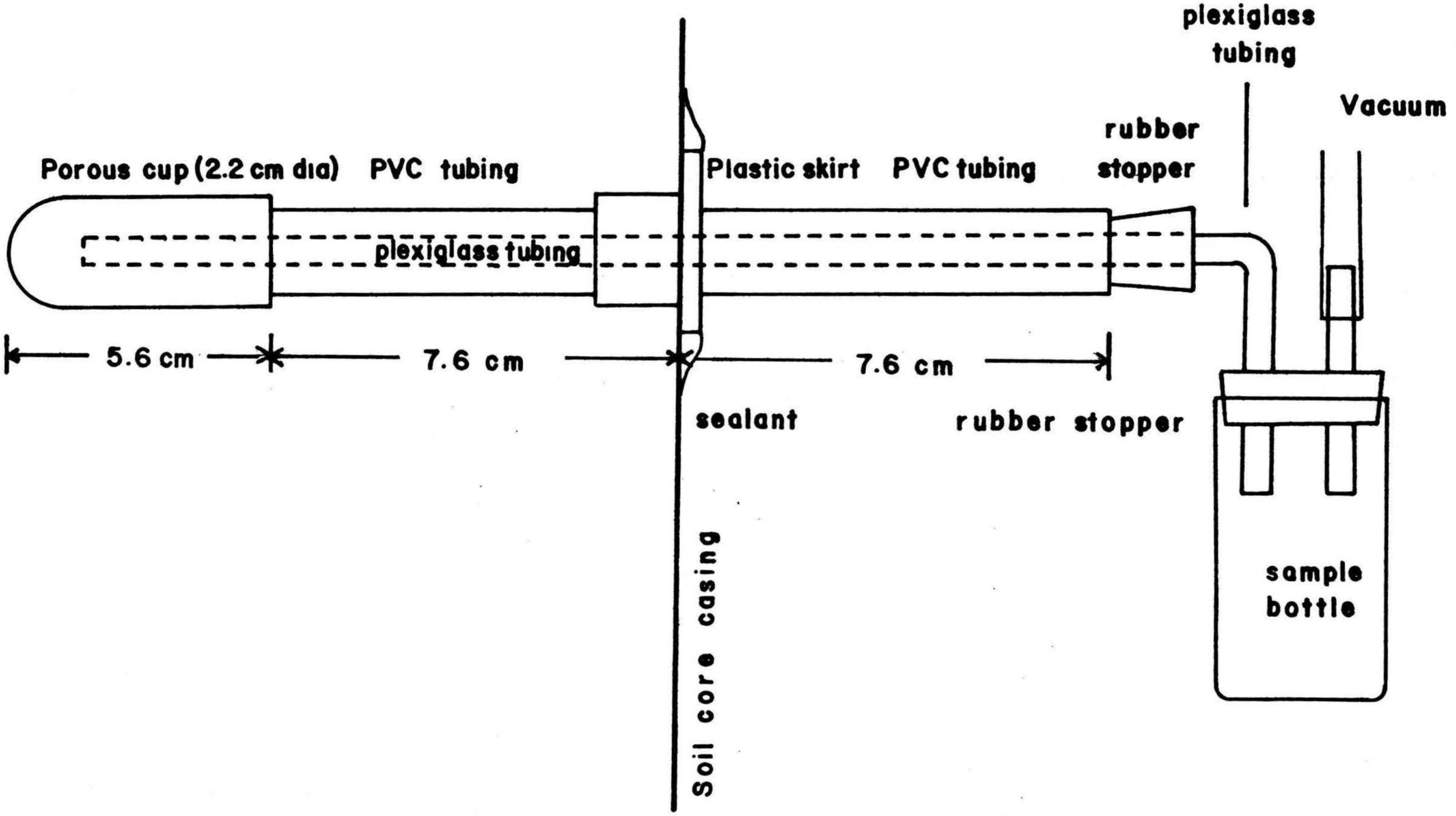

Figure 8. Diagram of soil water sampler.

caulking putty and a silicon sealant were used to seal the sampler in place. In addition to soil water samplers, each core also had a bottom port which allowed water to flow into or out of the core from the brine reservoir. This port consisted of a 25.4 cm (10 inch) length of PVC pipe with small holes drilled along the length of the pipe to allow for free water movement into or out of the soil. Nylon mesh was wrapped around the tube to prevent soil particles from falling into the pipe.

The brine solution used to simulate the water table was made up to be similar to that found in the field at 76 cm of depth. Also, the surface water solution was made up to approximately duplicate the chemical makeup of the surface water found in the field. Because each core contained a slightly different amount of soil, the cores were adjusted such that the soil surfaces in each column were at a common elevation relative to the brine and surface water levels. For the first set of conditions (Phase I), 46 cm (18 inches) of surface water was ponded on each core, while the brine level was maintained at 61 cm (24 inches) below the soil surface. Thus, a hydraulic head of 107 cm (3.5 feet) existed across each soil column. In Phase II, the levels were changed. The surface water was maintained at 15.2 cm (6 inches) above the soil surface, while the brine level was maintained at 15.2 cm (6 inches) below the soil surface. In this case 30.5 cm (1 foot) of head existed across the soil columns. Each set of conditions were maintained for 30 days.

During each phase, soil and surface water salinity was monitored by taking small soil water samples from the samplers in each core. A small hand vacuum pump was used to create a suction to draw a sample into the sample bottle. Generally, 40 cm of mercury was used as a suction pressure. This yielded 5-20 mℓ of sample in 2 hours or less. An effort was made to keep sample sizes as small as possible so as to avoid major disruption of the flow patterns within the soil columns. Electrical conductivity was measured on each sample using a YSI Micro-EC Probe. The Micro-EC Probe allows measurement of conductivity on samples as small as 5 mℓ.

Initially, during Phase I, samples were collected and analyzed three times a week. However, this was cut back to twice weekly during the final stages of Phase I and for Phase II. The change was made to further alleviate disruption of the water movement through the soil. In addition to soil water samples, the EC of surface and brine water samples was also determined. Water samples collected at the beginning, end, and once during the middle, of each phase were saved until the end of the experiment, when concentrations of Na^+, Ca^{++}, and Mg^{++} were determined. Other samples that were taken were discarded after EC was determined.

Between sampling times, the sampling bottle and tube were removed from the sampler and a rubber stopper inserted into the PVC tube. This procedure was enacted after the first three sampling periods of Phase I, when it was found that significant drainage into the bottles was occurring between sampling days.

Water movement through the cores was monitored by recording changes in elevation for each of the surface waters and for the brine reservoir between sampling periods. These elevation changes were then converted to volumes. Also, soil water sample volumes were recorded for each sampling period. After sampling, the elevations of all surfaces were brought back to their original level. During Phase II, the measuring procedure for the surface waters was changed by installing constant head reservoirs (inverted graduated cylinders) to feed water to the surface water as the level dropped. These were designed to maintain a constant water elevation and allow one to read the change in volume from the drop in water elevation within the graduated cylinder. In Phase I the tops of all the cores were covered with aluminum foil to minimize evaporation. In Phase II, the tops could not be covered, but evaporation was monitored on a reservoir containing a similar solution as that found in the surface waters.

In general, during any one sampling period, the data collected included the EC of the surface water, soil water, and brine solution; as well as soil water sample sizes and changes in volume for the surface waters and the brine reservoir. It should be noted that during Phase II leaks developed around the soil water samplers, making measurements of water movement difficult.

Soil Characterization

Upon completion of the soil core experiment, Core 1 was dismantled and the soil removed in 3 inch thick layers. Samples from each layer were analyzed to determine bulk density, saturated hydraulic conductivity, and soil texture. In addition, hydraulic conductivity versus water content curves were generated for each layer by use of a pressure plate apparatus similar to that used by *Gardner* [1956]. Because the three soil cores were taken from the field, in close proximity to each other, they were assumed to have the same general soil profile. Thus, the results of the soil characterization for Core 1 were assumed to be roughly applicable to the soil in Cores 2 and 3.

To generate the hydraulic conductivity versus soil moisture curves (K versus θ), 7.6 cm (3 inch) diameter, 7.6 cm (3 inch) thick, undisturbed soil samples were taken from the core at each layer. The samples were taken by pressing a 7.6 cm diameter, 7.6 cm tall, aluminum ring into the soil. The surrounding soil was removed and the ring and soil carefully pulled from the soil core and placed in a small airtight container. Another sample was taken in a similar manner using another 7.6 cm (3 inch) diameter ring, 15.2 cm (6 inches) tall. This ring was pressed into the soil 7.6 cm and removed and stored as described above. The small ring was used in a pressure plate apparatus, while the sample in the taller ring was used to determine saturated hydraulic conductivities for the soil in each layer.

The pressure plate apparatus consists of a 600 mℓ glass funnel with a fritted glass plate set in the bowl of the funnel. A cap fits over the funnel to create an airtight seal. A plastic hose is connected to the cap and runs to a pressure regulator. To operate the apparatus, the soil sample, in the 7.6 cm tall aluminum ring, is allowed to soak until saturated with a 0.01 M solution of $CaSO_4$ and is then placed on the fritted glass plate. The cap is then attached to the glass funnel, and the pressure applied by adjusting the pressure regulator. The amount of pressure applied is determined by the deflection of a mercury manometer. In theory, the pressure difference existing between the inside of the bowl and the outside of the glass plate acts as a suction causing water to be pulled out of the soil sample. The data collected consists of the volume of water recovered from the sample at each pressure applied. The pressure increments used in this experiment were 25, 50, 75, 100, 150, 200, 300, and 15,000 cm of water. After the last pressure increment, the sample was removed and a bulk density was determined.

To develop a hydraulic conductivity versus moisture content curve, the saturated hydraulic conductivity must be known. This was determined using the soil samples contained in the 15.2 cm tall aluminum rings. In this procedure, the soil samples contained in the rings were allowed to become saturated with a 0.01 M solution of $CaSO_4$. The rings and soil were then placed upright in a Buchner funnel and the 0.01 M $CaSO_4$ solution was ponded over the surface of the soil and the depth recorded. The time necessary for 10 to 15 mℓ of water to move through the sample was recorded. This data was then used to calculate the saturated hydraulic conductivity (K_s) using Darcy's Law.

The pressure plate data and the saturated hydraulic conductivity were used to develop a hydraulic conductivity versus soil moisture curve for each layer. The method used to accomplish this is similiar to that used by *Jackson* [1965]. The soil texture of each layer was determined by a hydrometer analysis. The procedure used was similar to that outlined by *Miller* [1978].

RESULTS

The results of the soil core experiment will be presented as well as the predicted results from the computer simulation (the observed data and the computed FORTRAN values are presented in Appendices I and II, respectively). These results will be explained in terms of the experimental procedures involved. Next, some physical properties of the specific soil layers found in the soil cores studied will be presented. Also, some of the chemical characteristics of the soil and surface water sampled will be discussed. Finally, some significant results from the field investigations will be presented along with a description of some of the seasonal changes that occurred in the marsh system.

Soil Core Experiment

During Phase I, EC was observed to decrease at all depths in each core (Figures 9, 10, 11). Prior to day 9 of the experiment, the sample bottles were left connected to the soil water samplers in the core between sampling days. During this time significant drainage of soil water into the bottles, was observed; particularly in the top and bottom samplers of the soil columns. This prompted concern that the normal flow of water through the column was being disrupted by this drainage. After day 9, rubber stoppers were inserted into the PVC tubes between sampling days to eliminate this drainage. This may account for some of the leveling off of the EC versus time plots after day 9. Water movement into the cores (as measured by changes in surface water elevation) was determined on an average volume per day

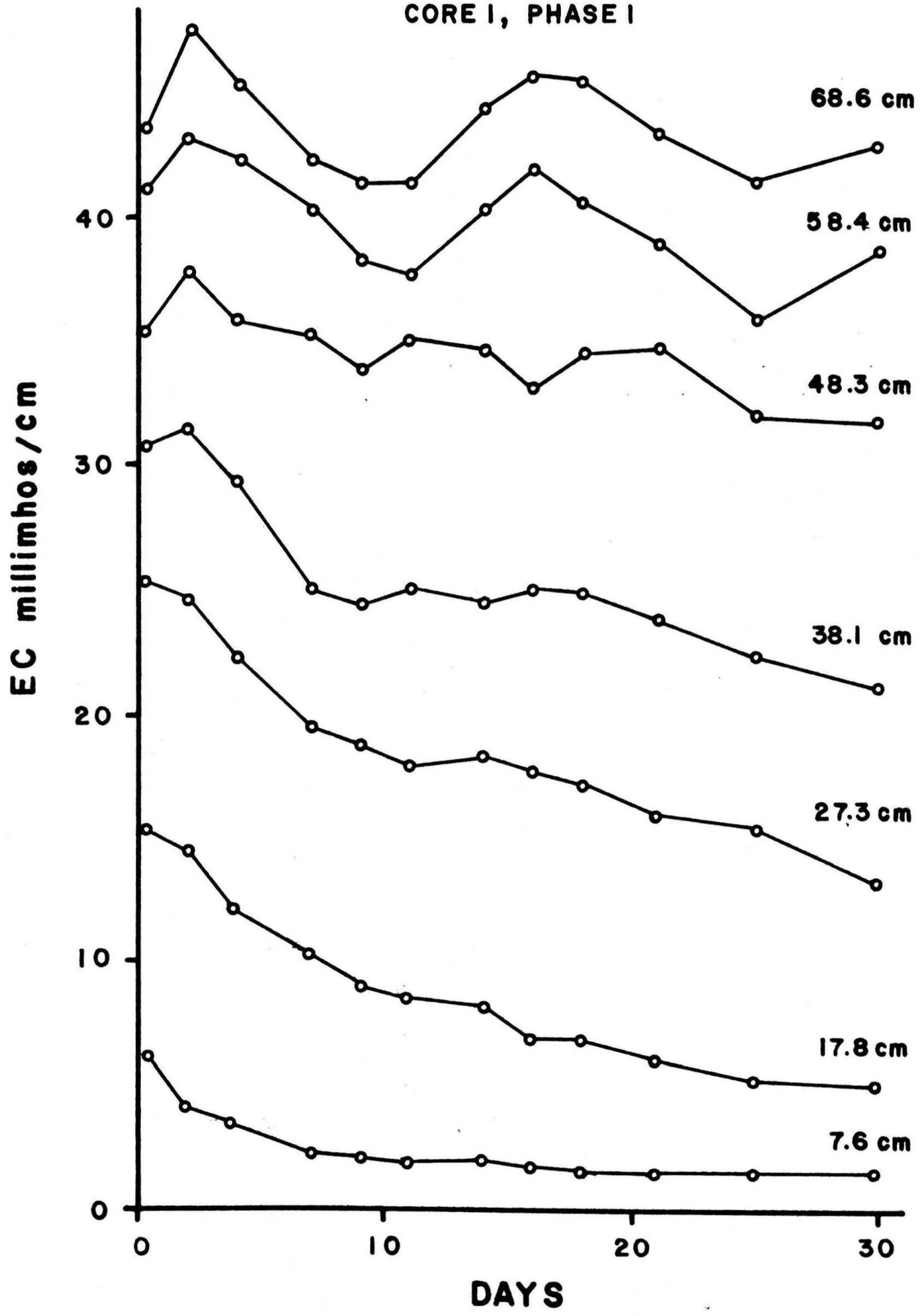

Figure 9. Change in EC with time at each sampler depth, Core 1, Phase I.

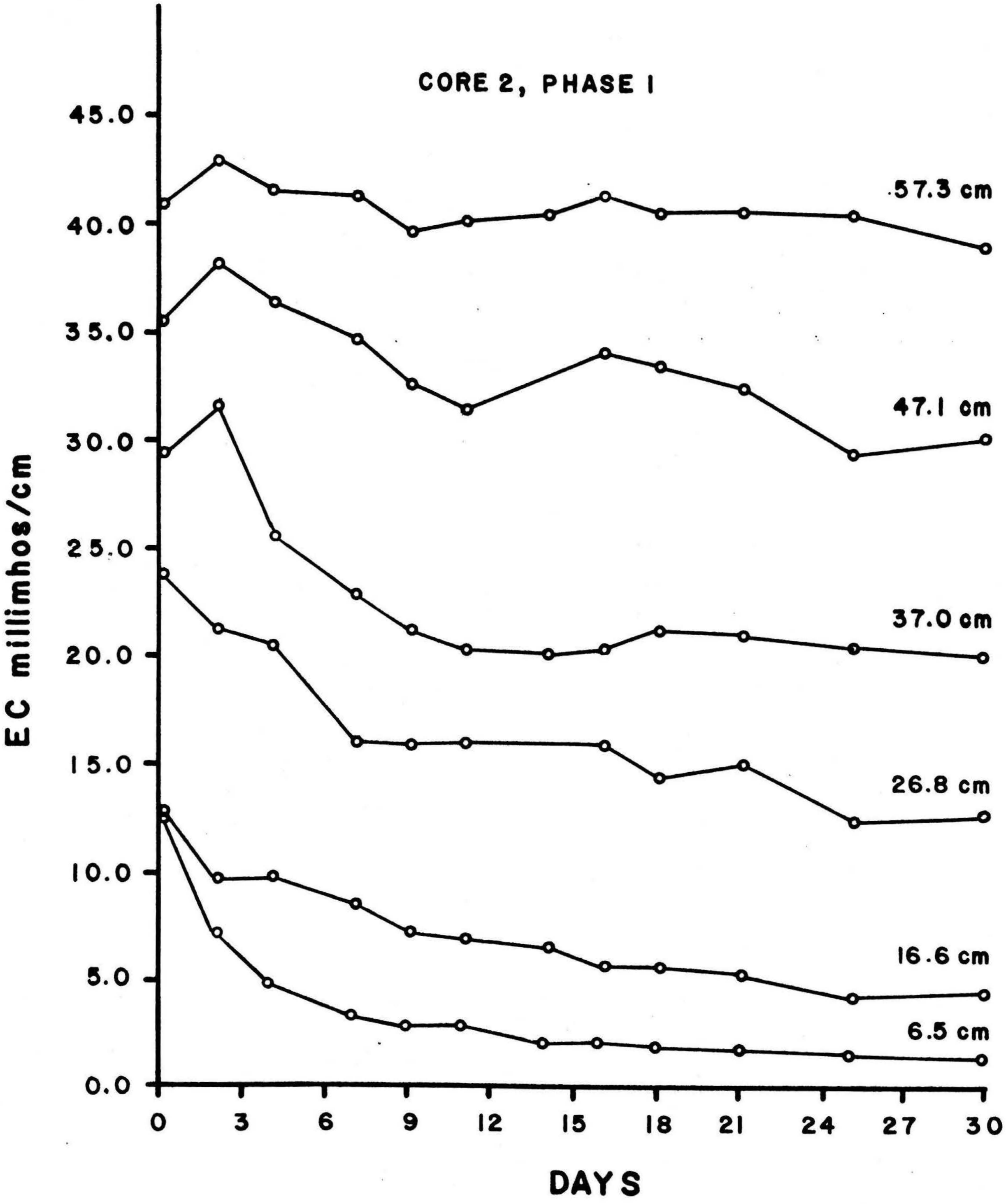

Figure 10. Change in EC with time at each sampler depth, Core 2, Phase I.

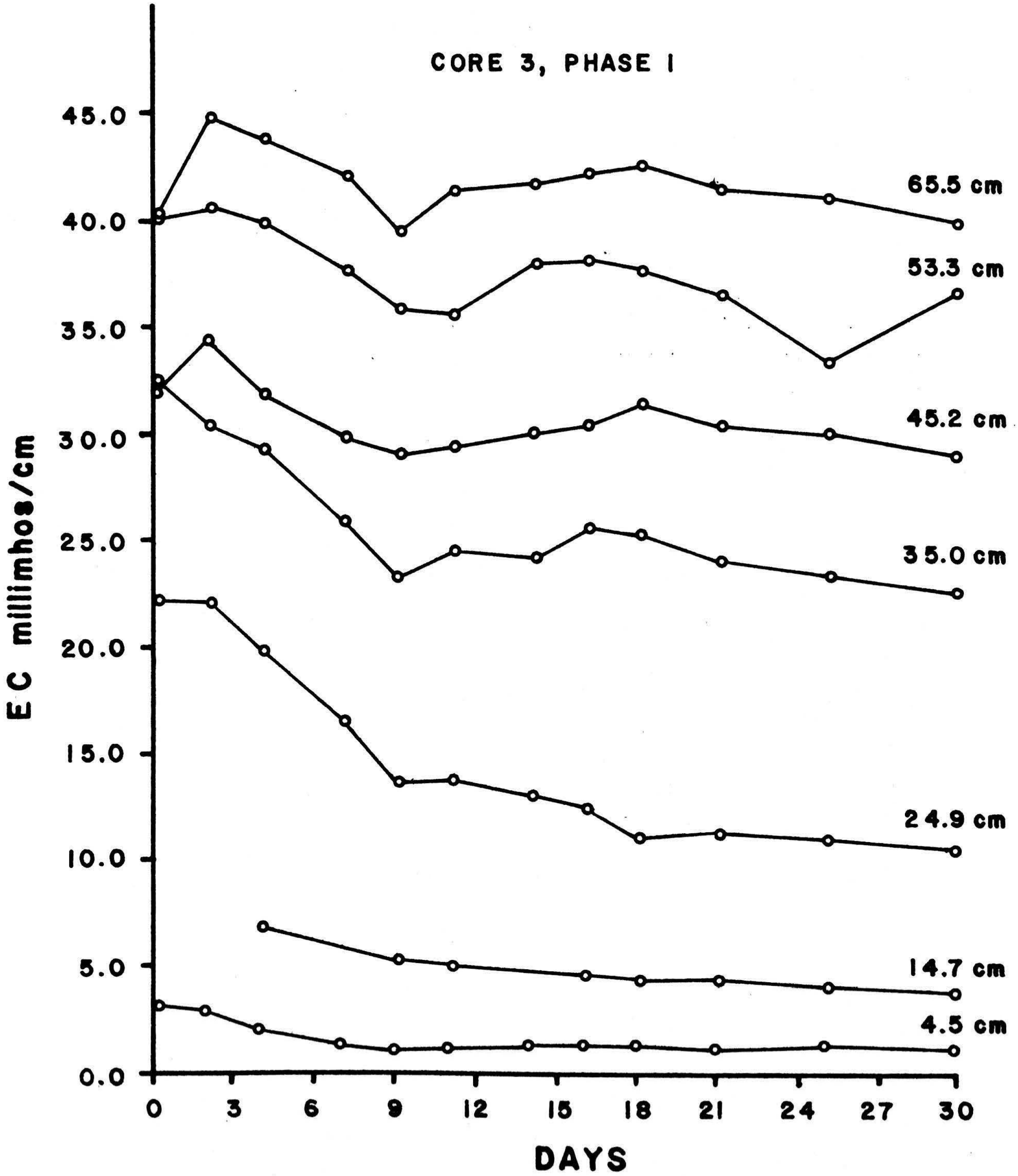

Figure 11. Change in EC with time at each sampler depth, Core 3, Phase I.

basis, before and after day 9. Table 2 shows these values for Phase I and Phase II.

In contrast to the large drop in EC at each depth for Phase I, salinity was observed to fluctuate very little during Phase II (Figures 12, 13, 14). This is consistent with the fact that much less water was moving into the column in response to the smaller hydraulic potential than in Phase I. This is indicated in Table 2 by the much smaller flow per day values observed in Phase II. These numbers have been adjusted for evaporation.

Unfortunately, several problems were encountered in the operation of the experimental system during Phase II. First, because the movement of water into the column was much less than in Phase I, the volume of samples taken became much more significant. Even though the number of sampling days were reduced from 12 in Phase I, to 9 in Phase II, the total volume of water taken in samples often was larger than the volume of water observed to move into the column from the surface water, between sampling days. As a result, water was forced to flow from the brine reservoir into the core, during sampling periods, to

Table 2. Flow into column from surface water (mℓ/day)

	Phase I		
	Before Day 9	After Day 9	Phase II
Core 1	127.5	61.0	7.5
Core 2	127.5	64.0	14.0
Core 3	127.5	72.0	8.0

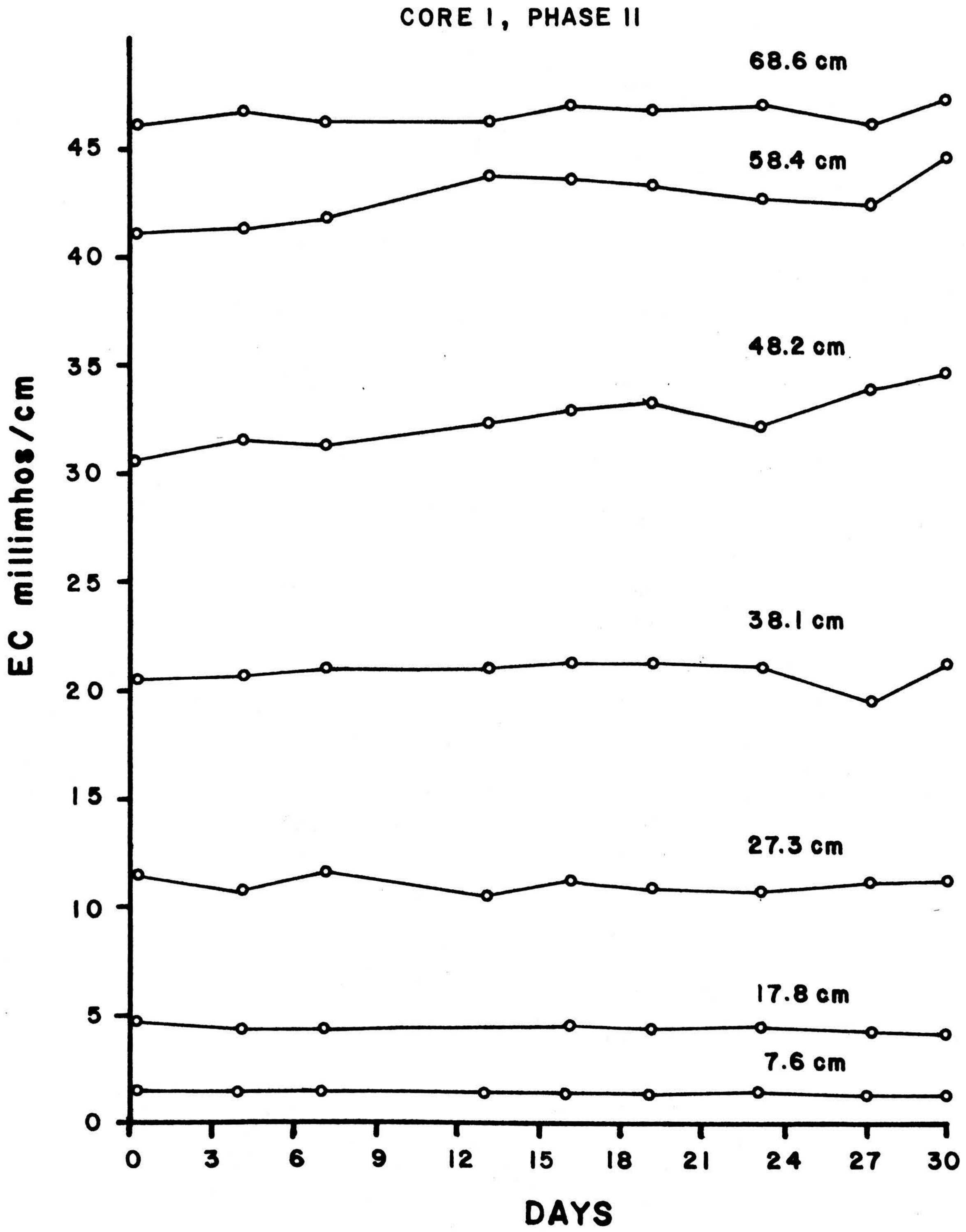

Figure 12. Change in EC with time at each sampler depth, Core 1, Phase II.

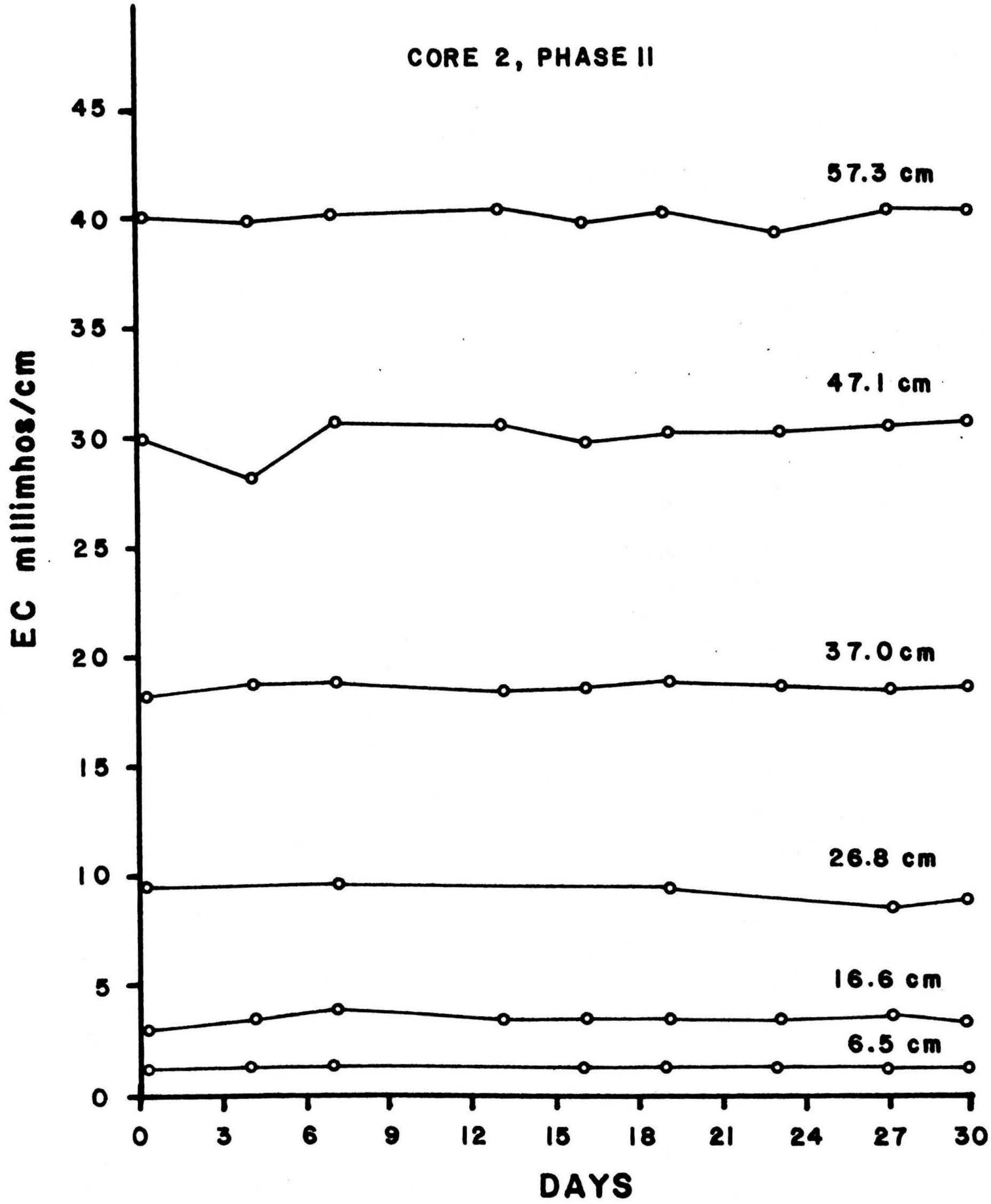

Figure 13. Change in EC with time at each sampler depth, Core 2, Phase II.

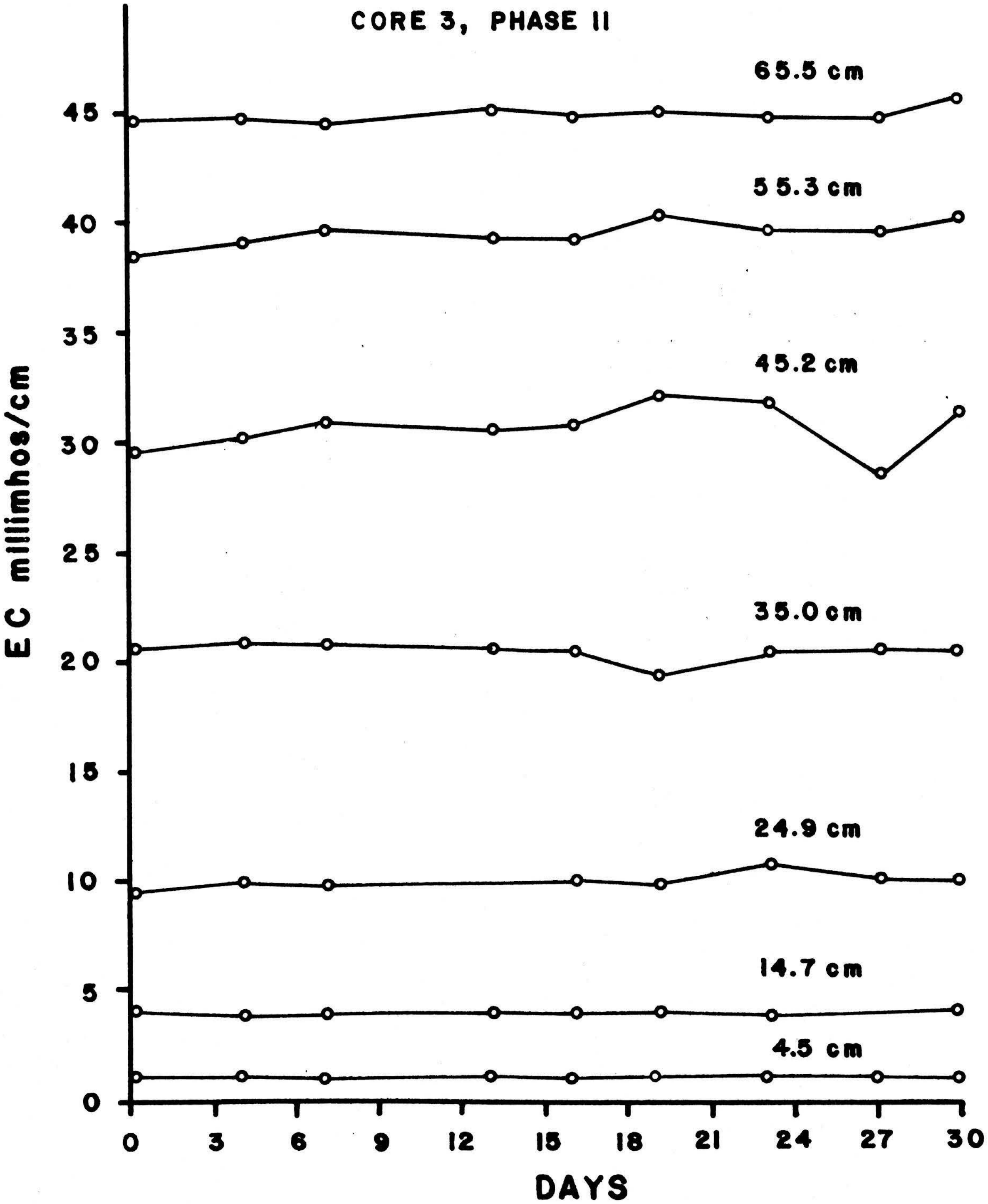

Figure 14. Change in EC with time at each sampler depth, Core 3, Phase II.

make up the difference. This was evidenced by a small drop in the brine water elevation during the sampling procedure. The second operational problem encountered was that of leakage which occurred around the sampling tubes. The resulting loss of water was difficult to quantify. However, in almost all cases, it amounted to no more than a few milliliters per day. Also, all the leaks occurred around the bottom 2 or 3 sampling tubes on each core. As a result, water was drawn from the brine reservoir into the bottom layers of the core, to replace this lost water. This too was evidenced by a drop in the brine water level between sampling days.

These problems are significant because they disrupted the flow of water through the column. Fortunately, these problems were minimal in Core 2. As a result, Core 2 came the closest to operating in the manner desired. In spite of the operational problems, only small EC changes were observed in any of the three cores.

Surface water EC was also monitored during the experiment. Figures 15 and 16 show how surface water EC varied during Phase I and Phase II respectively. In general, surface water EC was observed to increase in all cores, over the duration of the experiment. In comparison to the magnitude of changes that occurred in the soil water, particularly during Phase I, the changes in EC that occurred in the surface water are rather small. It should be noted, however, that because of the large volume of water ponded on each surface (14.8 liters in Phase I, 4.9 liters in Phase II), even a modest increase in salt concentration represents a fairly significant increase in salt content.

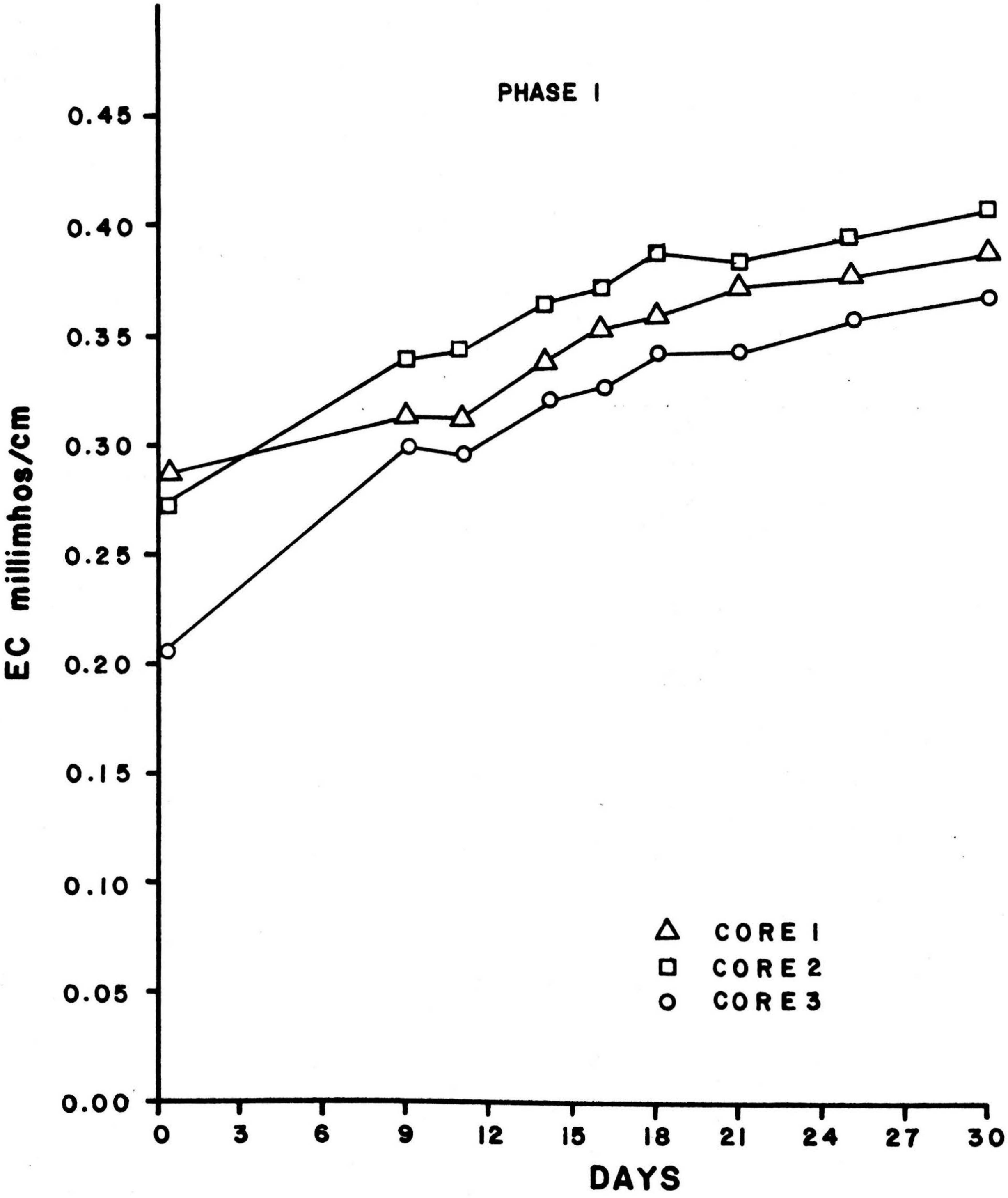

Figure 15. Change in surface water EC with time, Phase I.

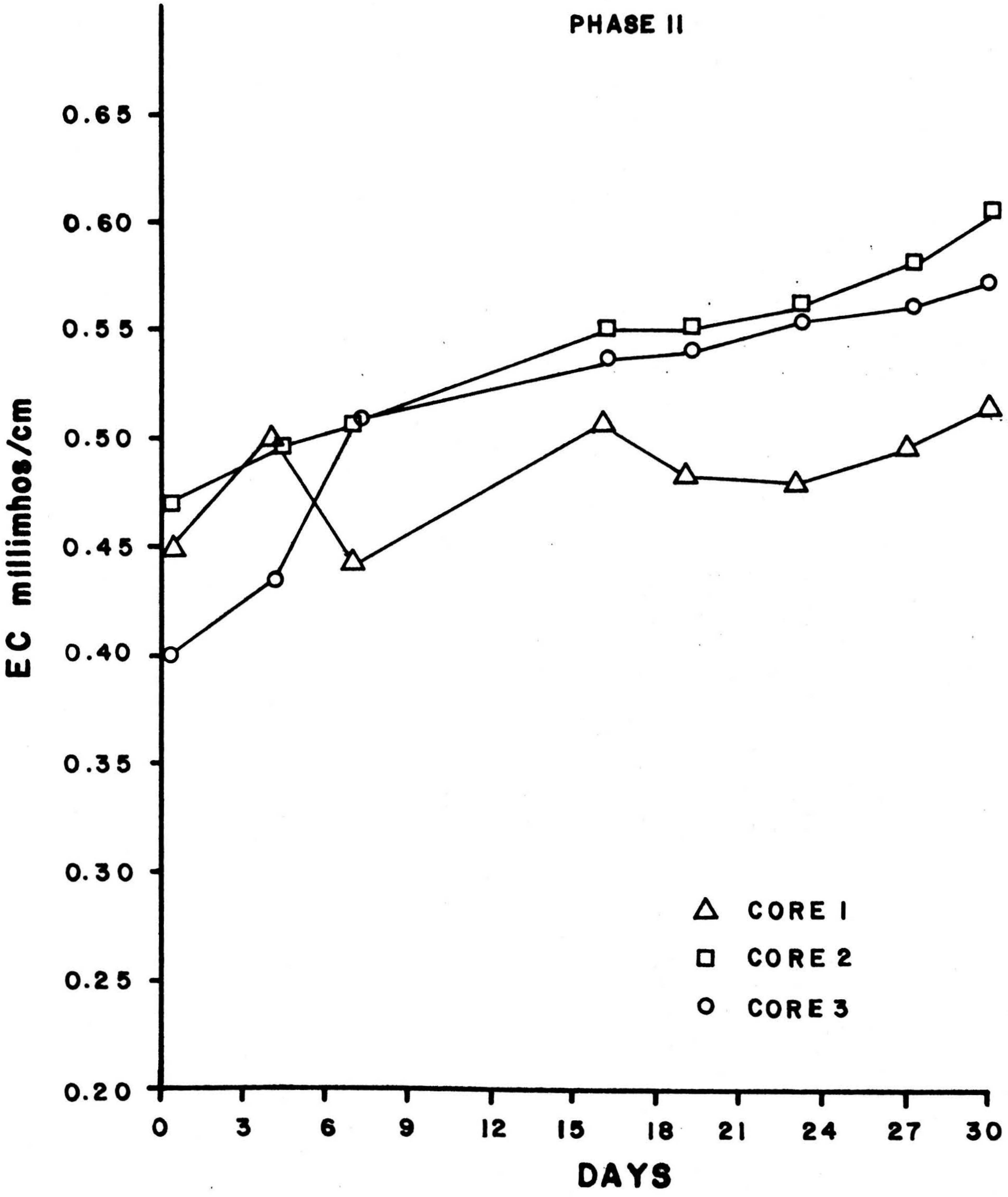

Figure 16. Change in surface water EC with time, Phase II.

Predicted Results

The predicted changes in EC with time are shown in Figures 17, 18, and 19, for Phase I, and Figures 20, 21, and 22, for Phase II. These predicted results are based on the flow of water into the core, as given in Table 2. For comparison the observed data is also plotted on the same axis.

In general, the model predicts a decrease in EC, in response to the flow of water through the column as shown in the figures for Phase I. For Phase II, on the other hand, the model predicts little change in EC, which is consistent with the observed results.

For Phase I, a flow of 127.5 mℓ/day was used for each core, until day 9 when the model switched to the smaller flow. This is consistent with the change in experimental procedure discussed earlier. Also, the model used day 2 as the initial conditions. The reason for this was that the sudden change in hydraulic loading, from the conditions in the field to those used in Phase I, caused rapid and varied changes in soil water salinities. The model was not designed to handle these changes. Thus, to enable the model to give better results, day 2 was chosen for initial conditions.

For Phase II, the model was designed to simulate the experimental operation, including the major operational problems. Included in the program was a sequence which simulated the brine solution being pulled into the core when insufficient water had moved into the core between sampling days. Included in the sequence is an added flow of 10 mℓ from the brine reservoir, which simulated leakage from the bottom sampling ports. In this way the model was able to predict changes in EC resulting from both downward movement of surface water and the upward flow of brine water.

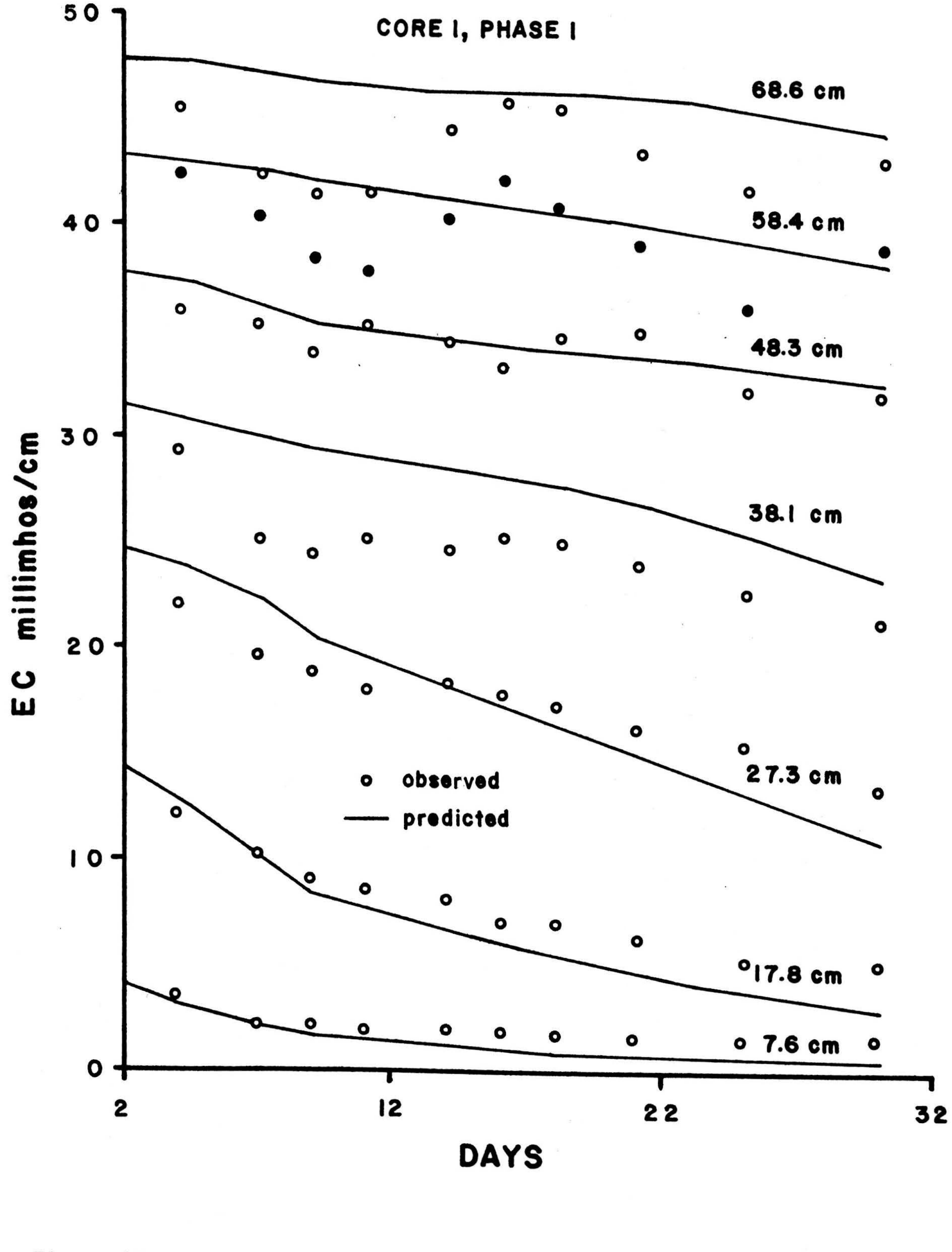

Figure 17. Predicted and observed changes in EC, Core 1, Phase I.

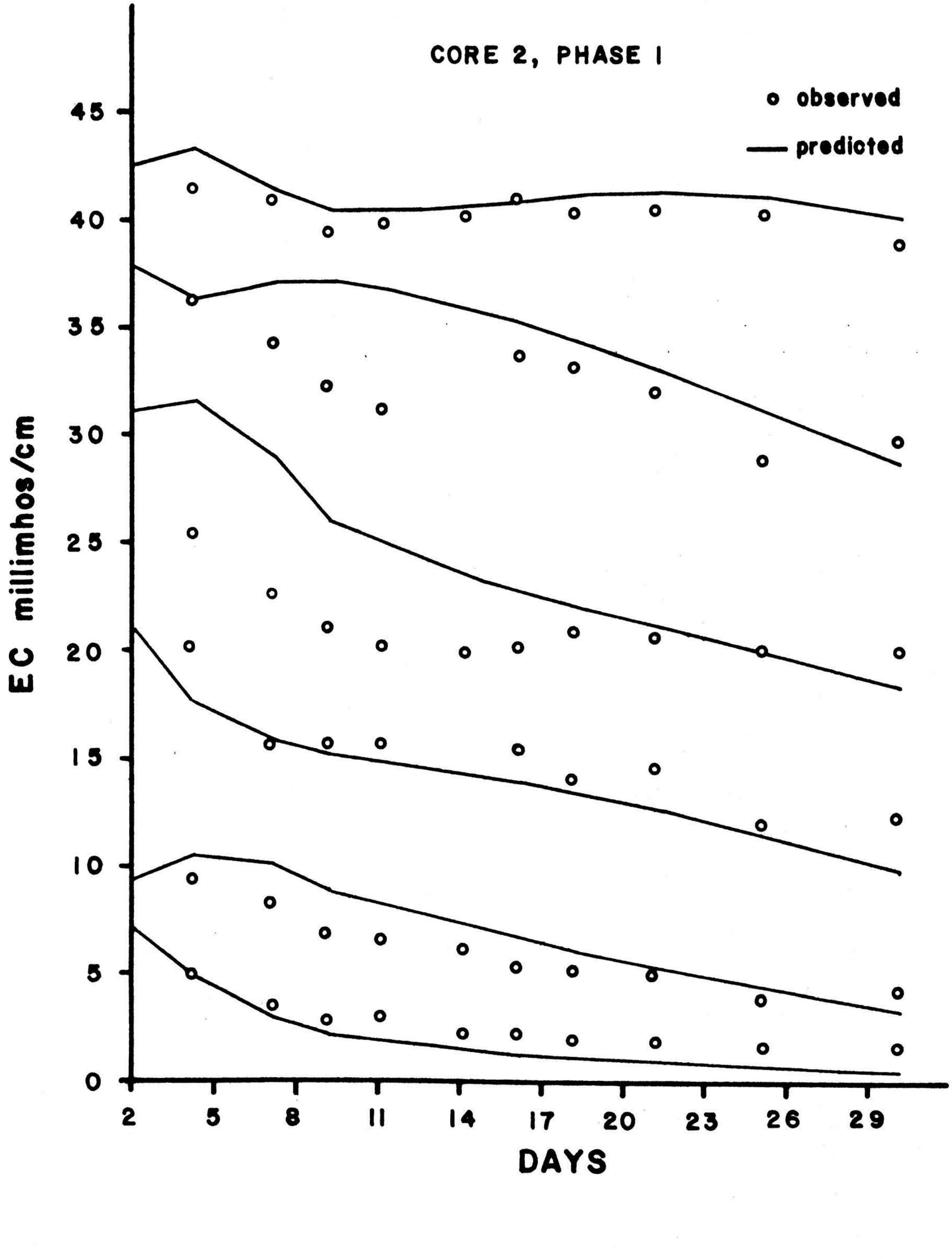

Figure 18. Predicted and observed changes in EC, Core 2, Phase I.

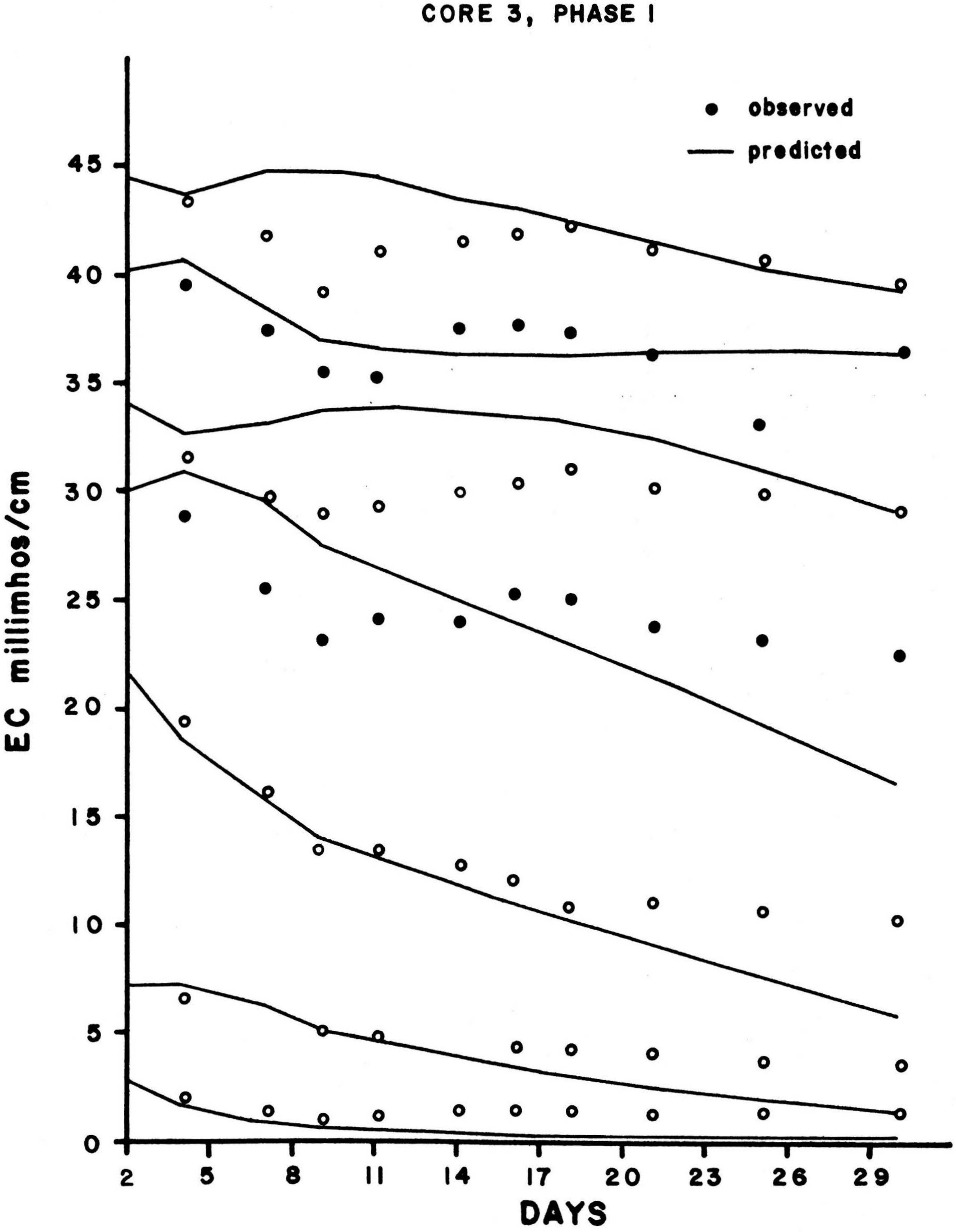

Figure 19. Predicted and observed changes in EC, Core 3, Phase I.

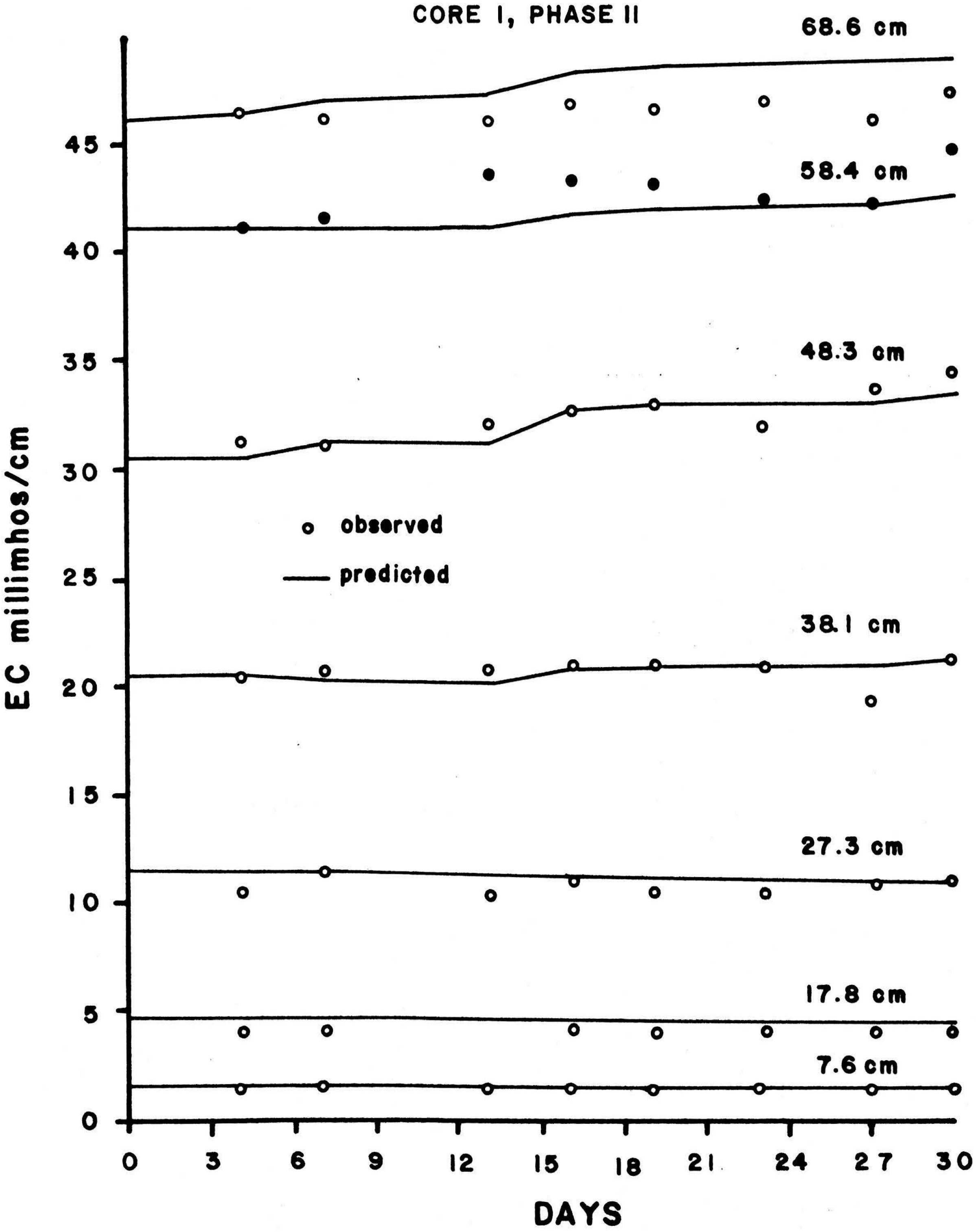

Figure 20. Predicted and observed changes in EC, Core 1, Phase II.

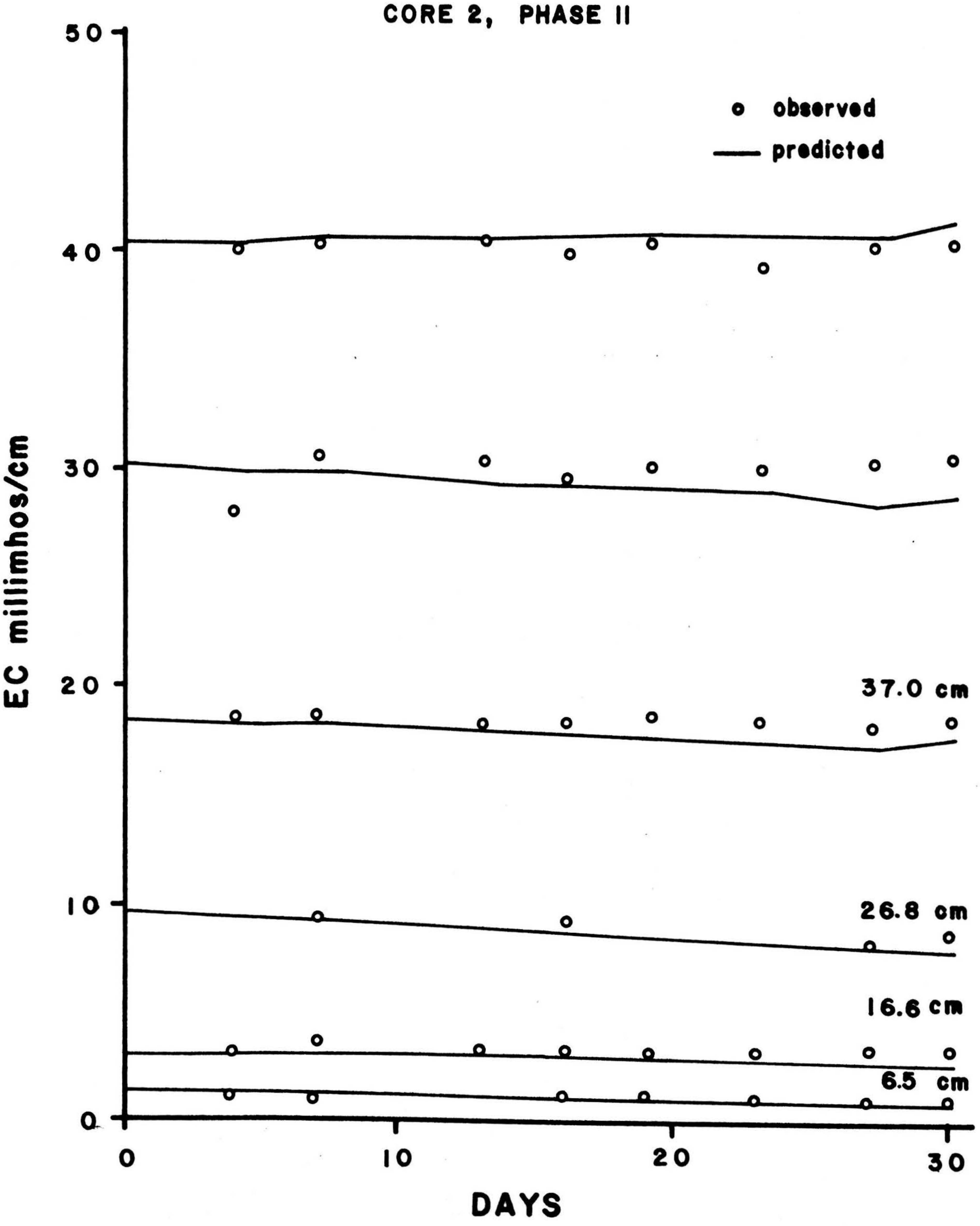

Figure 21. Predicted and observed changes in EC, Core 2, Phase II.

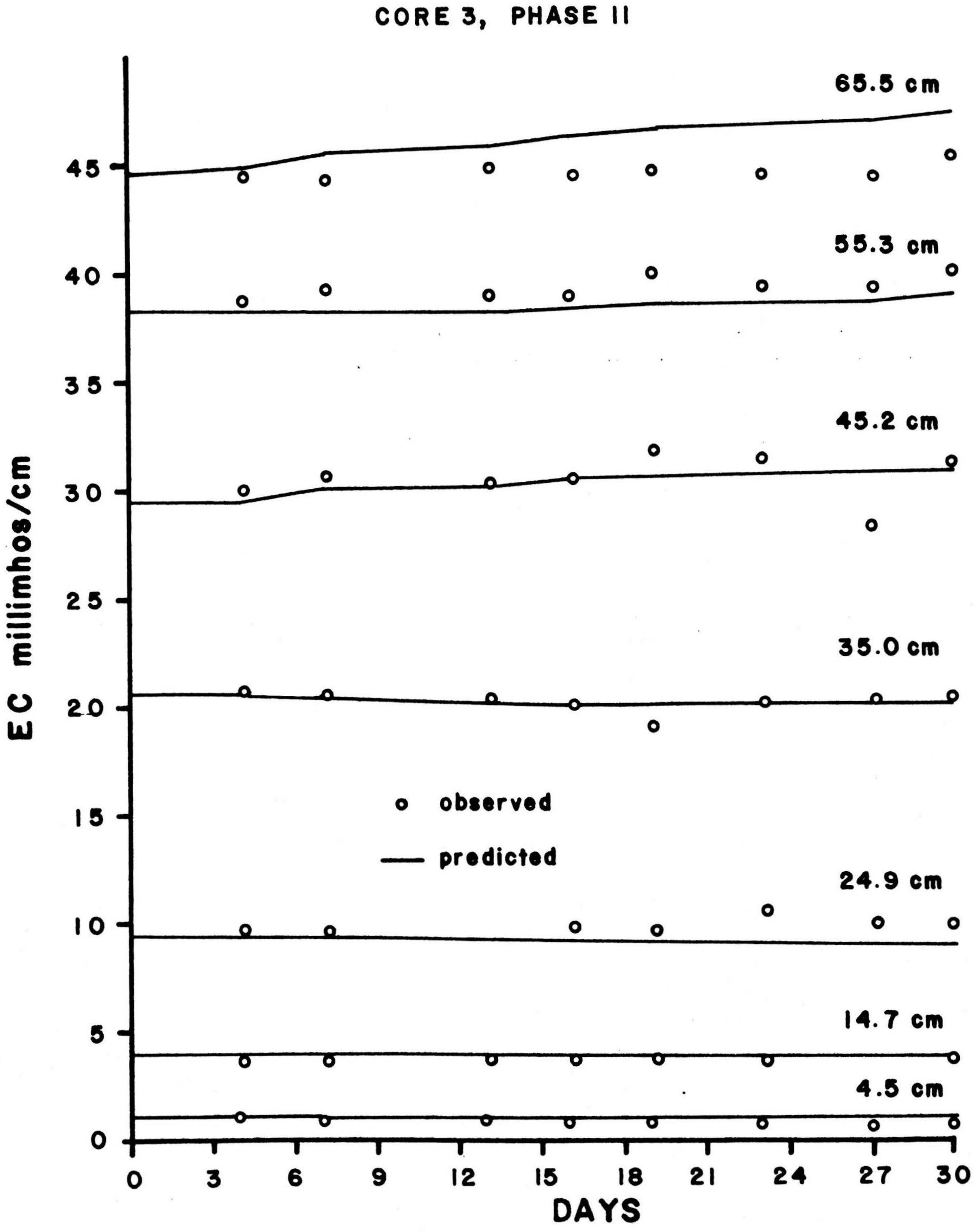

Figure 22. Predicted and observed changes in EC, Core 3, Phase II.

Cation Data

Some of the water samples taken from the soil cores during Phase I and II were stored and analyzed to determine Na^+, Ca^{++}, and Mg^{++} concentrations in the interstitial water. The purpose of this was to monitor the sodium adsorption ratio (SAR) of the soil water; in order to determine if changes in the soil hydraulic conductivity could potentially result from changes in solution composition. Unfortunately, the high concentrations of Na^+ existing in the water required sample dilutions of as great as 1000:1. This resulted in Ca^{++} and Mg^{++} concentrations that were too low to measure on the atomic absorption spectrophotomer being used. This, coupled with the lack of sensitivity inherent in large dilutions, makes the results useful only in showing general trends in solution composition over the course of the experiment.

Initial Na^+ concentrations were quite high in all three cores, ranging from roughly 100 meq/ℓ at the 0-8 cm depth to as high as 600-700 meq/ℓ at the 55-70 cm depths. On the other hand, Ca^{++} and Mg^{++} concentrations were initially quite low, ranging from less than 0.5 meq/ℓ to 30 meq/ℓ for Ca^{++}; and 2 meq/ℓ to 60 meq/ℓ for Mg^{++}. Based on these results, SAR values for the soil water sampled at the beginning of Phase I ranged from 100 to 200; indicating very large concentrations of Na^+ relative to the other cations. As the experiment progressed, Na^+ concentration were observed to decrease substantially; while little, if any, real changes were observed in Ca^{++} and Mg^{++} concentrations. As a result, SAR values at the end of Phase II ranged roughly from 10, at the top sampling depths, to 125 in the 30 to 50 cm depth range. SAR values for the surface waters remained quite low over the course of the experiment, usually varying from 2 to 5.

The major changes in solution composition occurred through the loss of Na^+ and increases in Mg^{++}. This was most pronounced at the lower depths of the columns. The brine water used had a high concentration of Mg^{++}. So, as water was pulled from the reservoir into the columns during Phase II, the Mg^{++} concentrations were observed to increase. As a result, SAR values in the lower portions of the column dropped to around 50 by the end of Phase II.

Results of Soil Characterization

The results of the soil characterization study are shown in Table 3. In general, the texture of the soil varied significantly with depth. This is consistent with the observed layering of the marsh sediments, with dense clay layers overlying more sandy layers, overlying clay layers, and so on. The bulk densities, on the other hand, are fairly constant throughout the column. However, the low bulk density found in the top layer of soil reflects the highly flocculant nature of this layer. The soil in the top 5 to 10 cm, in all marsh soils observed, was quite loosely packed and had a high organic matter content. Hence, the weight per unit volume of this layer was smaller than the denser more tightly compacted layers found deeper in the column. Consistent with these results are the saturated hydraulic conductivities. The highest K_s was found in the top layer, as expected, with the lowest values found in the layers containing the most clay.

In addition to the saturated hydraulic conductivities shown in Table 3, a hydraulic conductivity for the entire soil column was determined for each core. The values ranged from 0.07 cm/day for

Table 3. Characteristics of individual soil layers taken from Core 1.

Soil layer in cm below surface	Sand percent .05 dia. 2mm	Silt percent .002mm dia. .05mm	Clay percent dia. .002mm	Bulk density gm/cm^3	Saturated hydraulic conductivity cm/hr
0-7.6	20.5	48.5	31.0	1.03	.50
7.7-15.2	17.2	42.5	40.0	1.38	3.1×10^{-3}
15.3-22.9	15.9	43.8	40.3	1.42	2.7×10^{-3}
23.0-30.5	9.4	40.6	50.0	1.33	*
30.6-38.1	21.4	38.6	40.0	1.40	8.3×10^{-3}
38.2-45.7	34.4	44.6	21.0	1.44	.18
45.8-53.3	14.3	52.0	33.7	1.29	.11
53.4-61.0	21.6	49.0	29.4	1.35	.28
61.1-68.6	38.4	36.6	25.0	1.44	.27

* Significant disturbance of soil in sample prevented an accurate determination of hydraulic conductivity.

Core 1 to 0.06 cm/day for Core 2, and 0.05 cm/day for Core 3. Though these are rough estimates, they demonstrate the extremely low permeability of this soil.

Using data from the pressure plate apparatus, results were derived which relate water content to soil water matric potential and hydraulic conductivity (Table 4). Volumetric water content (in units of cm^3 of water per cm^3 of soil) varied curvilinearly with the log of the matric potential (Figure 23). In the 0-7.6 cm sample, water content was initially high, corresponding to the low bulk density, and decreased to 0.23 at 15,000 cm of pressure. For the 38.2 to 45.7 cm depth, water content was much lower at saturation (matric potential = 0), and decreased to a much lower level (.08) at 15,000 cm of pressure. Hydraulic conductivity decreases at approximately the same rate for both layers, though at 15,000 cm of matric potential the 38.2-45.7 cm sample will conduct considerably less water than the 0-7.6 cm layer (Figure 24). The curves presented in Figures 23 and 24 represent extreme cases. If the data in Table 4 were plotted for all the soil layers studied, the curves would all fall between those shown for the 0-7.6 cm and 38.2-45.7 cm depth.

Field Results

Figure 25 shows the variation in EC over a six month period of inflow and outflow surface water of Unit 1. Also, Figure 26 shows the variation in outflow from the marsh over the same period. EC and water flow rate was measured at the inlet structure to the marsh (Station 1). Thus, the average outflow EC was calculated as a weighted average with flow. As expected, the highest EC of both

Table 4. Water content with corresponding matric potential and hydraulic conductivity for soil layers studied.

Depth of Soil											
0-7.6 cm			7.7-15.2 cm			15.3-22.9 cm			30.6-38.1 cm		
θ	ψ	κ	θ	ψ	κ	θ	ψ	κ	θ	ψ	κ
.61	0	5.0×10^{-1}	.48	0	3.1×10^{-3}	.47	0	2.7×10^{-3}	.47	0	8.3×10^{-3}
.58	-10	5.9×10^{-2}	.46	-10	1.5×10^{-4}	.46	-10	3.2×10^{-5}	.46	-75	6.4×10^{-4}
.54	-25	9.0×10^{-3}	.44	-25	9.1×10^{-6}	.45	-100	5.5×10^{-6}	.45	-200	6.8×10^{-5}
.40	-1520	3.3×10^{-6}	.41	-300	2.9×10^{-7}	.40	-1620	3.9×10^{-7}	.40	-7460	1.4×10^{-6}
.35	-4470	1.0×10^{-6}	.35	-3140	2.4×10^{-8}	.35	-4575	1.2×10^{-7}	.37	-15000	3.6×10^{-7}
.30	-8600	4.3×10^{-7}	.30	-7450	7.7×10^{-9}	.30	-8570	4.8×10^{-8}	.30	-65720	1.5×10^{-8}
.23	-15000	1.3×10^{-7}	.23	-15000	1.7×10^{-9}	.23	-15000	1.3×10^{-8}	.23	-286100	6.2×10^{-10}

Depth of Soil											
38.2-45.7 cm			45.8-53.3 cm			53.4-61.0 cm			61.1-68.6 cm		
θ	ψ	κ	θ	ψ	κ	θ	ψ	κ	θ	ψ	κ
.46	0	1.8×10^{-1}	.51	0	1.1×10^{-1}	.49	0	2.8×10^{-1}	.46	0	2.7×10^{-1}
.44	-10	4.6×10^{-3}	.50	-10	1.0×10^{-2}	.48	-10	3.1×10^{-2}	.45	-10	2.6×10^{-2}
.42	-100	8.8×10^{-4}	.48	-75	4.6×10^{-4}	.45	-100	1.6×10^{-3}	.43	-100	2.6×10^{-3}
.38	-200	4.8×10^{-5}	.46	-300	9.9×10^{-5}	.43	-200	3.1×10^{-4}	.40	-200	2.0×10^{-4}
.36	-300	5.5×10^{-6}	.36	-3725	7.6×10^{-6}	.36	-2965	1.0×10^{-5}	.36	-1600	1.7×10^{-5}
.30	-1590	2.2×10^{-8}	.30	-7000	2.9×10^{-6}	.30	-8310	2.7×10^{-6}	.30	-6670	3.2×10^{-6}
.08	-15000		.18	-15000	3.4×10^{-7}	.24	-15000	7.4×10^{-7}	.23	-15000	6.3×10^{-7}

KEY: θ = Volumetric water content - cm^3 H_2O/cm^3 soil

ψ = Soil water matric potential - cm H_2O

κ = Hydraulic conductivity - cm/hour

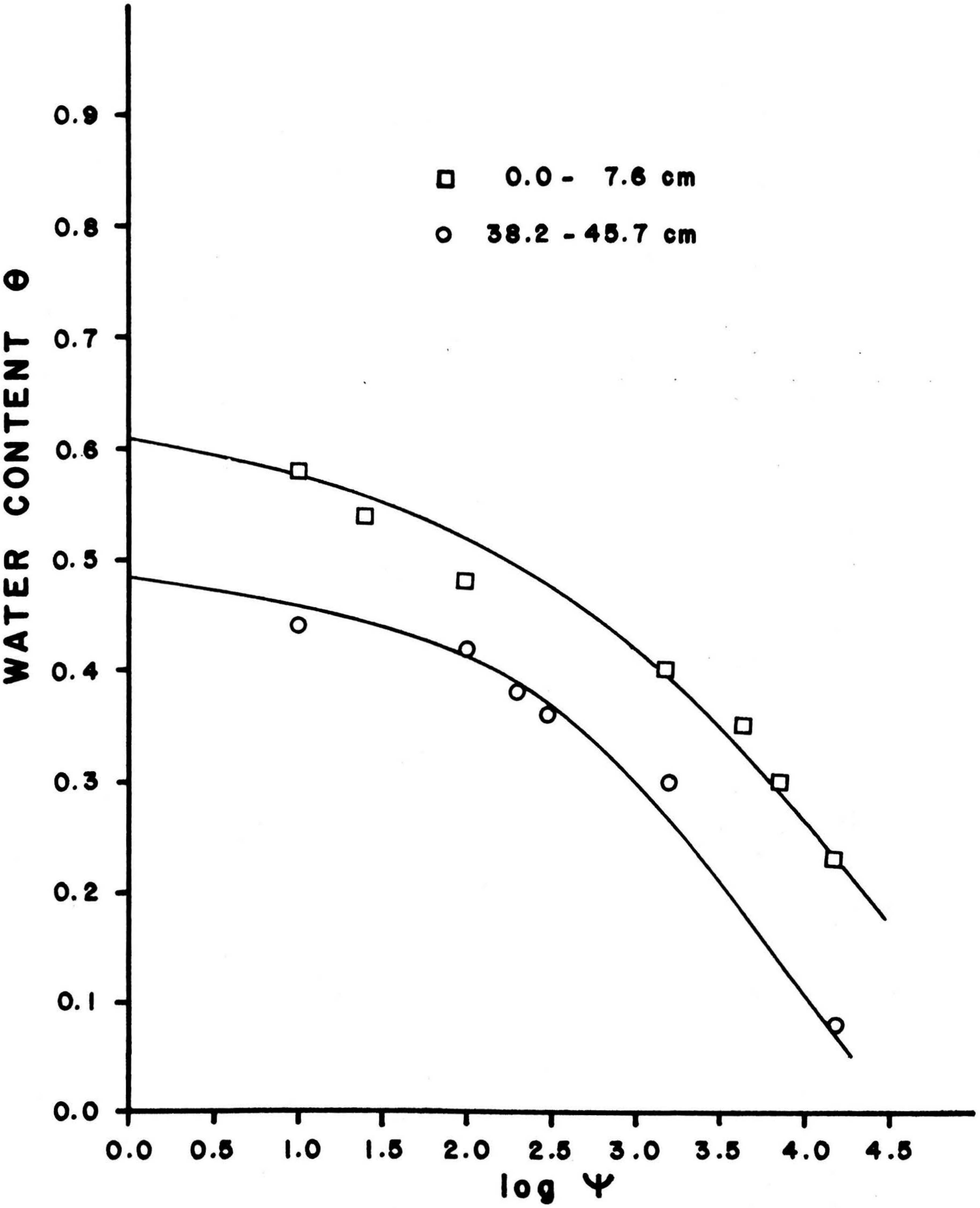

Figure 23. Volumetric water content vs. matric potential.

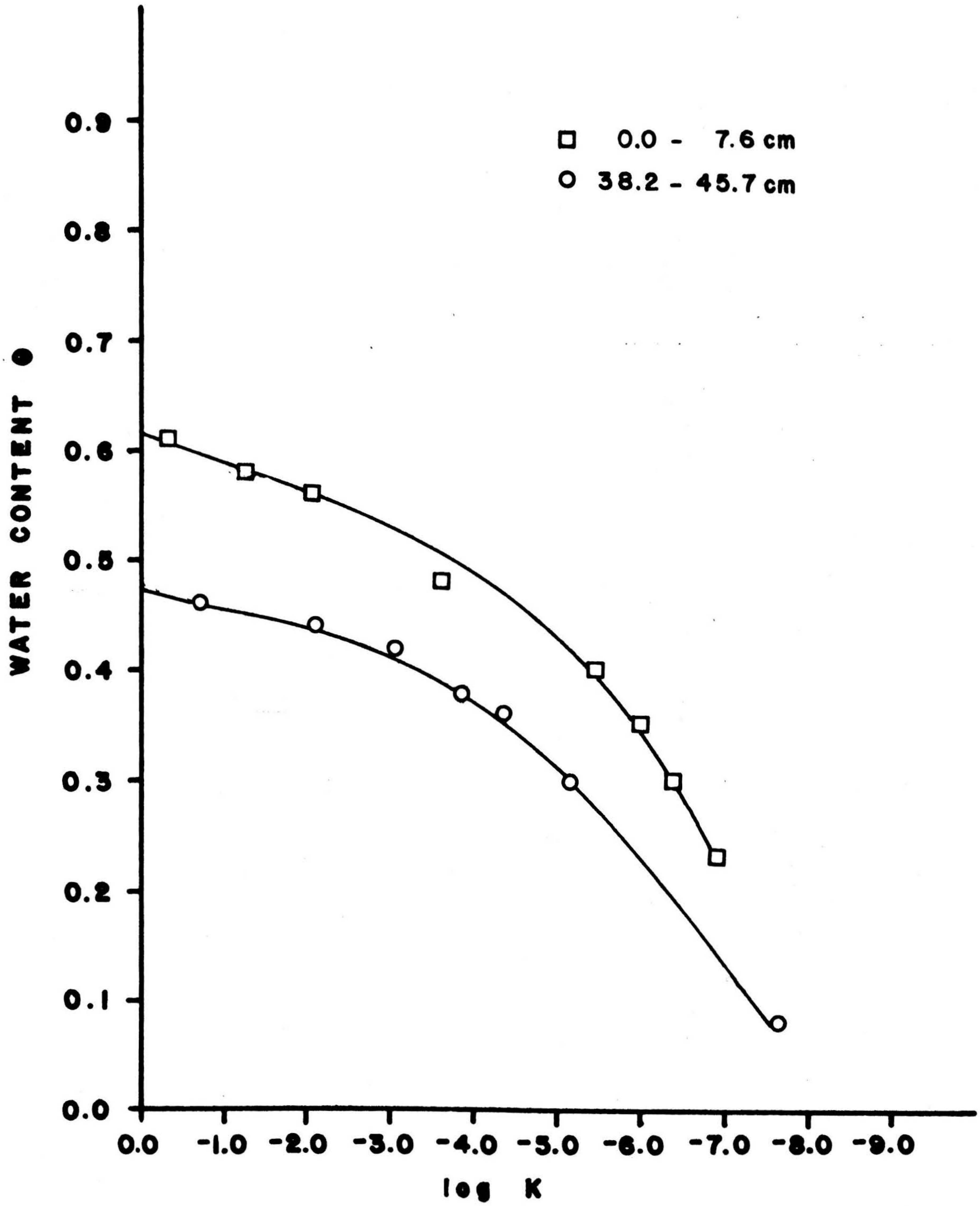

Figure 24. Volumetric water content vs. hydraulic conductivity.

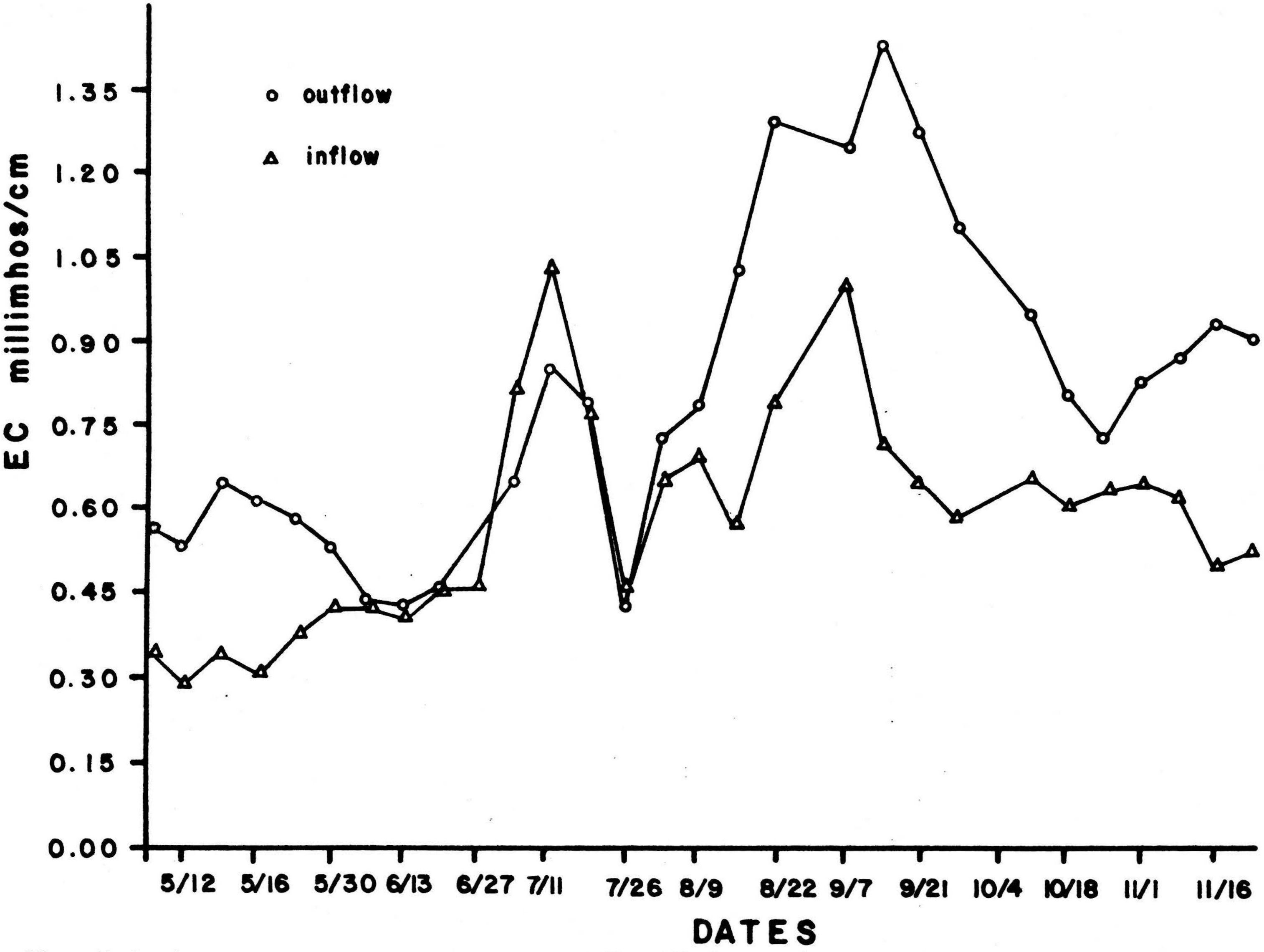

Figure 25. Variation in inflow EC and average outflow EC for spring, summer, and fall 1978.

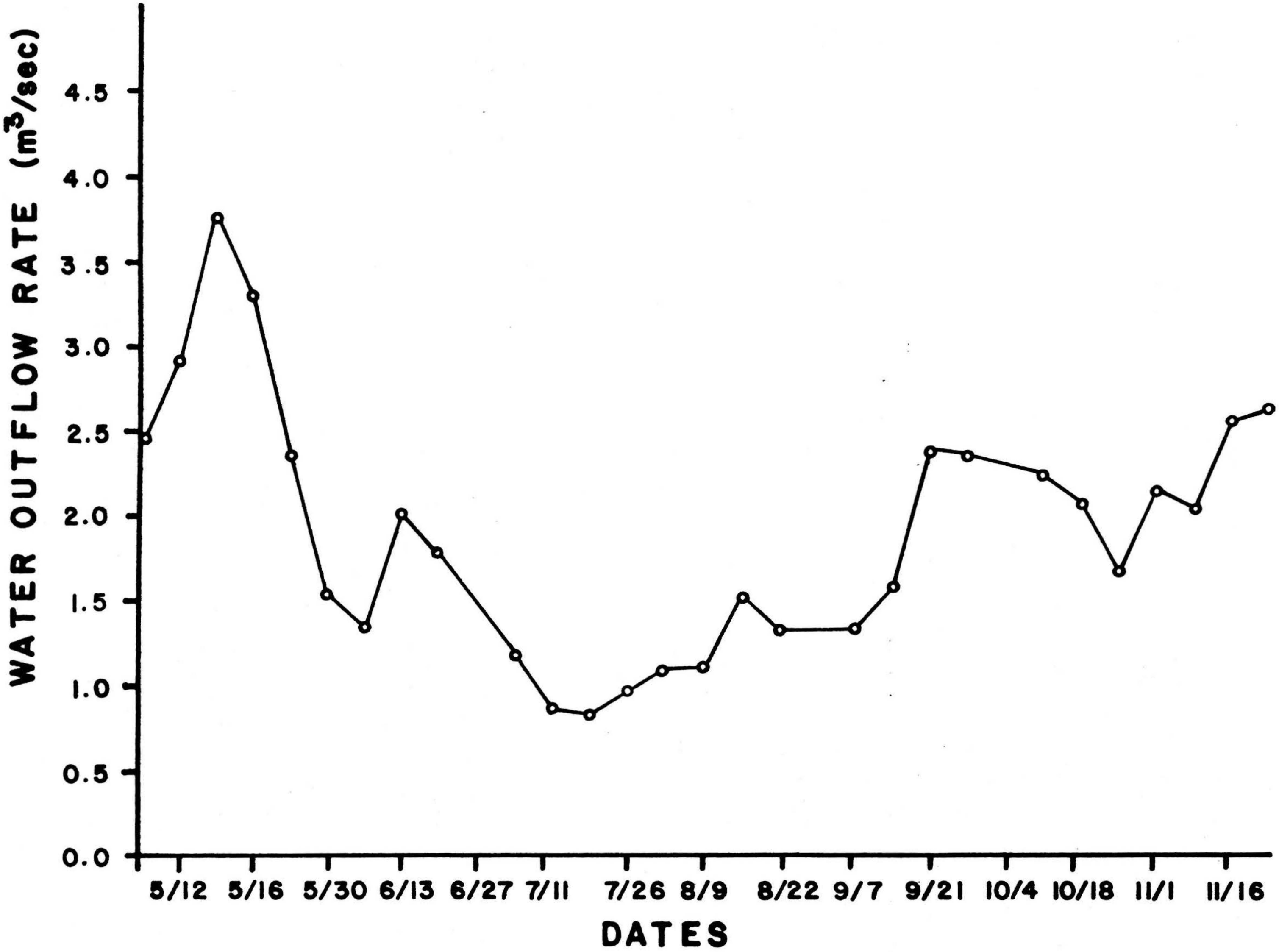

Figure 26. Variation in outflow rate for spring, summer, and fall 1978.

inflow and outflow water occurred in the summer months, while the largest flow occurred during the spring. Outflow decreased in the summer in response to evapotranspiration losses and a demand for Weber River water for irrigation, upstream from the marsh.

Variation in interstitial water EC is demonstrated in Figures 27 and 28 for two soil water sampling stations located in the study unit. These two stations were chosen because they typify two types of soil-surface-water systems found in the marsh. Well Station 2 is in an area that is covered with shallow water (1-15 cm) for most of the year; but it is dry during periods of low water level. On the other hand, Station 7 is in an area that is flooded at all times. Here water depths vary from a minimum of 15 cm during periods of low water level, to a maximum of 60 cm during periods of high water level. It should be noted that the period between July 5 and August 15, when the most pronounced change in EC occurred at each station, was a period of very low water level within the study unit. During this period the soil surface at Station 2 went dry for a period of approximately four weeks and then was reflooded.

Figure 29 shows the EC gradient with depth for Stations 7 and 5 at the two times indicated above. Station 5 was similar to Station 2 in that the surface dried and was reflooded over the same period. Also, the largest soil water EC was observed at Station 5.

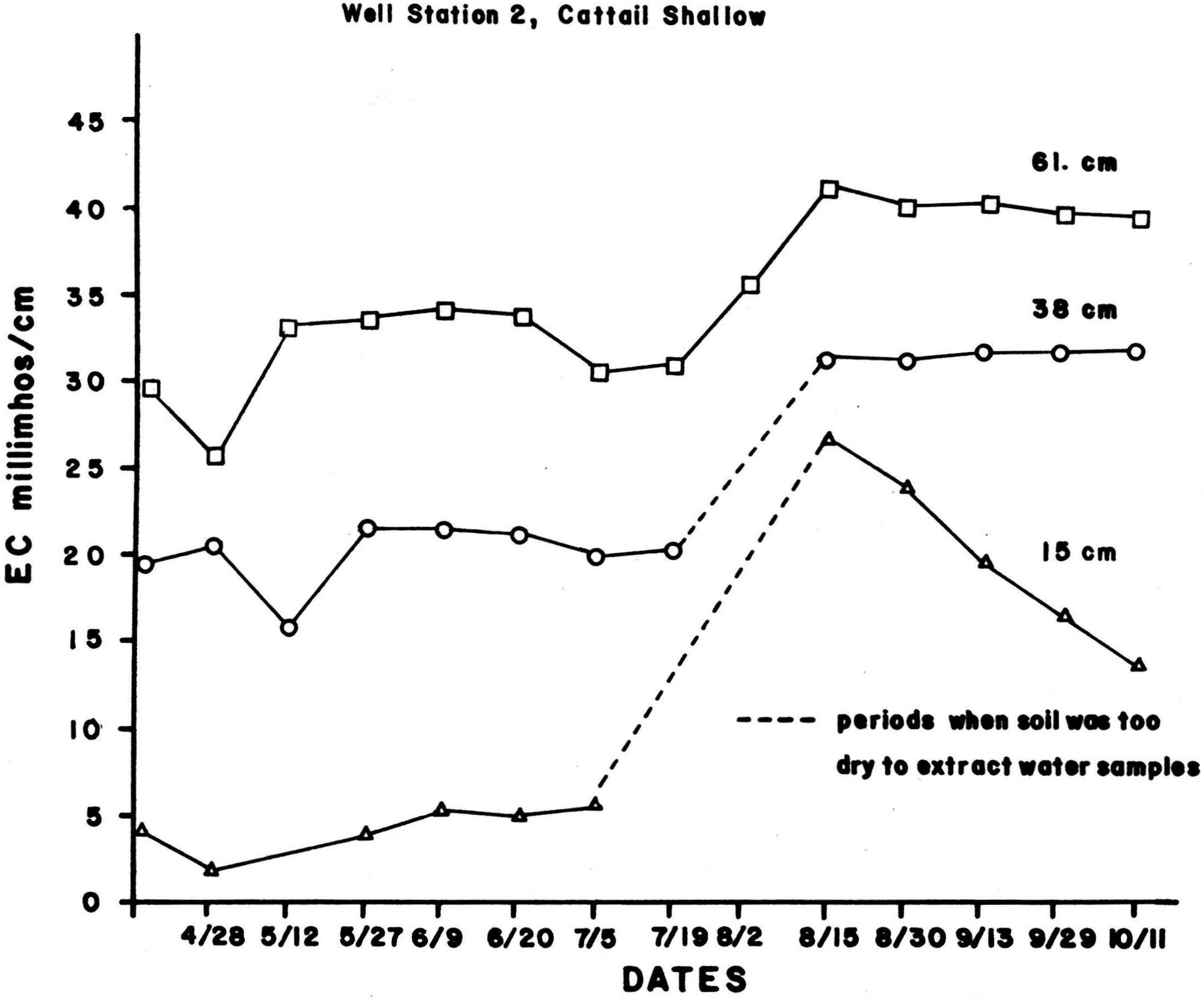

Figure 27. Soil water EC vs. time at each sampling depth for well station 2.

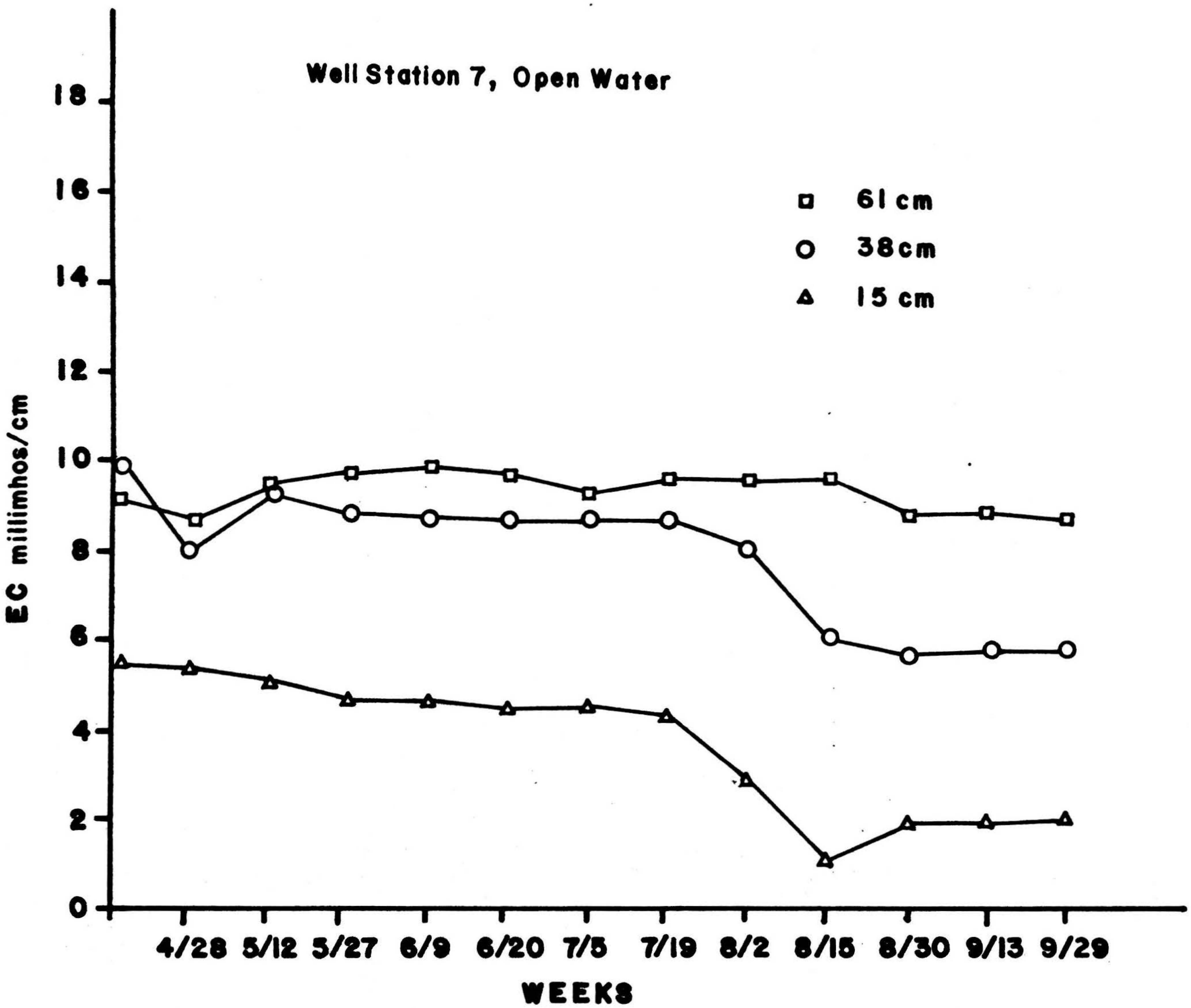

Figure 28. Soil water EC vs. time at each sampling depth for well station 7.

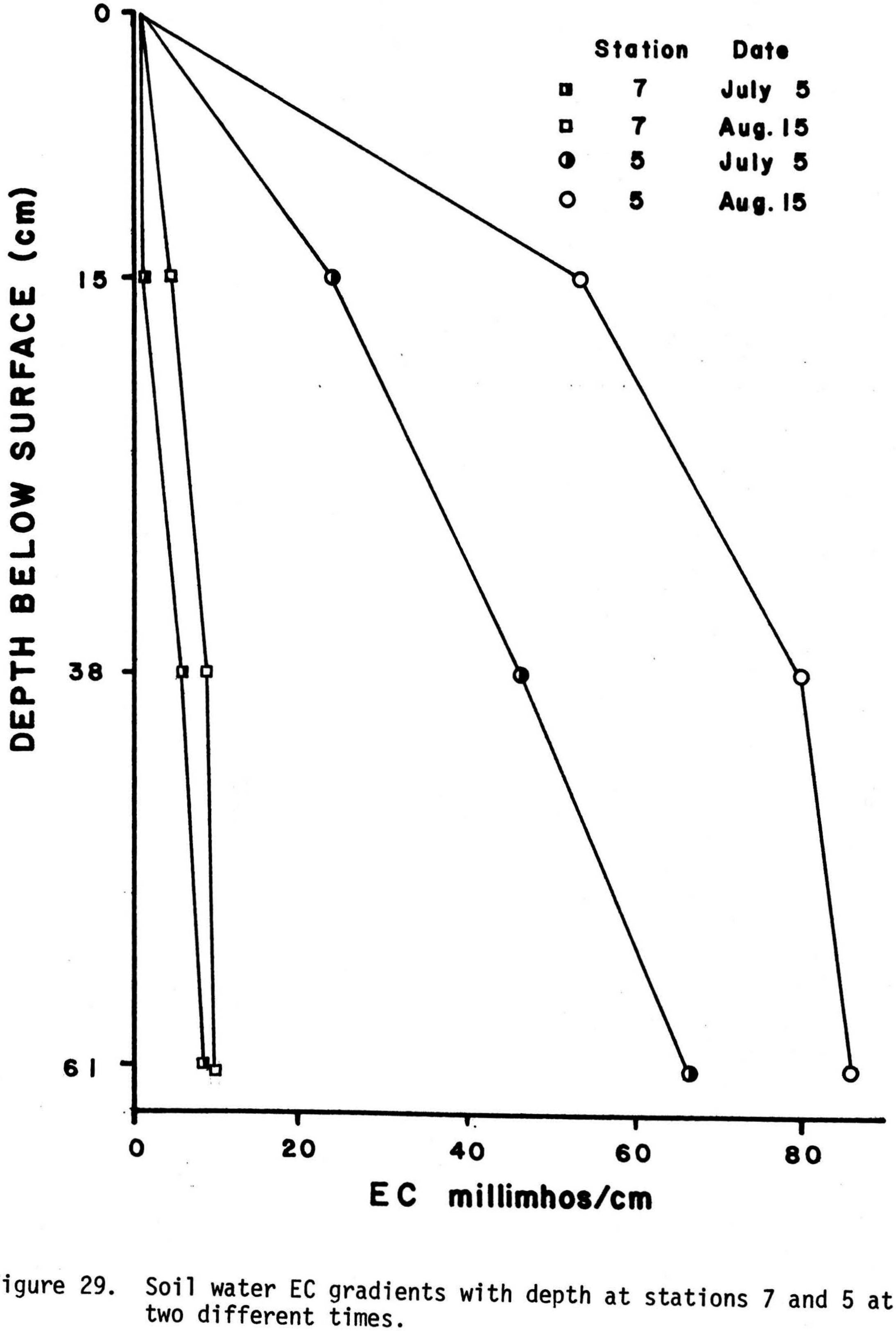

Figure 29. Soil water EC gradients with depth at stations 7 and 5 at two different times.

exist for the movement of salt through diffusion. Further evidence which contradicts the significance of this process is given by *Saxena* [1974]. In this paper, the author shows that diffusion coefficients are dependent on pore sizes in a soil. The author goes on to state that "...for small pores, as in closely packed particle matrices of fine silt size and less, the rate of diffusion is significantly reduced". The high clay and silt content, as well as bulk density of the soil (as given in Table 3), indicates that diffusion rates would be quite small for this soil. Also, *Peck* [1971] states that at a mean pore water velocity of greater than 10×10^{-6} cm/second, diffusion is less significant than water movement in the transport of salt. In Phase I, the mean pore velocities for each core were roughly 4×10^{-6} cm/second, indicating that diffusion was insignificant in this case. In Phase II, the mean pore velocities were calculated to be about 4×10^{-7} cm/second. Though this is the same order of magnitude as the number presented by *Peck*, the good agreement between the observed and predicted results in Phase II indicates that even at this low mean pore velocity, diffusion is insignificant.

Though diffusion is probably not important in the bulk portion of the soil column, evidence does exist that diffusion can occur between upper soil layers (first 8 cm) and the surface water in the soil core experiment. Figures 15 and 16 show that the EC of the surface water did in fact increase during the experiment. The increases in surface water EC during Phase II can be almost completely accounted for through evaporation. However, during Phase I evaporation losses were kept to a minimum. Therefore the increased EC of the surface water must have been a result of salt moving from the soil into the surface water.

To explain how diffusion could occur in this case, and not in the lower portions of the soil, one must examine the bulk densities in Table 3. The bulk density of the top layer of soil is considerably smaller than in the lower portions of the column. In addition, the top layer of soil was completely saturated, whereas the lower layers probably were not. As a result, the pore sizes in the top layer of soil were undoubtedly larger than in the lower layers. This means that even though the mean pore water velocity was constant throughout the column, the actual microscopic pore-water-velocity in the top layer of soil was much smaller than in the lower layers, due to the larger cross sectional area available for flow. Therefore, salts could have conceivably diffused from the upper layer of soil into the surface water, causing the observed increase in EC. An approximate diffusion coefficient for this system (based on the change in EC in the surface water during Phase I; assuming NaCl represented the majority of the salt) is 2.0×10^{-6} cm^2/second; which is the same order of magnitude as published values of diffusion coefficients for salt moving through a synthetic porous media [*Stoessel*, 1975].

Potential for Changes in Soil Hydraulic Conductivity

If the movement of water through the soil is the dominant mechanism controlling pore water salinity, then the ability of the soil to conduct water is important in determining how salinity will change under given conditions of hydraulic potential. As explained earlier, two processes that could act to alter the hydraulic conductivity of a soil are changes in the soil water chemical composition, and changes in the soil moisture content. The literature shows that waters of high SAR and low overall salt concentration can potentially cause reduced hydraulic conductivity in soils of high clay content [*Naghshineh-Pour, et al.*, 1970; *Yoron and Thomas*, 1968; *McNeal, et al.*, 1966; *McNeal, et al.*, 1968; and *Frenkel, et al.*, 1978].

Although the cation analyses presented earlier provides only general changes in the soil-water composition, sodium was the dominant cation in the soil-water system studied. This fact, coupled with the relatively high clay content of the marsh soil, indicated that significant reductions in hydraulic conductivity could occur. However, the observed decrease in SAR over the course of the experiment and the very high salt concentrations in lower portions of the column would indicate greater hydraulic conductivities. Based on the literature cited little if any decrease in hydraulic conductivity would occur in the lower soil layers due to the very high solute concentrations. On the other hand, in the upper portions of the column decreases in salt concentration could cause significant reduction in soil permeability even though SAR values were reduced. *Naghshineh-Pour, et al.*, [1970] shows that for the high clay soil studied, hydraulic conductivity was reduced by roughly fifty percent when leached with waters with SAR values ranging from 20 to 40 and electrolyte concentrations of from 10 to 40 meq/ℓ (EC = 1 to 4 mmhos). For comparison, this is the range of chemical composition found in water samples taken from the top two soil water samplers at the beginning of Phase II. Though much more study is required to predict how the Odgen Bay marsh soils will react to changes in solution composition, it appears that because of high sodium levels, hydraulic conductivity could vary significantly depending on how soil water SAR and salt concentrations fluctuate.

The other factor that could affect the flow of water through the soil is the moisture content of the soil. Figure 24 and Table 4 indicate how hydraulic conductivity of the soil can be significantly affected by changes in soil moisture. Conceivably, under field conditions, an area of soil that dries periodically may not rewet to saturation. In this case, the flow of water and salt through the soil could be drastically reduced; thus acting to maintain the salinity gradient in the soil. As the soil continues to rewet, the hydraulic conductivity will increase; thereby increasing the flux of water and solutes. As shown in Table 4, different soil layers will contain different amounts of water and conduct water at different rates at a given matric potential. Thus, the field situation becomes quite complex as different soil layers drain and rewet at varying rates.

Extrapolation of Experimental Results

The soil samples used in the soil core experiment represent conditions that exist in only one minute area of the marsh system. Therefore, direct application of the experimental results to the field situation would be very risky. However, the results can be used to indicate the potential changes that could occur over a period of time. For example, the results seem to indicate that under conditions of very small hydraulic potential, such as that tested during Phase II, little, if any, change in soil water EC could be expected in a 30 day period. On the other hand, under conditions where the movement of water is greater such as in Phase I, the change in soil water salinity can be significant.

To predict what might happen over a longer period of time, the model was used to determine the soil water EC at each depth in a soil core (Core 1) under constant conditions of hydraulic potential for a period of approximately one year. The values of surface water EC were chosen to approximate the yearly cycle in EC found in surface water in the field. Figure 27 shows the salinity gradient that existed initially in Core 1 (prior to Phase I), as well as the predicted final salinity gradient that would exist after one year under 30 cm and 61 cm of hydraulic head. Assuming a hydraulic conductivity for the soil of 0.06 cm/day, these hydraulic heads translate into flows of 8.6 mℓ/day and 17.8 mℓ/day respectively. Sample volumes were assumed to be zero in this simulation, to better reproduce the actual situation. Significant reduction in soil water EC occurred at all depths with the greatest reduction occurring in response to the most flow (Figure 30). Similarly, 30 cm of positive soil water pressure was used to simulate a condition of 8.6 mℓ/day of water moving upward through the column from the brine reservoir (Figure 31). Significant increases in EC were predicted from the model. Thus, it appears that over a long-term period, significant changes in soil water salinity can occur if water is allowed to flow through the soil.

Again, it must be emphasized that the actual field conditions can and probably will be quite different than those used in the model and during the experiment. For example, over a one year period, surface water level varied a great deal indicating that the hydraulic potential acting across the soil could change significantly from week to week. Also, for the one month that the water table level was monitored, the elevation of the water table was never observed to be more than a few centimeters below the surface water elevation. Considering the

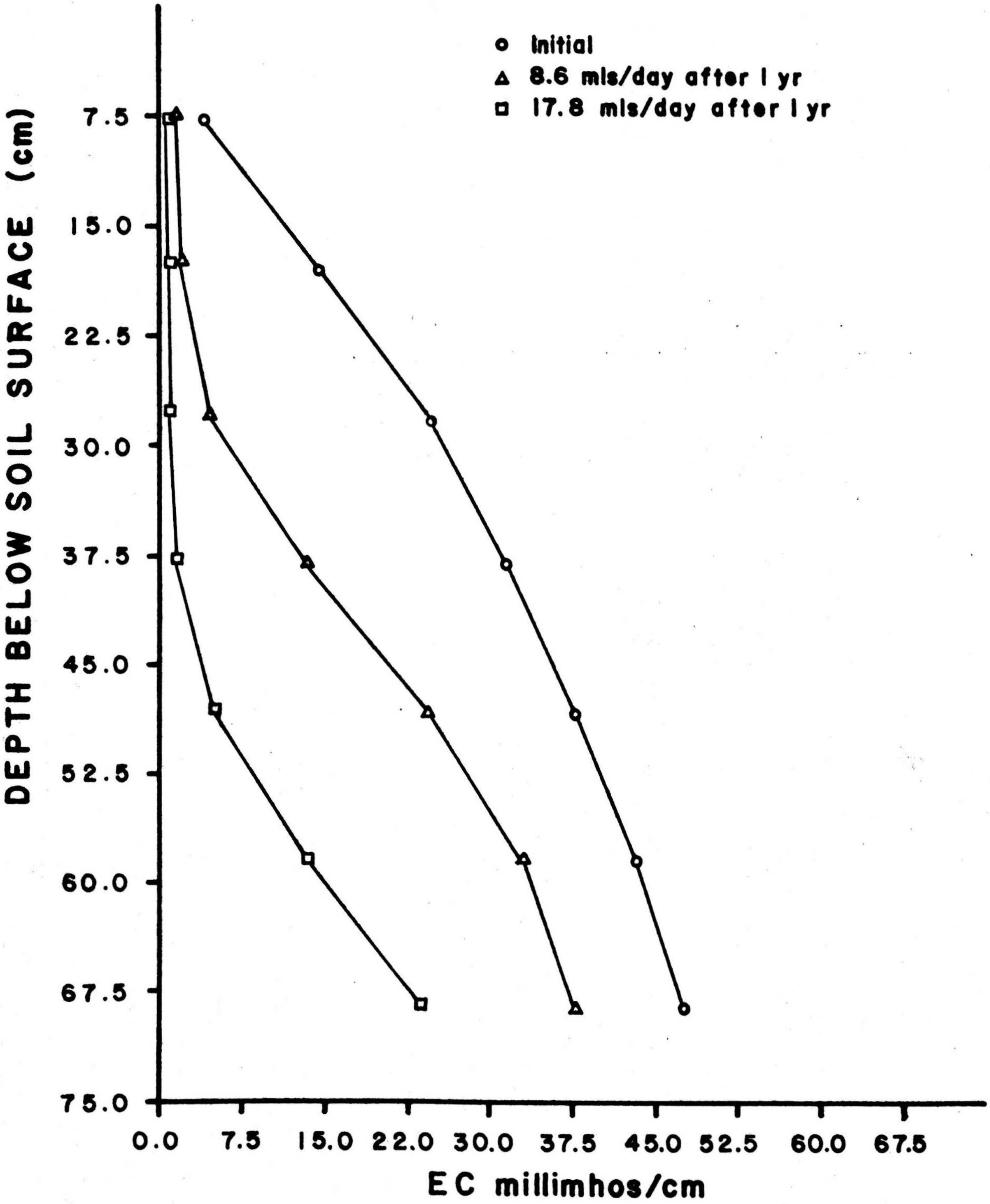

Figure 30. EC gradients with depth: Initial conditions, Core 1, Phase I and predicted based on two soil water flow rates.

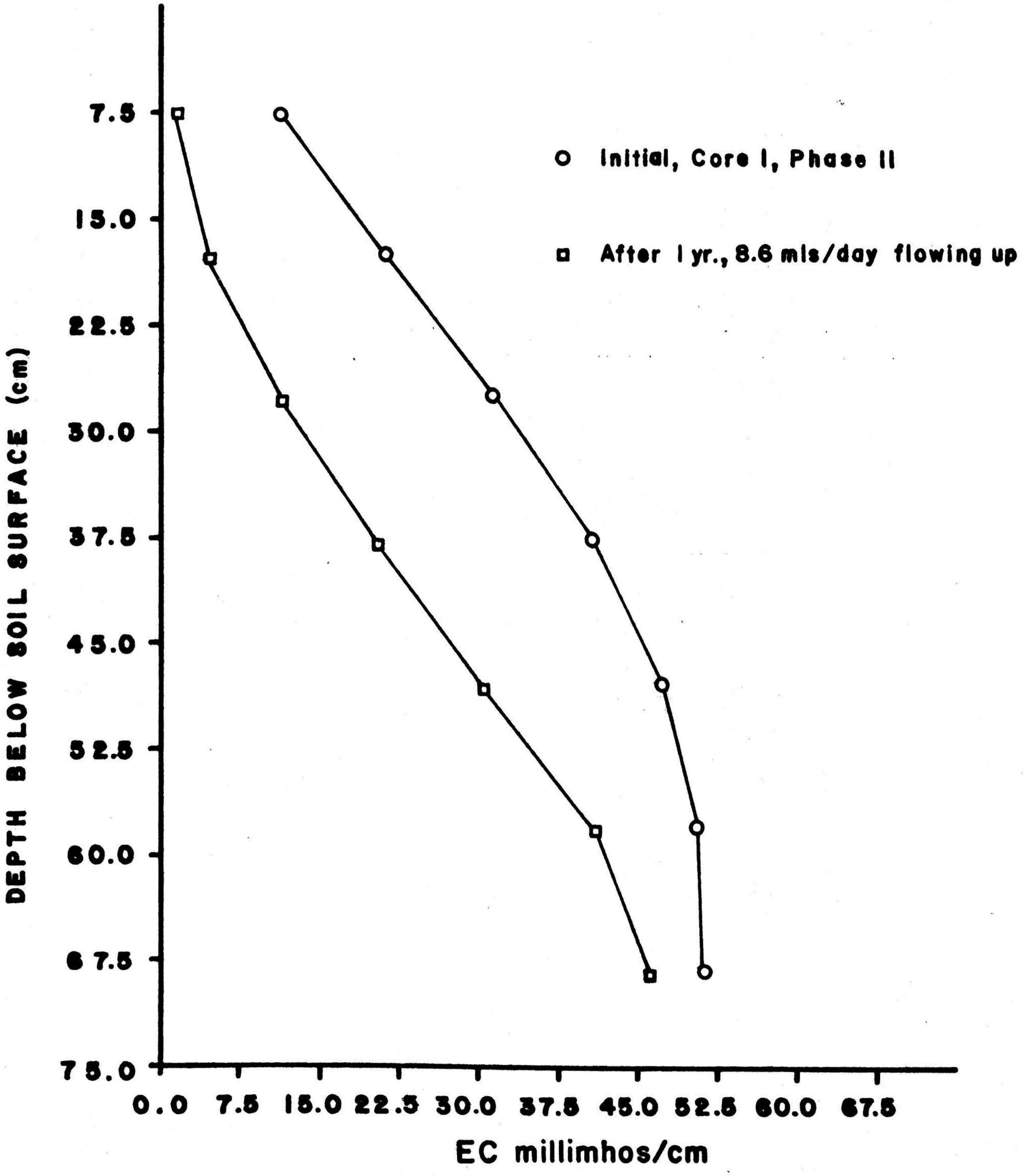

Figure 31. EC gradients with depth: Initial conditions, Core 1, Phase II and predicted based on an assumed flow rate.

extremely low hydraulic conductivity of this soil, differences between surface water elevation and water table elevation must be fairly large if appreciable water movement is to occur through these marsh sediments. Also, changes in hydraulic conductivity resulting from changes in soil moisture and solution composition could also affect the flow of water through the soil. Therefore, even though the model predicts substantial changes in interstitial water salinity over a long-term period, these and other factors act to alter the actual field situation.

Interpretation of Field Results

Both inflow and outflow EC vary significantly over a period of several months (Figure 25). Also, the flow of water through the marsh (as indicated by the outflow rate) was observed to fluctuate considerably with season. During periods of low flow (July 5 through August 1), inflow and outflow EC were not appreciably different. On the other hand, during periods of increasing flow rate, outflow EC is considerably higher than inflow EC. Also of importance is that evapotranspiration losses are highest during July [*Christiansen and Low*, 1970], the period that inflow and outflow EC are very similar. In May and September, when outflow conductivity was considerably larger than inflow EC, evapotranspiration rates are relatively small. Therefore, evapotranspiration does not appear to be the only mechanism controlling surface water electrical conductivity in the marsh.

Examination of the interstitial water EC data provides a possible explanation of the results presented in Figure 25. In Figure 27, soil water electrical conductivity was observed to increase substantially at Station 2, between July 5 and August 15. Also, Figure 29 shows how drastically soil water EC rose at Station 5 during the same period. As explained before, this period was one of low water level. During this period much of the marsh soil, including the areas around Stations 2 and 5, were left dry and exposed. During the same period, Station 7 remained completely inundated and actually showed a small decrease in soil water EC. Eventually, the water level rose in response to the increased flow of water through the marsh. As a result, many of the dry, exposed areas, including Stations 2 and 5, were reflooded. Apparently, the changes in EC that occurred during this period were related to the drying and reflooding of the marsh soil.

The process involved in this phenomena involves the movement of salts upward in the dry soil through the capillary rise of salty water from the lower soil layers. Conceivably, when soil dries out, highly saline water within the soil rises toward the surface due to capillary forces. The water evaporates, leaving the salt in the soil and on the surface. When the area is again reflooded, the deposited salt goes into solution; thus causing the observed jump in EC. At Stations 2 and 5, the soil was observed to dry as deep as 61 cm below the surface (as evidenced by an inability to draw water samples from the porous cup sampler). Thus, when water reflooded the area, soil water EC values rose dramatically. Surface water conductivity was also observed to be quite high (see Appendix IV). Therefore, the high outflow EC observed during periods of high surface water level, could be a result of salts deposited on the soil surface being resolubilized upon the return of high surface water. These salts eventually leave the system as water moves across the marsh.

At Station 7 EC changed very little. The small decrease that did occur could have been a result of the lowering of the water table during low flow; which allowed water and salts to flow through the soil in a manner similar to that which occurred in the soil core experiment. The large difference between the salinity gradients that exist at Stations 7 and 5 (as shown in Figure 29) may be a result of the water level cycles that seem to occur in the marsh system. In an area constantly under water, salts could be leached downward regularly, causing soil water salinity to be moderated. On the other hand, in areas that undergo a wetting and drying cycle, a large salinity gradient could be maintained by the regular upward movement of salts in the soil due to capillary action. Thus, two areas, only a few hundred yards apart, exhibit strikingly different soil salinity profiles.

CONCLUSIONS

Based on the results presented, the following conclusions can be made regarding salinity fluctuations in the marsh soils studied from Ogden Bay Waterfowl Management Area.

1. Under conditions of continuously wetted soil, the dominant mechanism affecting changes in interstitial water salinity is the movement of water and soluble salts through the soil profile.
2. The rate of water movement, as well as the salinity gradient that exists in the soil profile, will determine how much the soil water salinity will change in a given period of time.
3. In general, diffusion is a minor process affecting soil water salinity.
4. Due to the flocculent nature of the top layer of sediment in most marsh soils, diffusion of solutes can occur between the upper few centimeters of soil and the surface water.
5. The presence of water with high SAR values indicates that a potential exists for significant reduction in soil hydraulic conductivity when interstitial water with high solute concentrations is replaced with water of low solute concentration.
6. Though saturated soil conditions can exist at the soil surface and at lower layers of a soil profile, unsaturated intermediate layers can significantly reduce the hydraulic conductivity of a soil and, hence, reduce the amount of water flowing through the profile.
7. Based on the field data observed, the drying out of the marsh

soil during periods of low surface water level followed by the reflooding of the area, resulted in large increases in both interstitial water salinity and surface water salinity.

In general, it appears that the relationship between surface water salinity and soil water salinity revolves around the movement of water between the two. At present, the drying and reflooding of large areas of marsh soils appears to be the dominant mechanism acting to increase surface water salinity in the short term. However, more study of this phenomena is needed before generalizations can be drawn. Also, if the movement of water and salts between the soil and surface water is to be quantified for the entire marsh, many more experiments, similar to the soil core experiment described here, are needed using soil from several other areas of the marsh. This type of experiment can provide useful information if all variables are carefully controlled, some of the most important being sample sizes and sample frequency. Also, interpretation of the results must take into account all operational procedures, as well as the physical processes actually occurring in the field.

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APPENDICES

APPENDIX I. Observed results of soil core experiment.

Table 5. Observed results, Core 1, Phase I (all EC data expressed in millimhos/cm).

Sample depth below soil surface(cm)	Initial		Day 2		Day 4		Day 7[1]	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.286	---	≠	---	≠	---	≠	---
7.6	6.29	20.0	4.03	45.0	3.55	45.0	2.14	60.0
17.8	15.30	15.0	14.40	30.0	12.10	30.0	10.20	30.0
27.9	25.40	8.0	24.70	15.0	22.10	15.0	19.60	20.0
38.1	30.90	15.0	31.50	25.0	29.30	25.0	25.10	35.0
48.3	35.40	10.0	37.80	15.0	35.90	15.0	35.20	20.0
58.4	41.20	20.0	43.25	35.0	42.30	35.0	40.30	45.0
68.6	43.70	25.0	47.70	45.0	45.40	45.0	42.30	65.0
observed drop in surface water water volume between sampling days	---		No change		200 mls		420 mls	

---

* Sample volumes are stimates due to unmeasured volume water that drained into sample bottle between sampling days

≠ Surface water EC not reported due to a malfunction in the meter which yielded inconsistent values at low levels of EC

[1] Procedure changed on Day 7 - Sample bottles were removed between sampling days.

Table 5. Continued.

Sample depth below soil surface(cm)	Day 9		Day 11		Day 14		Day 16	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.314	---	.313	---	.339	---	.354	---
7.6	2.07	7.16	1.86	7.0	1.93	13.0	1.78	10.20
17.8	8.99	8.6	8.53	6.4	8.10	9.6	6.92	5.90
27.9	18.80	9.0	18.00	6.6	18.30	8.4	17.80	6.30
38.1	24.40	23.0	25.10	10.4	24.60	16.0	25.15	17.40
48.3	33.90	14.0	35.20	9.4	34.55	13.0	33.30	13.20
58.4	38.30	16.0	37.80	9.4	40.40	15.2	42.00	14.80
68.6	41.40	21.9	41.40	11.8	44.40	28.0	45.70	17.80
Drop in surface water volume between sampling days	400 mls		No change		360 mls			

Table 5. Continued.

Sample depth below soil surface(cm)	Day 18		Day 21		Day 25		Day 30	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.360	---	.373	---	.378	---	.389	---
7.6	1.69	9.6	1.68	11.2	1.67	11.2	1.64	15.0
17.8	6.90	7.7	6.19	7.7	5.24	5.8	5.05	8.6
27.9	17.20	8.8	16.10	8.0	15.40	7.1	13.30	10.4
38.1	24.90	16.5	23.90	25.0	22.60	15.8	21.20	24.0
48.3	34.70	8.5	34.95	13.4	32.20	7.8	32.00	12.2
58.4	40.70	12.2	39.00	11.0	36.10	11.3	38.83	44.3
68.6	45.4	13.6	43.40	16.5	41.60	11.3	43.00	44.3
Drop in surface water volume between sampling days	150 mls		240 mls		160 mls		370 mls	

Table 6. Observed results, Core 2, Phase I (all EC data expressed in millimhos/cm).

Sample depth below soil surface(cm)	Initial		Day 2		Day 4		Day 7[1]	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.271	---	≠	---	≠	---	≠	---
6.5	12.90	25.0*	7.05	45.0*	4.76	45.0*	2.76	7.4
16.6	12.90	15.0	9.37	25.0	9.50	25.0	6.86	9.0
26.8	23.90	6.0	20.40	8.0	20.20	8.0	15.70	6.1
37.0	29.20	20.0	31.20	40.0	25.30	40.0	20.90	19.2
47.1	35.40	25.0	37.80	35.0	36.10	35.0	32.30	8.6
57.3	40.80	25.0	42.60	40.0	41.30	40.0	39.30	11.5
Drop in surface water volume between sampling days	---		No change		200 mls		400 mls	

---

* Sampling volumes are estimates due to the unmeasured volume fo water that drained into bottles between sampling days

≠ Surface water EC not reported due to a malfunction in the meter which yielded inconsistent values at low levels of EC

[1] Procedure changed on day 7 - sample bottles were removed between sampling days

Table 6. Continued.

Sample depth below soil surface(cm)	Day 9		Day 11		Day 14		Day 16	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.340	—	.344	—	.366	—	.373	—
6.5	2.76	7.4	2.78	7.4	2.12	9.8	2.12	6.9
16.6	6.86	9.0	6.60	6.2	6.24	12.9	5.38	11.5
26.8	15.70	6.1	15.70	6.4	—	3.0	15.60	8.8
37.0	20.90	19.2	20.00	10.2	19.8	17.6	20.00	14.4
47.1	32.30	8.6	31.20	9.2	—	17.6	33.70	21.6
57.3	39.30	11.5	39.80	14.6	40.1	11.0	40.90	11.8
Drop in surface water volume between sampling days	400 mls		No change		460 mls		No change	

Table 6. Continued.

Sample depth below soil surface(cm)	Day 18		Day 21		Day 25		Day 30	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.388	—	.385	—	.396	—	.409	—
6.5	1.88	7.4	1.79	12.8	1.55	7.6	1.40	11.8
16.6	5.24	8.8	5.02	11.2	3.93	7.6	4.11	10.8
26.8	14.10	5.3	14.70	8.6	12.05	5.5	12.50	8.4
37.0	20.90	10.9	20.60	15.1	20.05	14.2	19.85	15.3
47.1	33.20	20.8	32.10	13.5	29.00	14.7	29.95	18.5
57.3	40.20	11.0	40.30	13.5	40.10	10.6	38.80	16.7
Drop in surface water volume between sampling days	100 mls		260 mls		100 mls		300 mls	

Table 7. Observed results, Core 3, Phase I (all EC data expressed in millimhos/cm).

Sample depth below soil surface(cm)	Initial		Day 2		Day 4		Day 7[1]	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.204	—	≠	—	≠	—	≠	—
4.5	3.21	25.0*	2.89	50.0*	1.98	50.0*	1.25	80.0*
14.7	—	3.0	7.22	15.0	6.57	15.0	no	15.0
24.9	22.2	10.0	21.75	20.0	19.50	20.0	16.20	23.0
35.0	32.35	10.0	30.10	15.0	29.00	15.0	25.60	18.0
45.2	32.00	20.0	34.10	40.0	31.60	40.0	29.00	45.0
55.3	40.10	15.0	40.30	20.0	39.60	20.0	37.40	25.0
65.5	40.10	35.0	44.50	60.0	43.40	60.0	41.70	70.0
observed drop in surface water volume between sampling days	—		No change		200 mls		420 mls	

---

* Sample volumes are estimates due to the unmeasured volume of water than drained into the sample bottles between sampling days

≠ Surface water not reported due to a malfunction in the meter which yielded inconsistent values at low levels of EC

[1] Procedure changed at day 7 - sample bottles were removed between sampling days

Table 7. Continued.

Sample depth below soil surface(cm)	Day 9 EC	Day 9 Sample Vol(mls)	Day 11 EC	Day 11 Sample Vol(mls)	Day 14 EC	Day 14 Sample Vol(mls)	Day 16 EC	Day 16 Sample Vol(mls)
surface water	.300	---	.296	---	.320	000	.327	---
4.5	0.92	17.2	1.11	11.2	1.30	17.8	1.32	11.8
14.7	5.02	6.9	4.70	10.3	---	3.0	4.26	8.9
24.9	13.40	12.0	13.50	10.6	12.80	16.2	12.20	13.5
35.0	23.20	11.40	24.20	13.8	24.00	8.0	25.30	15.4
45.2	28.80	24.0	29.20	20.5	29.90	29.8	30.20	33.1
55.3	35.50	9.6	35.30	11.6	37.70	10.0	37.80	7.6
65.5	39.20	40.0	41.10	16.4	41.40	23.4	41.90	19.2
observed drop in surface water volume between sampling days	400 mls		No change		580 mls		No change	

Table 7. Continued.

Sample depth below soil surface(cm)	Day 18		Day 21		Day 25		Day 30	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.342	—	.344	—	.358	—	.369	—
4.5	1.34	16.6	1.27	16.9	1.35	10.3	1.23	20.0
14.7	4.11	7.11	4.13	16.7	3.81	11.0	3.61	8.5
24.9	10.90	9.3	11.10	16.5	10.80	8.6	10.40	13.60
35.0	25.10	9.6	23.85	9.1	23.20	6.8	22.55	12.50
55.3	37.40	14.60	36.30	14.5	33.20	9.2	36.50	16.40
65.5	42.30	16.5	41.20	51.0	40.70	16.9	39.70	42.0
observed drop in surface water volume between sampling days	200 mls		200 mls		160 mls		230 mls	

Table 8. Observed results, Core 1, Phase II (all EC data expressed in millimhos/cm).

Sample depth below soil surface(cm)	Initial		Day 4		Day 7		Day 13		Day 16	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.445	---	.501	---	.442	---	---	---	.508	---
7.6	1.50	10.0	1.37	5.1	1.42	10.2	1.41	6.4	1.42	11.6
17.8	4.69	6.5	4.01	5.5	4.04	7.9	---	3.5	4.10	8.0
27.9	11.50	9.0	10.50	5.6	11.30	10.0	10.30	5.2	11.00	7.2
38.1	20.50	14.2	20.40	9.7	20.70	13.2	20.80	14.9	21.00	20.0
48.3	30.50	8.0	31.30	10.2	31.05	12.4	32.10	13.0	32.60	10.4
58.4	41.0	15.4	41.0	6.9	41.50	13.9	43.50	15.0	43.30	12.8
68.6	46.10	28.8	46.40	6.8	46.00	24.1	46.00	11.2	46.70	26.8
observed drop in surface water volume between sampling days	---		100 mls		80 mls		90 mls		240 mls	

Average drop in surface water volume per day	=	33.0 mls/day
Average evaporation loss per day	=	25.5 mls/day
Average flow into column	=	7.5 mls/day

Table 8. Continued.

Sample depth below soil surface(cm)	Day 19		Day 23		Day 27		Day 30	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.483	–––	.480	–––	.497	–––	.516	–––
7.6	1.41	11.4	1.47	8.8	1.40	8.0	1.38	11.5
17.8	4.07	5.8	4.10	6.0	4.03	6.2	3.88	31.6
27.9	10.50	5.4	10.50	6.1	10.90	7.6	11.00	9.0
38.1	21.00	9.0	20.80	7.2	19.20	7.0	21.20	14.0
48.3	33.00	7.4	31.95	6.0	33.70	8.2	34.50	10.0
58.4	43.10	8.8	42.50	7.4	42.20	5.4	44.60	13.7
68.6	46.50	6.2	46.80	17.8	45.90	7.3	47.25	13.2
observed drop in surface water volume between sampling days (minus evaporation)	35 mls		100 mls		210 mls		120 mls	

Table 9. Observed results, Core 2, Phase II (all EC data expressed in millimhos/cm).

Sample depth below soil surface(cm)	Initial		Day 4		Day 7		Day 13		Day 16	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.469	—	.494	—	.508	—	—	—	.549	—
6.5	1.26	10.0	1.31	6.4	1.28	10.0	—	1.5	1.33	10.3
16.6	3.04	8.8	3.16	6.2	3.56	9.5	3.18	11.5	3.22	9.5
26.8	9.61	6.4	—	3.0	9.36	8.6	—	2.0	—	2.0
37.0	18.30	11.8	18.50	8.8	18.60	12.6	18.20	8.2	18.30	9.2
47.1	30.10	17.8	27.90	5.1	30.50	13.2	30.30	15.3	29.50	7.6
57.3	40.10	12.2	39.90	8.9	40.20	12.6	40.40	11.6	39.8	12.2
observed drop in surface water volume between sampling days	—		170 mls		215 mls		215 mls		190 mls	

Average drop in surface water volume per day = 39.5

Average evaporation loss per day = 25.5

Average flow into column = 14.0 mls/day

Table 9. Continued.

Sample depth below soil surface(cm)	Day 19		Day 23		Day 27		Day 30	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.549	---	.560	---	.580	000	.604	---
6.5	1.35	9.5	1.36	5.6	1.29	7.6	1.31	10.4
16.6	3.20	7.2	3.25	8.0	3.39	6.5	3.30	11.40
26.8	9.27	7.8	---	4.5	8.32	4.8	8.86	32.0
37.0	18.60	7.2	18.40	13.3	18.30	8.4	18.60	15.8
47.1	30.00	7.8	30.00	10.0	30.30	8.1	30.50	14.0
57.3	40.30	8.4	39.30	12.8	40.30	8.6	40.40	12.8
observed drop in surface water volume between sampling days	115 mls		165 mls		110 mls		No change	

Table 10. Observed results, Core 3, Phase II (all EC data expressed in millimhos/cm).

Sample depth below soil surface(cm)	Initial		Day 4		Day 7		Day 13		Day 16	
	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)	EC	Sample Vol(mls)
surface water	.461	---	.433	---	.509	---	---	---	.536	---
4.5	1.13	8.4	1.17	5.9	1.05	11.2	1.20	16.4	1.12	10.8
14.7	4.12	6.5	3.62	6.0	3.65	10.8	3.76	9.6	3.82	13.0
24.9	9.54	13.6	9.76	7.3	9.63	12.3	---	2.0	9.90	7.6
35.0	20.70	17.7	20.70	8.3	20.60	9.4	20.40	13.8	20.30	8.0
45.2	29.60	9.7	30.00	9.8	30.70	12.7	30.40	18.3	30.60	17.0
55.3	38.40	9.4	38.80	10.0	39.40	11.8	39.10	14.2	39.00	8.0
65.5	44170	11.2	44.50	6.9	44.30	12.1	44.90	14.2	44.60	11.4
observed drop in surface water volume between sampling days	---		125 mls		No change		145 mls		280 mls	

Average drop in surface water volume per day = 33.5

Average evaporation loss per day = 25.5

Average flow into column = 8.0 mls/day

Table 10. Continued.

Sample depth below soil surface(cm)	Day 19 EC	Sample Vol(mls)	Day 23 EC	Sample Vol(mls)	Day 27 EC	Sample Vol(mls)	Day 30 EC	Sample Vol(mls)
surface water	.539	---	.552	---	.559	---	.573	---
4.5	1.18	6.2	1.17	7.9	1.19	9.4	1.12	11.0
14.7	3.83	7.4	3.74	6.3	---	4.8	3.99	8.8
24.9	9.72	8.6	10.60	8.8	10.00	8.2	9.99	13.2
35.0	19.20	5.4	20.30	6.3	20.40	6.1	20.50	7.3
45.2	31.90	15.3	31.60	11.6	28.40	15.5	31.30	10.8
55.3	40.10	9.4	39.40	8.2	39.40	9.9	40.20	11.8
65.5	44.80	7.5	44.55	7.6	44.50	8.0	45.45	15.7
observed drop in surface water volume between sampling days	75 mls		55 mls		170 mls		155 mls	

APPENDIX II. Development of a computer model.

APPENDIX II

As explained in the theory section, the relationship which describes the change in salt content of a given soil section is given as:

$$F(C_1 - C_2) = SW(\Delta C_s) \tag{A1}$$

If equation A1 is to be compatible with a computer program, C_1, C_2, and C_s must be defined in more specific terms. For example, C_1 is assumed to be the average salt concentration entering the section between sampling days:

$$C_1 = \frac{C_{11} + C_{12}}{2} \tag{A2}$$

where C_{11} = inflow salt concentration at time 1

C_{12} = inflow salt concentration at time 2

Similarly, C_2, the outflow salt concentration is defined as:

$$C_2 = \frac{C_{21} + C_{22}}{2} \tag{A3}$$

where C_{21} = outflow salt concentration at time 1

C_{22} = outflow salt concentration at time 2

C_s can be defined as the difference between the average salt concentration existing in the section at time 2 and the average salt concentration at time 2.

$$\Delta C_s = \frac{C_{12} + C_{22}}{2} - \frac{C_{11} + C_{21}}{2} \tag{A4}$$

Thus, equation 3 can be written as:

$$F\left(\frac{C_{11} + C_{12}}{2} - \frac{C_{21} + C_{22}}{2}\right) = SW\left(\frac{C_{12} + C_{22}}{2} - \frac{C_{11} + C_{21}}{2}\right) \qquad \text{(A5)}$$

If the initial salt concentrations are known (in this case C_{11} and C_{21}) and the inflow concentration at time 2 is known (C_{12}), then equation 7 can be solved for the outflow concentration at time 2 (C_{22}).

In using this model on the actual soil core data, the input data included the initial EC of each sampler in the column, the surface water EC on each sampling day, the water content for each section, the average flow of water through the column in milliliters per day, and the volume of sample taken on each day. Starting with the first sampling day after the initial EC readings, the concentration in the top sampling port of the soil column was calculated using equation A5. In this case, the inflow salt concentration at time 2, C_{12}, is the salt concentration in the surface water at time 2. In the next iteration, the salt concentration at the second sampling port is calculated using the outflow salt concentration just calculated as the new inflow concentration.

This is continued until a concentration at each depth in the column is calculated. The computer then moves to the next sampling day and repeats the process using the calculated values from the previous day to predict the next days values. Thus, an array is established in the computer, similar to that shown below:

$$\begin{array}{rlllll} & C_{11} & C_{12} & C_{13} \; . \; . \; . \; . \; . & C_{1N} & \text{sampling days} \\ \text{sampling} & C_{21} & C_{22} & & & \\ \text{port} & C_{31} & & & & \\ \text{depths} & & & & & \\ & C_{M1} & & & & \end{array}$$

The array is filled in step by step with predicted concentrations. All the concentrations are then converted to EC. A listing of the FORTRAN program is given to further illustrate the model.

```
                 B6700  F O R T R A N   C O M P I L A T I O N   M A R K   2.9.19
                                                                 G O
                                                                 = =

C
C      PROGRAM NAME - MARSH SOIL SALINITY
C
C
C      THIS PROGRAM IS DESIGNED TO DETERMINE SOIL WATER EC
C      AT EACH SAMPLE DEPTH ON EACH SAMPLE DAY
C      OVER THE ENTIRE DURATION OF THE EXPERIMENT.
C
C      N=NUMBER OF SAMPLE DAYS
C      M=NUMBER OF SAMPLE DEPTHS
C      ECI=INITIAL EC OF SOIL WATER SAMPLES
C      SURE=SURFACE WATER EC OVER THE DURATION OF THE EXPERIMENT
C      SAMP=VOLUME OF SAMPLE TAKEN FROM EACH PORT
C      DAYS=NUMBER OF DAYS BETWEEN EACH SAMPLING
C      VP=TOTAL VOLUME OF WATER ASSUMED TO BE IN EACH LAYER
C      D=SAMPLE DEPTHS
C      C=BRINE CONCENTRATION IN MEQ/LIT
C

       DIMENSION ECI(7),C(9,11),SURE(11),DAYS(10),SAMP(8,10),VP(7),D(8)
       N=9
       M=7
       READ(5,20)(ECI(I),I=1,M)

       READ(5,10)(SURE(I),I=1,N)
       READ(5,21)((SAMP(I,J),I=1,M+1),J=1,N-1)
       WRITE(6,602)((SAMP(I,J),J=1,N-1),I=1,M+1)

   602 FORMAT(9F6.2)
       READ(5,30)(DAYS(I),I=1,N-1)
       WRITE(6,35)(DAYS(I),I=1,N-1)
    35 FORMAT(8F5.0)
       READ(5,40)(VP(I),I=1,M)
       READ(5,50)(D(I),I=1,M+1)
       READ(5,60)(C(M+2,J),J=1,N)
    10 FORMAT(11F5.3)
    20 FORMAT(7F6.2)
    21 FORMAT(8F5.2)
    30 FORMAT(10F2.0)
    40 FORMAT(7F6.0)
    50 FORMAT(8F5.1)
    60 FORMAT(11F6.0)
C
C       CHANGE SAMPLE EC TO MEQ/LIT
C
       DO 5 I=1,M
       Z=(ALOG10(ECI(I))*1.095)+1.
       EQL=10.**Z
     5 C(I+1,1)=EQL
C
C       CHANGE SURFACE WATER EC TO MEQ/LIT
C
       DO 15 I=1,N
       Z=(ALOG10(SURE(I))*1.095)+1.
       EQL=10.**Z
    15 C(1,I)=EQL
```

```
C
C      CALCULATE CONCENTRATIONS OVER DURATION OF EXPERIMENT
C
       DO 999 J=1,N-1
       FPD=7.5
       F=FPD*DAYS(J)
       FS=F
       DO 25 I=1,M
       X1=FS/VP(I)
       X2=C(I,J)+C(I,J+1)-C(I+1,J)
       X3=C(I,J)-C(I,J+1)+C(I+1,J)
       X4=X1+1.
       C(I+1,J+1)=((X1*X2)+X3)/X4
       WRITE(6,223)I+1,J+1,FPD,FS,C(I+1,J+1),SAMP(I+1,J)
   223 FORMAT(1X,I3,I4,4F8.2)
       FS=FS-SAMP(I+1,J)
C
C      IF VOLUME OF WATER WITHDRAWN IN SAMPLES IS GREATER THAN THE TOTAL  VOLUME
C      OF WATER FLOWING THROUGH THE COLUMN, THEN WATER WILL BE DRAWN FROM
C      THE BRINE RESEVOIR.
C
       IF(FS.LE.0.)GO TO 55
    25 CONTINUE
       IF(I.EQ.M+1)GO TO 999
    55 K=I+2
       FS=0
       DO 65 L=K,M+1
    65 FS=FS+SAMP(L,J)+10.0
       I=M+1
       DO 75 L=K,M+1
       X1A=FS/VP(I-1)
       X2A=C(I+1,J)+C(I+1,J+1)-C(I,J)
       X3A=C(I,J)+C(I+1,J)-C(I+1,J+1)
       X4A=X1A+1.
       C(I,J+1)=((X1A*X2A)+X3A)/X4A
       WRITE(6,223)I+1,J+1,FPD,FS,C(I+1,J+1),SAMP(I+1,J)
       FS=FS-SAMP(I+1,J)-10.0
    75 I=I-1
   999 CONTINUE
       WRITE(6,240)
   240 FORMAT(1H1)
       WRITE(6,200)
   200 FORMAT(3X,"CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF
     C EXPERIMENT"/)
       TOTD=0.
       DO 9 I=1,N-1
       TOTD=TOTD+DAYS(I)
     9 DAYS(I)=TOTD
       WRITE(6,205)
       WRITE(6,210)(DAYS(I),I=1,N-1)
   210 FORMAT(3X,"DEPTH",1X,"INITIAL",3X,F2.0,10F8.0/)
   205 FORMAT(45X,"ELAPSED TIME IN DAYS"/)
C
C      PRINT RESULTS - CONTENTS OF CONCENTRATION ARRAY
C
       WRITE(6,101)(D(I),(C(I,J),J=1,N),I=1,M+1)
   101 FORMAT(1X,F5.1,9F8.2/)
   100 FORMAT(1X,F5.1,9F8.3/)
C
C      CHANGE MEQ/LIT BACK TO EC
C
```

```
      DO 799 J=1,N
      DO 45 I=1,M+1
      X=(ALOG10(C(I,J))-1.)/1.095
   45 C(I,J)=10.**X
  799 CONTINUE
      WRITE(6,220)
  220 FORMAT(/)
      WRITE(6,230)
  230 FORMAT(3X,"EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPER IMENT
     CIMENT"/)
      WRITE(6,295)
      WRITE(6,210)(DAYS(I),I=1,N-1)
C
C      PRINT RESULTS - EC IN CONCENTRATION ARRAY
C
      WRITE(6,100)(D(I),(C(I,J),J=1,N),I=1,M+1)
      END
```

APPENDIX III. Predicted results of soil core experiment.

Table 11. Predicted results, Core 1, Phase I.

CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS									
DEPTH	INITIAL	4.	7.	9.	11.	14.	16.	18.	21.	25.	30.
0.0	2.58	2.64	2.74	2.81	2.80	3.06	3.21	3.27	3.40	3.45	3.56
7.6	46.01	33.39	20.88	15.58	13.68	11.12	9.79	8.75	7.46	6.29	5.26
17.8	185.53	162.20	126.50	101.57	91.46	78.30	70.51	63.29	53.55	42.50	31.91
27.9	334.97	323.17	297.79	271.51	258.61	238.74	225.53	212.72	192.69	166.23	136.45
38.1	437.17	427.38	414.60	404.50	399.69	391.27	384.80	377.89	364.58	342.52	311.90
48.3	533.77	525.94	509.15	494.63	488.85	481.53	478.27	475.40	471.44	463.52	452.48
58.4	618.59	613.39	605.92	597.19	592.69	584.60	579.34	573.66	564.02	550.59	537.21
68.6	688.62	685.46	676.78	669.01	666.56	663.85	663.32	662.37	659.17	649.14	634.44

EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS									
DEPTH	INITIAL	4.	7.	9.	11.	14.	16.	18.	21.	25.	30.
0.0	0.290	0.296	0.307	0.314	0.313	0.339	0.354	0.360	0.373	0.378	0.389
7.6	4.030	3.008	1.959	1.499	1.331	1.102	0.981	0.885	0.765	0.655	0.556
17.8	14.400	12.737	10.150	8.306	7.548	6.550	5.952	5.392	4.630	3.749	2.885
27.9	24.700	23.905	22.184	20.389	19.502	18.129	17.211	16.316	14.907	13.026	10.877
38.1	31.500	30.855	30.011	29.343	29.024	28.465	28.035	27.575	26.687	25.208	23.141
48.3	37.800	37.293	36.205	35.260	34.884	34.407	34.194	34.006	33.748	33.229	32.506
58.4	43.250	42.917	42.440	41.881	41.593	41.074	40.737	40.371	39.752	38.887	38.023
68.6	47.700	47.500	46.951	46.458	46.303	46.131	46.097	46.037	45.834	45.196	44.261

Table 12. Predicted results, Core 2, Phase I.

CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS									
DEPTH	INITIAL	4.	7.	9.	11.	14.	16.	18.	21.	25.	30.
0.0	2.48	2.66	2.90	3.07	3.11	3.33	3.40	3.55	3.52	3.63	3.76
6.5	84.87	57.47	32.19	22.38	18.84	14.58	12.47	10.70	9.86	7.01	5.58
16.5	115.89	130.33	124.05	107.78	99.39	87.03	79.09	71.69	61.41	49.23	37.11
26.8	278.97	233.07	205.98	196.23	191.37	183.27	177.51	170.74	159.73	142.32	120.71
37.0	432.61	438.79	397.94	356.53	340.44	318.87	307.82	297.43	283.42	264.56	242.05
47.1	533.77	511.87	523.47	523.75	517.56	503.55	493.43	480.50	460.17	429.13	394.31
57.3	608.42	619.20	589.11	574.56	574.98	578.55	582.28	586.43	583.34	584.23	568.54

EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS									
DEPTH	INITIAL	4.	7.	9.	11.	14.	16.	18.	21.	25.	30.
0.0	0.280	0.298	0.323	0.340	0.344	0.366	0.373	0.388	0.385	0.396	0.409
6.5	7.050	4.938	2.909	2.067	1.783	1.411	1.224	1.064	0.895	0.723	0.587
16.5	9.370	10.431	9.970	8.769	8.144	7.214	6.610	6.043	5.246	4.287	3.312
26.8	20.900	17.736	15.843	15.157	14.814	14.240	13.831	13.348	12.560	11.304	9.725
37.0	31.200	31.606	28.908	26.148	25.069	23.614	22.865	22.160	21.204	19.912	18.359
47.1	37.800	36.381	37.133	37.151	36.750	35.841	35.182	34.339	33.010	30.971	28.667
57.3	42.600	43.289	41.364	40.430	40.457	40.686	40.925	41.192	41.314	41.051	40.042

Table 13. Predicted results, Core 3, Phase I.

CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS									
DEPTH	INITIAL	4.	7.	9.	11.	14.	16.	18.	21.	25.	30.
0.0	1.89	2.11	2.45	2.67	2.64	2.87	2.94	3.09	3.11	3.25	3.36
4.5	31.60	18.23	8.39	5.63	4.82	3.82	3.51	3.24	3.17	3.07	3.13
14.7	87.12	86.36	73.70	60.53	53.65	44.72	39.34	34.81	28.56	21.82	15.80
24.9	291.42	247.27	205.46	180.94	169.10	150.03	138.99	128.45	113.53	93.27	71.88
35.0	415.94	428.13	407.56	378.24	363.08	339.74	325.23	310.55	288.85	255.44	216.65
45.2	476.83	455.69	462.90	471.99	474.01	470.31	467.82	462.91	452.20	430.01	397.70
55.3	572.54	578.38	545.89	521.92	515.57	512.08	511.34	511.19	513.52	514.63	511.67
65.5	638.20	625.61	642.15	642.38	638.59	622.73	616.08	606.67	592.18	571.63	556.89

EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS									
DEPTH	INITIAL	4.	7.	9.	11.	14.	16.	18.	21.	25.	30.
0.0	0.218	0.241	0.277	0.299	0.296	0.320	0.327	0.342	0.344	0.358	0.369
4.5	2.860	1.731	0.852	0.591	0.514	0.416	0.384	0.358	0.350	0.340	0.346
14.7	7.220	7.163	6.197	5.178	4.637	3.927	3.494	3.124	2.607	2.039	1.518
24.9	21.750	18.720	15.807	14.075	13.231	11.865	11.062	10.293	9.199	7.684	6.057
35.0	30.100	30.904	29.546	27.599	26.586	25.021	24.043	23.051	21.575	19.284	16.591
45.2	34.100	32.716	33.189	33.784	33.915	33.674	33.511	33.190	32.488	31.029	28.893
55.3	40.300	40.675	38.583	37.033	36.621	36.395	36.347	36.337	36.489	36.560	36.368
65.5	44.500	43.698	44.751	44.767	44.525	43.514	43.089	42.488	41.561	40.241	39.293

Table 14. Predicted results, Core 1, Phase II.

CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

ELAPSED TIME IN DAYS

DEPTH	INITIAL	4.	7.	13.	16.	19.	23.	27.	30.
0.0	4.12	4.69	4.09	4.46	4.76	4.51	4.48	4.65	4.85
7.6	15.59	14.59	14.89	13.91	13.34	13.33	13.02	12.52	12.10
17.8	54.32	54.07	53.17	52.25	52.29	51.75	51.03	50.47	50.35
27.9	145.03	143.09	143.49	140.47	140.11	140.07	139.10	137.92	138.30
38.1	273.13	272.82	268.40	266.66	276.27	277.63	277.01	276.75	282.24
48.3	422.00	421.52	433.01	431.56	454.63	459.29	459.46	459.44	466.27
58.4	585.00	584.00	584.56	584.88	594.52	597.98	599.93	601.10	603.30
68.6	663.36	666.67	675.98	679.74	694.90	698.92	701.28	702.91	707.88

EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

ELAPSED TIME IN DAYS

DEPTH	INITIAL	4.	7.	13.	16.	19.	23.	27.	30.
0.0	0.445	0.501	0.442	0.478	0.508	0.483	0.480	0.497	0.516
7.6	1.500	1.412	1.438	1.352	1.301	1.301	1.272	1.228	1.190
17.8	4.690	4.671	4.599	4.527	4.530	4.487	4.430	4.385	4.376
27.9	11.500	11.359	11.388	11.169	11.143	11.140	11.070	10.984	11.011
38.1	20.500	20.479	20.175	20.056	20.715	20.808	20.766	20.748	21.123
48.3	30.500	30.468	31.226	31.131	32.647	32.952	32.964	32.962	33.409
58.4	41.100	41.036	41.071	41.092	41.710	41.932	42.057	42.131	42.592
68.6	46.100	46.310	46.900	47.138	48.097	48.352	48.501	48.603	48.917

Table 15. Predicted results, Core 2, Phase II.

CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS							
DEPTH	INITIAL	4.	7.	13.	16.	19.	23.	27.	30.
0.0	4.36	4.62	4.76	4.97	5.19	5.19	5.30	5.51	5.76
6.5	12.88	11.94	11.35	10.37	9.83	9.54	9.08	8.58	8.15
16.6	33.79	33.49	33.27	32.14	31.87	31.33	30.52	29.82	29.46
26.8	119.15	114.90	112.86	107.09	105.31	103.55	100.57	97.65	96.32
37.0	241.20	239.36	239.31	234.33	233.15	232.21	229.22	226.31	232.76
47.1	415.94	410.76	410.65	402.31	401.41	400.20	397.91	394.63	396.43
57.3	569.43	569.80	574.34	573.80	575.78	577.74	577.08	576.08	539.74

EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

		ELAPSED TIME IN DAYS							
DEPTH	INITIAL	4.	7.	13.	16.	19.	23.	27.	30.
0.0	0.469	0.494	0.508	0.528	0.549	0.549	0.560	0.580	0.604
6.5	1.260	1.176	1.123	1.034	0.985	0.953	0.916	0.869	0.830
16.6	3.040	3.016	2.998	2.904	2.882	2.838	2.770	2.712	2.682
26.8	9.610	9.296	9.146	8.718	8.585	8.454	8.232	8.013	7.914
37.0	18.300	18.172	18.169	17.823	17.741	17.676	17.468	17.265	17.714
47.1	30.100	29.758	29.750	29.198	29.138	29.058	28.906	28.688	28.308
57.3	40.100	40.124	40.416	40.381	40.508	40.634	40.591	40.528	41.404

Table 16. Predicted results, Core 3, Phase II.

```
CONCENTRATION IN MEQ/LIT AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

                                           ELAPSED TIME IN DAYS

DEPTH INITIAL    4.       7.       13.      16.      19.      23.      27.      30.
0.0     4.28    4.00     4.77     4.92     5.05     5.08     5.22     5.29     5.43

 4.5   11.43   11.20    10.09     9.41     9.05     8.81     8.43     8.14     7.85

14.7   47.13   46.33    46.93    46.31    46.13    45.63    45.03    44.40    44.16

24.9  118.20  117.15   116.37   115.03   115.19   114.75   113.77   112.81   112.68

35.0  276.05  274.61   271.48   269.01   270.00   270.09   269.35   268.46   271.61

45.2  408.38  409.03   417.72   418.98   424.56   427.51   427.69   427.85   430.21

55.3  543.05  541.89   542.28   542.69   545.58   548.11   549.18   550.61   556.37

65.5  641.34  645.82   655.50   660.49   668.52   673.91   676.82   679.77   685.93


EC IN MILLIMHOS/CM AT SOIL DEPTHS FOR DURATION OF EXPERIMENT

                                           ELAPSED TIME IN DAYS

DEPTH INITIAL    4.       7.       13.      16.      19.      23.      27.      30.
0.0    0.461   0.433    0.509    0.523    0.536    0.539    0.552    0.559    0.573

 4.5   1.130   1.109    1.008    0.946    0.913    0.891    0.855    0.828    0.802

14.7   4.120   4.056    4.104    4.054    4.040    4.000    3.952    3.902    3.882

24.9   9.540   9.463    9.405    9.306    9.318    9.286    9.213    9.142    9.133

35.0  20.700  20.602   20.387   20.218   20.285   20.291   20.241   20.180   20.396

45.2  29.600  29.643   30.217   30.301   30.669   30.864   30.876   30.886   31.041

55.3  38.400  38.325   38.350   38.376   38.563   38.726   38.796   38.888   39.259

65.5  44.700  44.985   45.600   45.918   46.427   46.769   46.953   47.140   47.530
```

APPENDIX IV. Field data.

Table 17. Inflow - Outflow - EC Data.

Date	Inflow EC mmhos/cm	Average Outflow EC mmhos/cm	Outflow Rate m^3/sec
April 25	.348	.564	2.44
May 2	.288	.533	2.93
May 9	.342	.650	3.78
May 16	.304	.615	3.30
May 23	.370	.584	2.37
May 30	.421	.534	1.54
June 6	.423	.442	1.35
June 13	.401	.432	2.02
June 20	.461	.466	1.78
July 5	.808	.651	1.14
July 11	1.030	.855	.87
July 19	.766	.788	.84
July 26	.457	.431	.97
August 1	.648	.727	1.11
August 9	.649	.786	1.12
August 16	.570	1.028	1.53
August 22	.781	1.292	1.33
September 7	1.000	1.246	1.35
September 13	.716	1.433	1.60
September 21	.647	1.273	2.40
September 27	.585	1.103	2.36
October 11	.657	.952	2.25
October 18	.606	.806	2.09
October 25	.632	.728	1.69
November 1	.647	.829	2.17
November 9	.621	.873	2.06
November 22	.500	.932	2.58
November 29	.527	.907	2.65

Table 18. Conductivity variation of surface and interstitial water. Vertical profiles in selected locations in Unit I-E, Ogden Bay Refuge, Utah, 1978, as shown in Figures 15, 16, and 29.

Station	Depth (In.)	3/31	4/11	4/14	4/28	5/12	5/27	6/9	6/20	7/5	7/19	8/2	8/15	8/30	9/13	9/29	10/11
1W	0	--	--	--	--	393	--	--					1194			768	772
Bulrush	6	2960	2890	2203	2696	2036	2736	2782	2788	2696	2668	1997	2316	2420	2690	2786	3072
Shallow	15	6260	7880	5539	3210	6657	7038	7394	7470	7656	8176	7245	6282	5031	5317	5080	5391
	24	1250	14500	12179	7330	13253	13650	13980	13803	13741	14280	11752	11780	10234	10307	10383	10676
2W	0	--	--	--	--	410							8108			901	930
Cattail	6	6490	11900	4294	1877	--	3884	5335	4971	5586	--	--	26818	23918	19623	16619	13643
Shallow	15	19300	20700	19482	20590	15836	21570	21495	21191	19979	20357	--	31390	31252	31765	31688	31808
	24	31300	32500	30090	25680	33250	33590	34223	33820	30544	30992	35700	41217	40100	40278	39620	39437
3W	0	--	--	--	--	853							6210		2683	1750±	1250±
Phrag-	6	28700	29400	25908	21614	27350	26650	26518	25223	23593	--		37562	49446	50549	48950	47233
mites	15	38400	36900	33456	26108	36420	31650	36409	35404	31977	--		49274	46638	47717	47578	47543
	24	37800	43700	39924	29211	40690	41328	41720	41135	37041	38480	19674	44892	45456	47463	46575	46548
4W	0	--	--	--	--	2843	--						15712		13875	4900±	
Phrag-	6	8940	9140	6599	4751	7014	6309	7984	6311	13235	--	--	24560	32562	32461	31238	
mites	15	--	9950	12036	9876	13040	--	14888	14288	17377	8809	--	14053	24192	25894	25816	
	24	28900	28900	28764	23754	30620	26140	30890	31362	28209	31613	--	28750	24069	23598	23570	
5W	0	--	--	--	--	656	1673	1085	1694	1295	--		2422	5765	1879	1113	
Water	6	8090	33100	27540	23433	25440	15435	26961	26250	23638	--		53014	49372	49257	45600	
over	15	10500	56600	53448	27499	57540	57700	61904	56967	46085	--		79561	79561	61403		
"flat"	24	--	62600	58956	42586	63750	67185	69564	68131	66515	65490		85926	85615	70834		
6W	0	--	--	--	--	509	496	506	598	1310	808	676	941	999	1251	766	
Bulrush	6	3030	4860	5722	1744	3113	4853	3312	3241	3421	3167	2837	2612	2495	3030	3115	
Deep	15	5260	5630	3998	3510	4163	4544	4544	4760	4865	5109	5980	5892	5510	6313	6328	
	24	11500	14500	13974	11984	14430	14710	14344	14104	14164	14485	13062	12378	11363	12583	12542	
7W	0	--	--	--	--	418	465	484	536	753	828	752	994	1025	1182	808	
Open-	6	6030	7100	5467	5350	5088	4657	4654	4459	4518	4312	2907	1117	1944	1933	2007	
sago	15	8980	10200	10139	8004	9275	8818	8741	8668	8707	8682	8089	6088	5678	5770	5786	
	24	--	11400	9160	8667	9479	9727	9877	9695	9296	9628	9568	9610	8816	8835	8680	
8W	0	--	--	--	--	368	605	532	517	800	811	767	973	1332	1172	730	800
Cattail	6	6230	8950	6212	4462	5910	6074	6147	6426	6858	6880	5621	3392	3153	3160	3167	3582
deep	15	6410	7530	5712	5650	6238	6350	6575	6742	6842	7148	7431	7738	7610	7452	7452	7505
	24	8880	10700	9792	9042		9600	10012	10038	9972	10406	10250	9104	8994	8673	8012	8356
9W	0	--							--				6567		2812	1000±	753
Alkali	6	--							23522	25336	--	--	80426	56216	56872	52078	47693
Bulrush	15	--							48012	43297	--	--	77976	56330	56352	54327	54912
shallow	24	--							54796	46132	--	--	76102	83021	63174	59576	58422