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A MODEL FOR ESTIMATING TIME-VARIANT RAINFALL INFILTRATION AS A FUNCTION OF ANTECEDENT SURFACE MOISTURE AND HYDROLOGIC SOIL TYPE

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16. Abstract Recent research indicates that the use of remote sensing techniques for the measurement of near surface soil moisture could be practical in the not too distant future. Other research has shown that infiltration rates, especially for average or frequent rainfall events, are extremely sensitive to the proper definition and consideration of the role of the soil moisture at the beginning of the rainfall. Thus, it is important that an easy to use, but theoretically sound, rainfall infiltration model be available if the anticipated remotely sensed soil moisture data is to be optimally utilized for hydrologic simulation.

A series of numerical experiments with the Richards' equation for an array of conditions anticipated in watershed hydrology were used to develop functional relationships that describe temporal infiltration rates as a function of soil type and initial moisture conditions.

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BY

Harold A. Wilkening, III

Robert M. Ragan

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CHAPTER 1

INTRODUCTION

Soil moisture conditions in the near surface soil regime are important drivers to many key hydrologic processes such as infiltration, moisture redistribution, and evapo-transpiration. Recent progress in microwave and thermal inertia techniques show great promise as tools for defining the spatial as well as temporal variations in soil moisture. Most hydrologic models in present use, however, were designed without recognizing the growth of remote sensing technology and are not structured to take full advantage of this information. In fact, models were deliberately simplified in their original development because of the absence of spatial and temporal soil moisture information that remote sensing technology may be able to provide.

The potential use of soil moisture measurements in the proper definition of watershed rainfall-runoff predictions was demonstrated in a recent study by Wilkening (1981) using a physically-based mathematical model to study the infiltration process. The results of Wilkening's numerical simulations showed that rainfall excess volumes from single event storms are quite sensitive to antecedent or initial moisture conditions, especially in more frequent rainfall events. Wilkening showed that rainfall excess predictions were quite sensitive to near surface moisture conditions, soil hydraulic properties and storm characteristics such as intensity, duration and temporal distribution. Using such a rigorous physically-based model, the impact of initial moisture conditions on the infiltration process and resulting rainfall excess can be properly represented. The use of this type of model is not practical on a watershed scale, however, because of the excessive computational costs involved in solving the governing physical equations and the large amount of spatially varying data required to represent the response of soils including the soil moisture-retention curve, the saturated conductivity and the unsaturated conductivity function.

Rather than attempt to directly incorporate rigorous, but cumbersome, mathematical models into applications oriented hydrologic simulation, an alternative approach is to conduct numerical experiments with such models and analyze the results to establish relatively simple functional relationships that accurately describe infiltration capacity within some range of conditions. These functional relationships, describing infiltration capacity in terms of initial moisture conditions, general soil texture classifications, and accomodating temporal rainfall variations, would be computationally efficient and could significantly improve current infiltration components within both single event and continuous simulation models. In particular, these theoretically sound, yet simple infiltration models that logically incorporate the importance of initial soil moisture conditions would be of tremendous value when interfaced with accurate remotely-sensed soil moisture measurements made on watershed basis.

Several efforts have conducted that use numerical simulations of theoretical infiltration events as a tool to develop practical methods for use in watershed analysis. Smith (1972), demonstrated a process of curve fitting to describe the properties of a complex theoretical model of porous media flow. The parameters of a parametric equation were fit to time varying infiltration capacity curves that resulted from numerical simulations using experimental soil data from several different soils. Bloomfield et al., (1981) used numerical simulations to examine the validity of empirical equations relating infiltration capacity to soil moisture storage in the upper soil zone. The Bloomfield group demonstrated that serious conceptual errors in traditional infiltration equations could be improved when the parameters were fit to infiltration-storage relationships derived from numerical simulations.

The objective of the present study was to develop a simple but theoretically sound, infiltration model that defines infiltration capacity in terms of general soil properties available from soil survey maps and near-surface initial soil moisture conditions derived from some antecedent precipitation index, a model state variable or, ideally accurate periodic remotely sensed measurements.

The approach was to use Richard's equation, describing unsaturated porous media flow, to generate a synthetic infiltration series for an array of conditions that could be encountered in the field within each of the SCS hydrologic soil groups. The results are expressed in a general form similar to the infiltration-storage approach used in many single event and continuous simulation models such as USDAHL (Holtan, et al., 1973), HEC1 (U.S. Army Corps of Engineers, 1973), Stanford Watershed Model (Crawford and Linsley, 1966), and SSARR (U.S. Army Corps of Engineers, 1972). Infiltration-storage relationships, in which the infiltration capacity at any time within a rainfall event is defined as a function of the volume of moisture stored in some upper layer, are widely used, especially in continuous simulation models, because infiltration capacity can be related to a period by period accounting of moisture within the model components.

CHAPTER 2

REVIEW OF THE INFILTRATION-STORAGE CONCEPT WITHIN HYDROLOGIC MODELS

The infiltration-storage approach is examined by reviewing the functional form in two different models: USDAHL and SSARR. The underlying assumptions used in these models and all other models employing an infiltration-storage approach are examined.

1. EXAMPLES WITHIN TWO HYDROLOGIC MODELS

1.1. USDAHL Model. Within USDAHL (United States Department of Agriculture Hydrology Laboratory), an agricultural watershed model, Holtan's equation is used to express infiltration capacity ... "as an exhaustion function of available moisture storage in the surface horizon of a soil diminishing to a low constant rate of intake associated with ding strata." The equation, based on physical concepts as shown in Fig. 1, bu rived from field infiltration data is given as:

$$f = a\, S_a^{1.4} + f_c \qquad (1)$$

where: f is the infiltration capacity, in in/hr;

S_a is available soil porosity, undepleted by moisture, in inches;

f_c is rate of intake after prolonged wetting in in/hr;

a is a coefficient of pore space continuity, estimated as the product of vegetative density of crop maturity and stage of growth.

A graphical representation of Eq. (1) and its estimation of a typical time varying infiltration capacity for a soil subject to a steady rainfall is presented

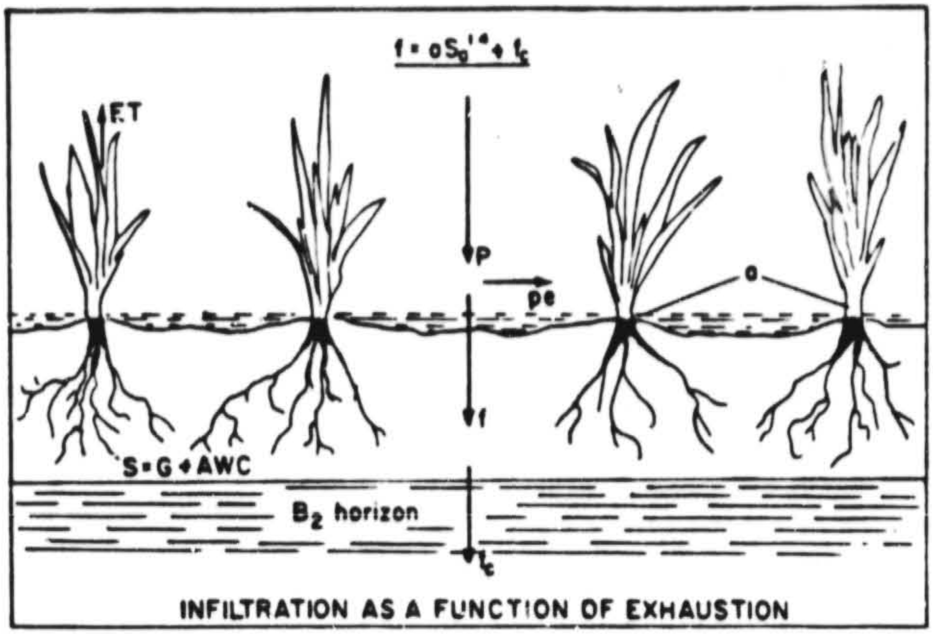

FIGURE 1. PHYSICAL CONCEPT OF HOLTAN'S INFILTRATION MODEL.
(GLYMPH AND HOLTAN, 1969)

as Figure 2. The model rests on the premise that the major zone of hydrologic activity lies above some controlling depth which controls profile drainage so that some potential storage volume, S, can be computed from total porosity and known or assumed antecedent moisture conditions. In USDAHL, primarily used in agricultural applications, the controlling depth, or impeding strata is considered to the B-horizon in agricultural soils.

Extensive information relating the parameters to soil texture classes are available through soil surveys of the Soil Conservation Service (SCS). Values of f_c, the steady state limiting drainage rate in the controlling B-horizon, are given in Table 1 for each SCS hydrologic soil group. The hydrologic capacities of texture classes are defined in Table 2. The total moisture storage capacity, S, is the difference between moisture content at saturation and wilting point. The total storage is divided into large pores, G, which are drained by gravity and plant available porosity, AWC, depleted only by plant uptake. Using this approach the moisture-retention is defined as three points: saturation, field capacity, and wilting point as shown in Fig. 3. Drainage and plant uptake are determined as a function of the current storage values of G and AWC. The vegetative parameter, a, is defined in various landcovers in Table 3. An advantage of this model is that the parameters can be estimated easily when on SCS soil surveys are available.

1.2 Corps of Engineers SSARR. The SSARR (Streamflow Synthesis and Reservoir Regulation) model is a hydrologic model of a river basin throughout which streamflow can be synthsized. The model is structured around three main components: a generalized watershed model, a river system model, and a reservoir regulation model.

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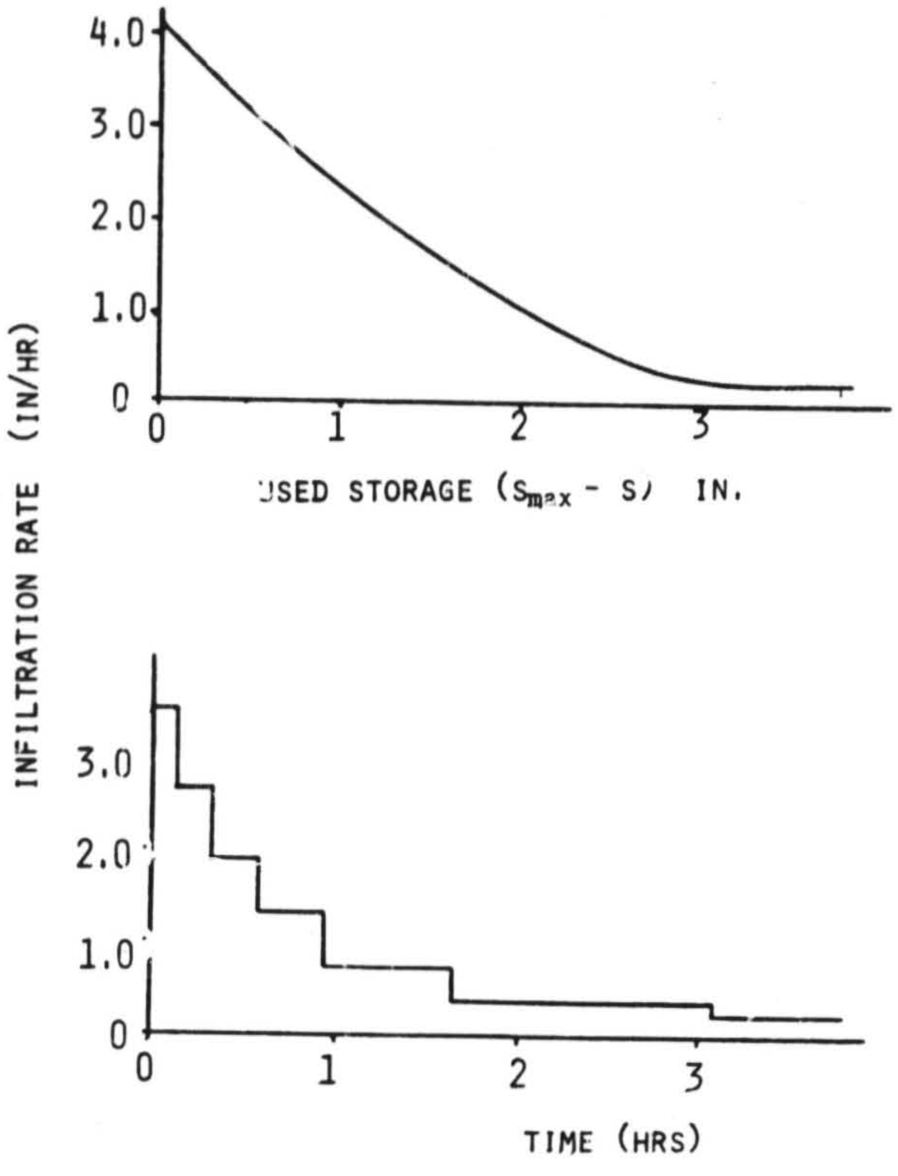

FIGURE 2 EXAMPLE OF HOLTAN'S INFILTRATION EQUATION (UNIQUE FUNCTION FOR PARTICULAR SOIL)

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TABLE 1.--Hydrologic soil groups used by the Soil Conservation Service

Hydrologic soil group	Soils included[1]	Final constant infiltration rate, (fc), in./hr.
A	(Low runoff potential) Soils having high infiltration rates even when thoroughly wetted, consisting chiefly of sands or gravel that are deep and well to excessively drained. These soils have a high rate of water transmission.	0.30 -0.45
B	Soils having moderate infiltration rates when thoroughly wetted, chiefly moderately deep to deep, moderately well to well drained, with moderately fine to moderately coarse textures. These soils have a moderate rate of water transmission.	0.15 - 0.30
C	Soils having slow infiltration rates when thoroughly wetted, chiefly with a layer that impedes the downward movement of water or of moderately fine to fine texture and a slow infiltration rate. These soils have a slow rate of water transmission. (high runoff potential)	0.05 - 0.15
D	Soils having very slow infiltration rates when throroughly wetted, chiefly clay soils with a high swelling potential; soils with a high permanent water table; soils with a clay pan or clay layer at or near the surface; and shallow soils over nearly impervious materials. These soils have a very slow rate of water transmission.	0 to 0.05

TABLE 2.--Hydrologic capacities of texture classes

Texture class	Storage capacity S	Large pores G	Plant-available porosity AWC
	(percent)	(percent)	(percent)
Coarse sand	24.4	17.7	6.7
Coarse sandy loam	24.5	15.8	8.7
Sand	32.3	19.0	13.3
Loamy sand	37.0	26.9	10.1
Loamy fine sand	32.6	27.2	5.4
Sandy loam	30.9	18.6	12.3
Fine sandy loam	36.6	23.5	13.1
Very fine sandy loam	32.7	21.0	11.7
Loam	30.0	14.4	15.6
Silt loam	31.3	11.4	19.9
Sandy clay loam	25.3	13.4	11.9
Clay loam	25.7	13.0	12.7
Silty clay loam	23.3	8.4	14.9
Sandy clay	19.4	11.6	7.8
Silty clay	21.4	9.1	12.3
Clay	18.8	7.3	11.5

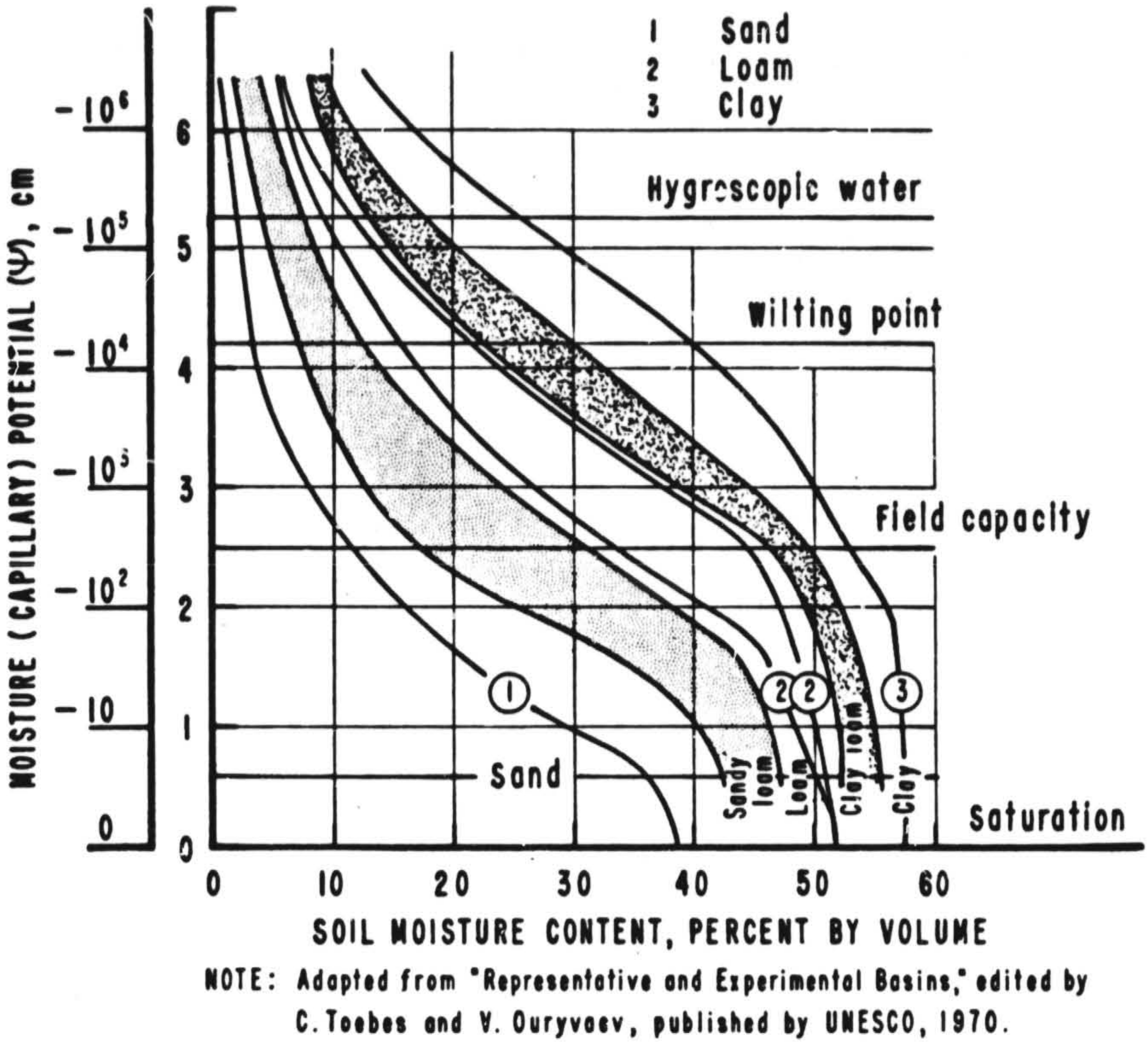

FIGURE 3. SOIL MOISTURE CONSTANTS BY TEXTURE

TABLE 3

VEGETATIVE PARAMETER a IN INFILTRATION EQUATION $f = aS_a^{1.4} + f_c$

Land Use of Cover		Condition & Basal Area Rating*		
Fallow	After Row Crop	0.10	After Sod	0.30
Row Crops	Poor	0.10	Good	0.20
Small Grains		0.20		0.30
Hay (Legumes)		0.20		0.40
Hay (Sod)		0.40		0.60
Pasture (Bunch Grass)		0.20		0.40
Temporary Pasture (Sod)		0.40		0.60
Permanent Pasture (Sod)		0.80		1.00
Woods and Forests		0.80		1.00

*Adjustments needed for "weeds" and "grazing"

(from Glymph and Holtan, 1969)

Within the watershed model, watersheds are separated into relatively homogeneous units for independent anaylsis before they are added to the system. A major element of the watershed model is the soil moisture index (SMI) which functions as an indicator of the relative soil wetness and is used to determine runoff. The percent of the total rainfall input that becomes available for runoff during some period empirically derived relationships of SMI vs. runoff percent (ROP). Because runoff is expressed in terms of rainfall intensity, rainfall intensity may be included as a third variable in SMI-ROP relationships as shown in Fig. 4. The total generated runoff for period (RGP) is computed as:

$$RGP = ROP \times P_n \qquad (2)$$

where P_n is the net precipitation for the period.

The soil moisture index, a state variable representing current moisture conditions in the upper soil layer, is updated for each period by:

$$SMI_2 = SMI_1 + (P_n - RGP) - \frac{PH}{24} \times KE \times ETI \qquad (3)$$

where SMI_1 and SMI_2 are soil moisture indices at the beginning and end of the period; PH is period length in hours; ETI is an evapotranspiration index, in inches per day; and KE is a factor for evaporation reductions on rainy days. When soil moisture conditions are approximately at the wilting point, the SMI is a small number which yields little or no runoff. When precipitation recharges soil moisture, the value of SMI increases until it reaches a maximum value considered to represent the soil field capacity and the runoff percent is assumed to approach 100 percent.

The generated runoff for a period (RGP) represents not only rainfall

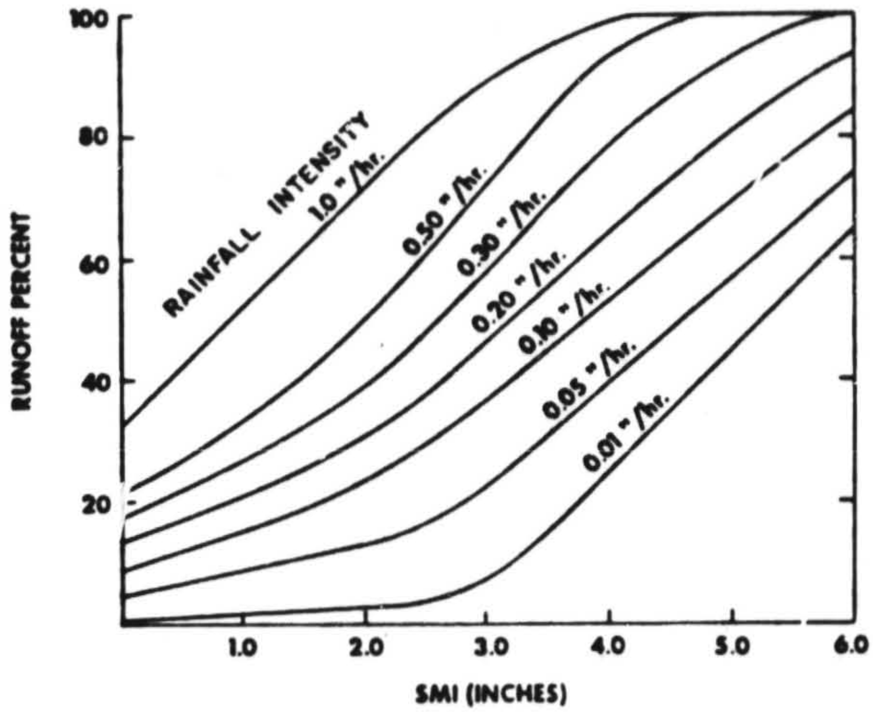

FIGURE 4. THREE VARIABLE SOIL MOISTURE INDEX RELATIONSHIPS DEVELOPED FOR BIRD CREEK. SMI (INCHES) vs. RAINFALL INTENSITY (INCHES PER HOUR) vs. RUNOFF PERCENT

(U.S. Army Corps of Engineers, 1972)

excess not entering the soil regime but also subsurface (upper storage) lateral flow and baseflow runoff. The partitioning of RGP into these three components is accomplished through two additional empirical parameters: a baseflow index and a surface-subsurface runoff ratio relationship.

2. EVALUATION OF INFILTRATION-STORAGE APPROACH

The USDAHL and SSARR models illustrate two methodologies for estimating runoff for a time period by water budget accounting of moisture movement through the various hydrologic components. In both cases, the infiltration capacity is described as a function of the current soil moisture conditions (or at least a parameter related to moisture conditions as in SSARR). This type of approach is advantageous for several reasons:

1) Infiltration recovery during intermittant lulls in a rainfall event or between rainfall events can be accounted for through soil drainage and ET algorithms.
2) Infiltration capacity can be described for non-uniform rainfall distributions.
3) Computations are much easier than a more rigorous mathematical approach to modeling infiltration.
4) Difficult to attain soil data such as the soil moisture retention and unsaturated hydraulic conductivity functions are not required to estimate infiltration rates.

One of the key assumptions of the infiltration-storage relationship is that infiltration capacity can be related to soil moisture storage without regard to the vertical distribution of water in the soil profile. A result of this assumption is the existance of a single-valued unique relation between the infiltration capacity and the volume of stored moisture in some defined surface layer within a particular soil. For example, the estimation of parameters in Holtan's equation is based on soil properties or, as with the ROP vs SMI reltaionship in SSARR, calibrated with rainfall-runoff data to obtain unique infiltration-storage function particular soil or group of soils within a watershed unit.

An examination of the above assumption was made by Bloomfield et. al. (1981) using a numerical solution of the unsaturated flow equation to model ponded infiltration into a semi-infinite soil column. The results of a series of infiltration simulations using a range of uniform initial moisture conditions were expressed in terms of available storage in the soil. An arbitrary effective depth of soil was chosen as 200cm and the initial stored contents (US_o) were calculated for each run as:

$$US_o = 200 (\theta_i - \theta_r) \tag{4}$$

where θ_i is the uniform initial water content and θ_r is the residual water content following drainage. A maximum volume of available storage was defined as:

$$US_{max} = 200 (\theta_{max} - \theta) \tag{5}$$

where θ_{max} is the saturated water content. To calculate successive store contents, the following relation was used:

$$US_{t+1} = US_t + \Delta I_{t+1} \tag{6}$$

where US_t is the store contents at the end of time period t and ΔI_{t+1} is the infiltration volume in time period t+1.

Using Eqs. (4), (5), and (6), Bloomfield expressed the ponded infiltration capacity for each initial moisture condition in terms of an infiltration-storage relationship. Comparison of these relationships, shown in Fig. 5, indicated that when the vertical distribution of soil moisture is considered, the relation between infiltration capacity and soil water storage is not unique for the particular soil but is described by a family of curves dependent on the initial soil water storage or initial uniform moisture content.

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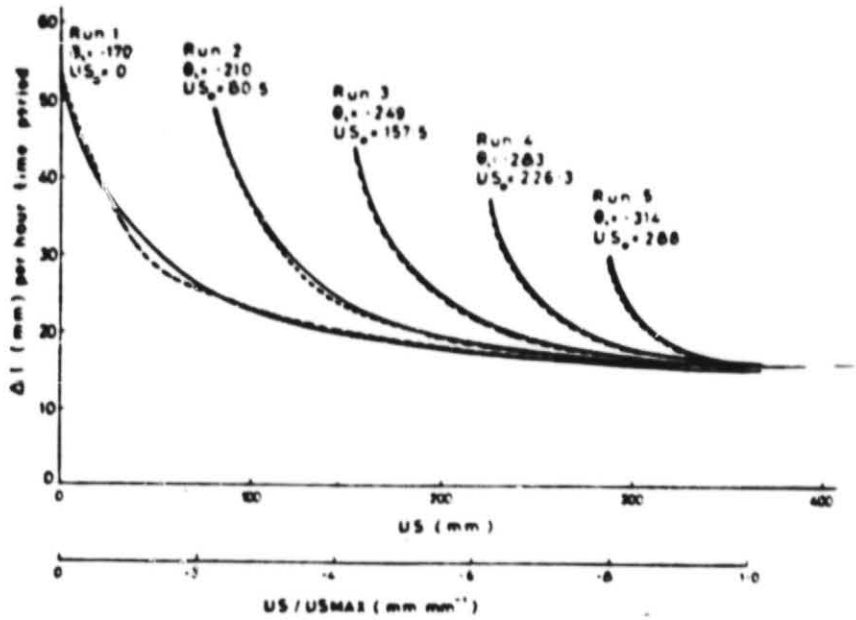

Solid lines indicate numerical results, dashed lines predicted using (9) and (10).

(BLOOMFIELD, PILGRIM AND WATSON, 1981)

FIGURE 5. NONUNIQUE ΔI - US RELATIONSHIPS

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CHAPTER 3

DEVELOPMENT AND USE OF REPRESENTATIVE INFILTRATION-STORAGE RELATIONSHIPS FOR WATERSHED MODELING

1. GENERAL APPROACH

If infiltration estimates made using infiltration-storage retaionships are to be improved, the work of the Bloomfield group shows that the parameters of the function should be defined not only for particular soil properties but also for various initial moisture conditions existing in the soil. This improvement provided by inclusion of initial soil moisture in a functional relationship definition of the impact of watershed moisture conditions on predicted rainfall excess volumes.

It is a difficult task, however, to define the parameters in terms of both soil properties and initial moisture conditions. For example, the calibration of a simple empirically-based infiltration storage equation for a particular watershed would require rainfall-runoff data resulting from an array of initial soil moisture conditions. Extensive field tests would be required to enlarge the already existing results of infiltrometer data on field plots used to define parameters for various soil types (such as Holton's equation).

Another approach, as used in the present study, is to use the theoretical equations of unsaturated moisture flow to simulate rainfall infiltration behavior for an array of different soils and initial moisture conditions.

A similar approach was used by Smith (1972) to develop a parametric infiltration equation to describe the results of theoretical infiltration simulations. By expressing the results in terms of general infiltration-storage relationships, a series of representative functions can be developed. These functions can then be used to define the dependence of the infiltration-storage relationship on the initial moisture for the range of conditions used in the numerical experiments.

2. DEVELOPMENT OF REPRESENTATIVE INFILTRATION-STORAGE FUNCTIONS

2.1 Infiltration Model. An array of infiltration events was simulated using a mathematical model to describe rainfall infiltration into a semi-infinite soil profile. In theory, the one-dimensional infiltration capacity at the soil surface is based on the vertical movement of moisture in the near-surface unsaturated soil regime. Unsaturated moisture flow in a non-swelling soil can be described by Richard's equation as:

$$C(\psi)\frac{\partial}{\partial t} = \frac{\partial}{\partial z}K(\psi) \quad (\frac{\partial\psi}{\partial z} - 1) \qquad (7)$$

in which ψ is the soil water pressure head (cm); $C(\psi)$ is the specific moisture capacity defined as the slope of the moisture-retention curve; $K(\psi)$ is the unsaturated hydraulic conductivity function and z is the depth below the soil surface (cm).

The solution of Richard's equation requires the specification of boundary conditions. The upper boundary condition at the surface until the soil becomes saturated and ponding occurs is:

$$R = K(\psi) \quad (\frac{\partial \psi}{\partial z} - 1) \tag{8}$$

in which R is the rainfall flux at the surface. During this phase of the infiltration event, the infiltration rate is equal to the rainfall rate. After surface ponding occurs; the upper boundary condition becomes:

$$\psi(0,t) = D \tag{9}$$

in which $\psi(0,t)$ is the soil pressure head at the surface and D is the depth of ponded water at the surface. During this phase, the infiltration capacity, R is defined by Darcy's Law as:

$$R = -K_s \quad (\frac{\partial \psi}{\psi z} - 1) \tag{10}$$

in which K_s is the saturated soil hydraulic conductivity.

The lower boundary condition in the soil column can be denoted as some depth, L, below the soil surface where the effect of the infiltration event on the initial soil conditions can be assumed negligible. The lower boundary condition becomes

$$\psi(L,t) = \psi(L,0) \qquad \text{for } 0< t< \infty \tag{11}$$

The initial boundary condition is the initial volumetric moisture content through the soil profile.

An implicit finite difference formulation of Eq. (7) was solved numerically subject to the relevant boundary conditions. An implicit linearization of the non-linear functions was made using the predictor-corrector method. A complete explanation of the formulation of the model equations, the implementation of the numerical solution, and verification of the numerical solutions are given in Wilkening (1981).

The theoretical model predicts infiltration rates into a bare, nonswelling soil based on the following input data: initial soil moisture conditions in the soil at the beginning of the rainfall event; a description of the basic soil hydraulic properties given by the moisture retention function and the unsaturated hydraulic conductivity function; and the rainfall hyetograph.

2.2 Simulated Conditions. A series of numerical simulations were run for a wide veriety of soils with distinct hydrologic characteristics. Soils were selected that are representative of each of the four SCS hydrologic soil groups given in Table 1. The SCS classification is used extensively in many watershed models because soils can be assigned to hydrologic soil classes based on texture information from routine soil surveys. As mentioned earlier, the basic description required for each soil consisted of the soil moisture retention curve and the unsaturated hydraulic conductivity function. The representative soil properties were selected as follows:

A. A soil texture classification was chosen that was representative of each SCS soil group based on the description given in Table 1. Because so many soils fall in the SCS B and C soil groups, two soil texture classes were selected for each of these hydrologic classes. The soil textures selected for each SCS soil group are given in Table 4.

TABLE 4. REPRESENTATIVE SOIL PARAMETERS

(1) Soil Texture	(2) SCS Hydrologic Soil Group	SCS Steady State Infiltration Rate (3) f_c (cm/min)	(4) K_S (cm/min)	(5) Porosity, $\bar{\theta}_s$	Cooefficient of Power Curve (Eq. 13) (6) $\bar{\Psi}_s$ (log)	Exponent of Power Curve (Eq. 13) (7) $\bar{b}$	Moisture Content Field Capacity (8) θ_{fc}	Moisture Content Wilting Point (9) θ_{wp}
Loamy Sand	A	0.013-0.019	0.102	0.410	1.78	4.38	0.12	0.05
Sandy Loam	B	0.006-0.013	0.043	0.435	7.18	4.90	0.20	0.09
Loam	B	0.006-0.013	0.011	0.45	14.6	3.59	0.25	0.13
Sandy Clay Loam	C	0.002-0.006	0.007	0.42	8.63	7.12	0.25	0.15
Silty Clay Loam	C	0.002-0.006	0.004	0.477	14.6	7.75	0.32	0.19
Silty Clay	D	0. -0.002	0.002	0.492	17.4	10.4	0.37	0.26

(3) From Table 1.

(4) From Rawls et al., 1981.

(5) (6) (7) From Clapp and Hornberger, 1978.

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B. Because of the variability of hydrologic properties even within a soil texture class, representative moisture-retention curves derived from a statistical analysis of desorption data (Rawls et al., 1981; Clapp and Hornberger, 1978) were selected for each soil texture class. An extensive compilation of hydrologic properties was summarized and used to estimate the parameters of the Brooks and Corey equation by Rawls et al., (1981). The Brooks and Corey equation provides a reasonable estimation of the $\psi - \theta$ curve for tensions less than 50 cm (Brakensiek, et al., 1981) and is given as:

$$S_e = (\psi_b/\psi)^{\lambda} \tag{12}$$

where S_e is the effective saturation, ψ is the capillary pressure head (cm), ψ_b is the bubbling pressure (cm), and λ is the pore size distribution index. The parameters of an empirical power curve were estimated for each soil texture by Clapp and Hornberger (1978) based on desorption data by Holtan et al. (1968). The power curve was defined as:

$$\psi = \psi_s(\theta/\theta_s)^{-b} \tag{13}$$

where θ_s is the saturated water content or porosity; θ is the water content; and both ψ_s and b are empirical parameters. In addition, a parabolic modification of the power curve was proposed to account for gradual air entry near saturation. The functions resulting from Eqs. (12) and (13) were compared for specific soil texture classes and found to be very similar. The power curve, given by Eq. (13) was used with representative parameter values in

Table 4 from Clapp and Hornberger (1978). The resulting moisture-retention curves are given in Fig. 6.

C. Representative unsaturated hydraulic conductivity functions were described by a method proposed by Campbell (1974). The unsaturated hydraulic conductivity function is described based on a knowledge of the power curve describing the moisture retention function and and estimate of K_s. The equation is given as:

$$K(\psi) = K_s(\theta/\theta_s)^{2b+3} \tag{14}$$

where $K(\psi)$ is the hydraulic conductivity at a specified soil pressure, ψ; and b and θ_s are defined as in the power curve given in Eq. (13). An estimate of a representative value of K_s for each soil texture is difficult because of the large variations of this property even within a soil texture class. A set of mean values of saturated hydraulic conductivity values for each soil texture class given by Rawls et al. (1981) were used in this study.

For each soil given in Table 4, infiltration simulations were run for a range of initial moisture conditions that would be expected to occur under field conditions. Soil moisture conditions that were tested ranged from near the wilting point to near saturation. Moisture conditions at wilting point, field capacity and saturation, defined by the moisture retention curve are given in Table 4. A summary of the simulated initial conditions is given in Table 5. The vertical distributions in each initial moisture profile is assumed to be uniform. The implications of this assumption are discussed in a later section.

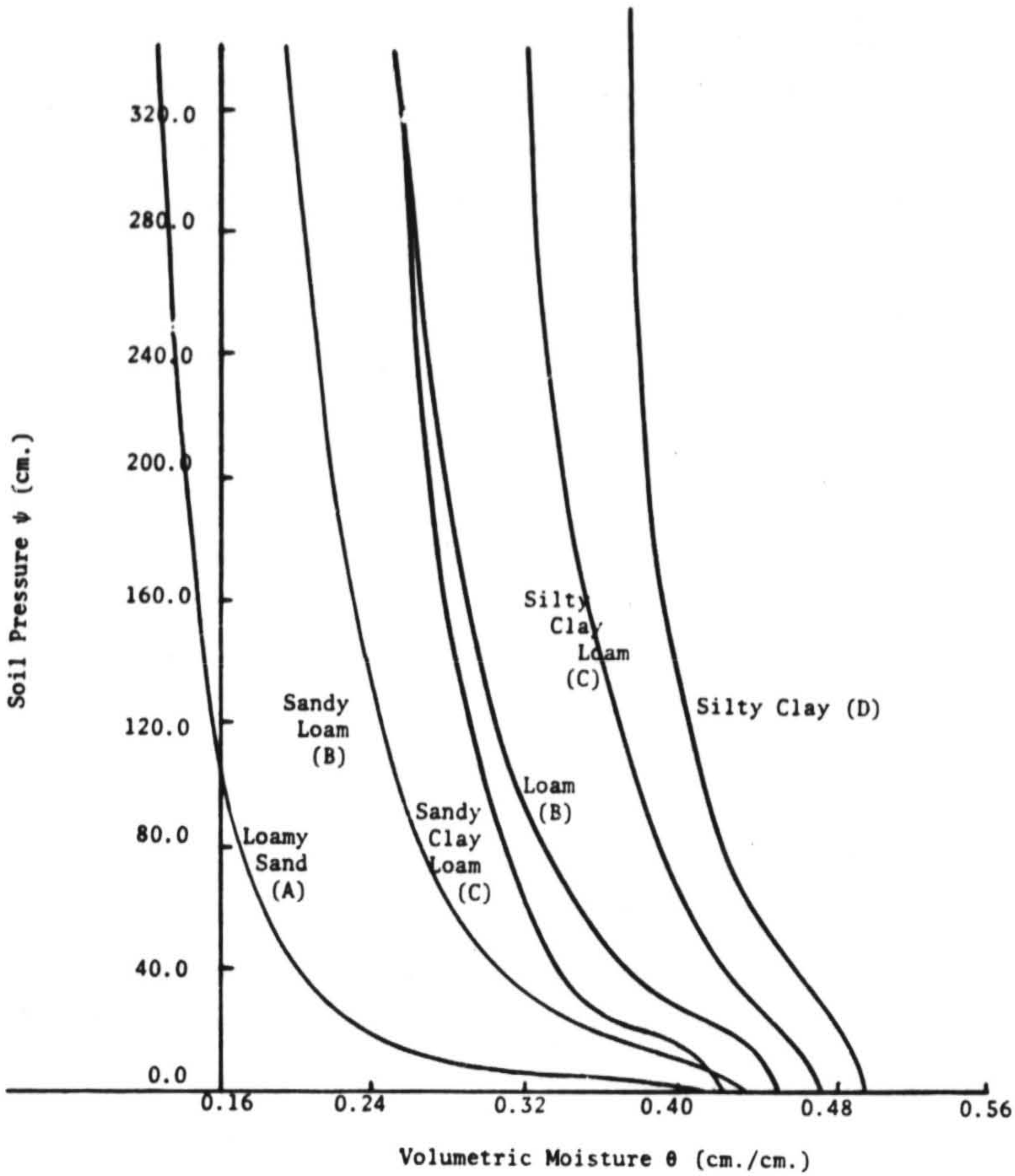

FIGURE 6. REPRESENTATIVE MOISTURE-RETENTION CURVES

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TABLE 5

INITIAL MOISTURE CONDITIONS SIMULATED
FOR EACH REPRESENTATIVE SOIL

Soil	SCS Group	Initial Volumetric Moisture	% Saturation
Loamy Sand	A	0.24	59
		0.38	93
Sandy Loam	B	0.28	64
		0.40	92
Loam	B	0.19	40
		0.22	49
		0.28	62
		0.32	71
		0.36	80
		0.40	89
Sandy Clay Loam	C	0.22	52
		0.30	71
		0.38	90
Silty Clay Loam	C	0.25	52
		0.35	69
		0.45	94
Silty Clay	D	0.34	69
		0.40	81
		0.46	93

2.3 Development of Infiltration-Storage Functions. The results of each numerical simulation defined infiltration capacity as a function of time for a particular soil, initial moisture condition and rainfall event. An example of results of these simulations is shown in Fig. 7. where the infiltration capacity of a loam soil with an uniform initial moisture content of 0.28 cm/cm is shown during three rainfall intensities.

Results such as those shown in Fig. 7 were expressed in an infiltration-storage relationship as follows:

A. An incremented time period was defined for period by period accounting of moisture movement. This time period should correspond to the time period used to describe the hyetograph in the particular simulation model in which the infiltration storage relationship would be used. A time period of 15 minutes was chosen as a representative value to be used in this study. However, infiltration-storage relationships could be developed from the simulation results for any time period of interest. The time period is denoted as Δt.

B. Two quantities were defined as follows. The depleted storage, ΔS_t, was defined as:

$$\Delta S_t = \Sigma I_t \tag{15}$$

where ΔS_t is the total volumetric change in moisture storage from initial moisture conditions at time t; and ΣI_t is the total volume of infiltration from the rainfall event at time t. The infiltration volume over a particular time period, $\Delta I_{t,\ t+\Delta t}$, is

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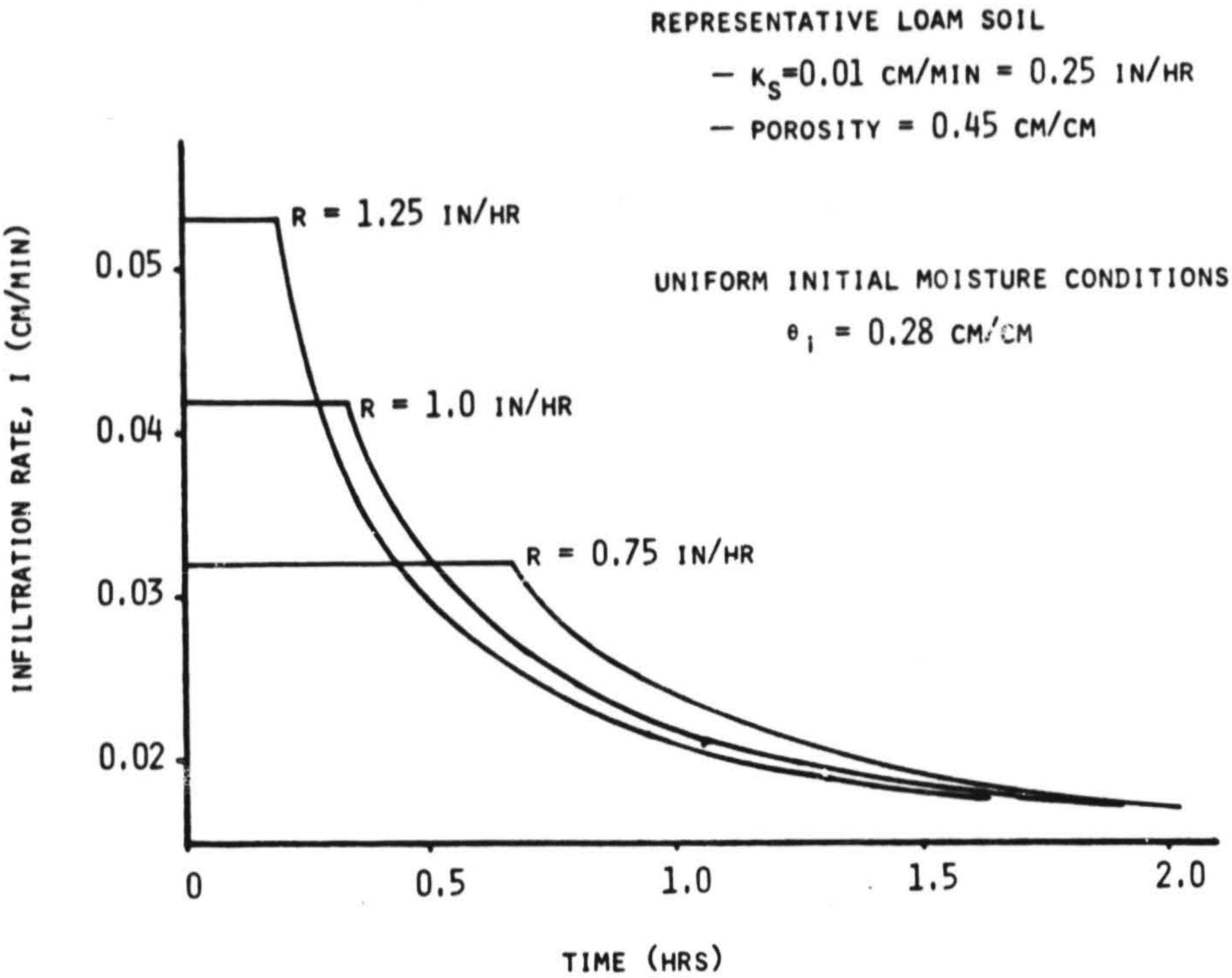

FIGURE 7. TIME VARYING INFILTRATION CAPACITY RESULTING FROM NUMERICAL SIMULATIONS

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defined as:

$$\Delta I_{t,t+\Delta t} = \Delta S_{t+\Delta t} - \Delta S_t \quad (16)$$

where ΔS_t and $\Delta S_{t+\Delta t}$ are the depleted storage at the beginning and end of the time period, respectively.

C. Using equations (15) and (16), the infiltration capacity over a time step, $\Delta I_{t,t+\Delta t}$ can be expressed as a function of the depleted storage at the beginning of the time step ΔS_t. An example of this transformation is shown in Table 6 for the results of a particular simulation shown in Fig. 7. From the numerical results, the cumulative infiltration volume at 15 minute time increments are determined. Using Equation (15), ΔS_t can be equated to the cumulative infiltration volumes as shown in Table 6, cols. (1) and (2). Using Equation (16), infiltration volumes over the time step are defined (col. 3).

The infiltration-storage relationships resulting from the rainfall simulations for a loam soil with an initial moisture conditions of 0.28 cm/cm (Fig. 7) are shown in Fig. 8. When the ΔI - ΔS relationship is used to describe the infiltration capacity, a unique function, such as shown in Fig. 8, describes the infiltration behavior of a soil with a specific initial moisture condition regardless of the rainfall rate. Of course, prior to surface saturation and ponding, the infiltration rate into the soil would be limited by the rainfall rate as shown in Fig. 8.

The fact that the ΔI - ΔS relationship for a specific soil and initial moisture condition is independent of the rainfall rate is significant because

TABLE 6

INFILTRATION-STORAGE RELATIONSHIP
DERIVED FROM RESULTS OF A NUMERICAL SIMULATION

(1) t(min)	(2) S_t(cm)	(3) I(cm)
0	0	
		.786
15.0	0.786	
		0.539
30.0	1.325	
		0.405
45.0	1.730	
		0.347
60.0	2.077	
		0.310
75.0	2.387	
		0.284
90.0	2.671	

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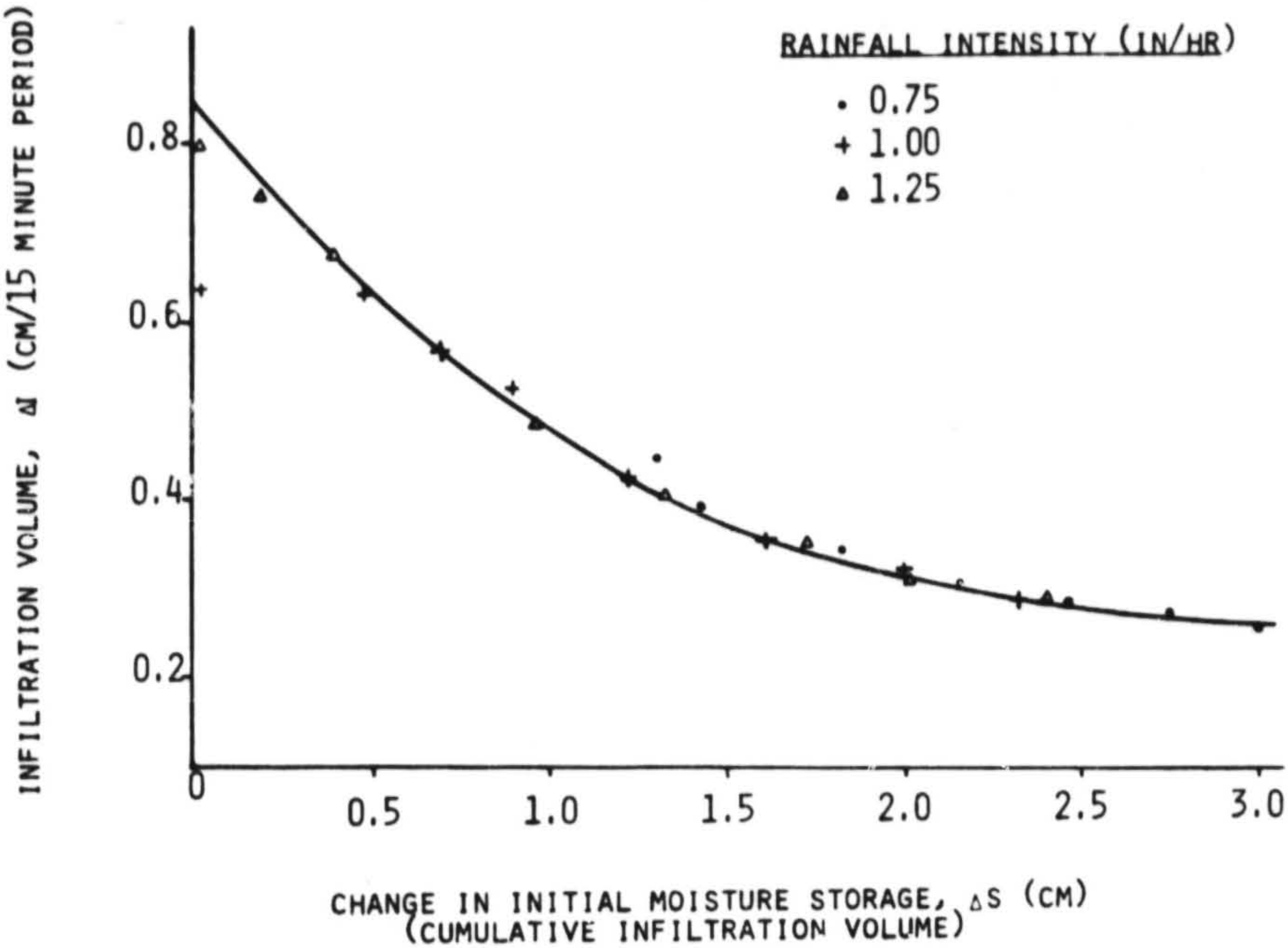


FIGURE 8 INFILTRATION-STORAGE RELATIONSHIPS FOR A LOAM SOIL (θ_i = 0.28)

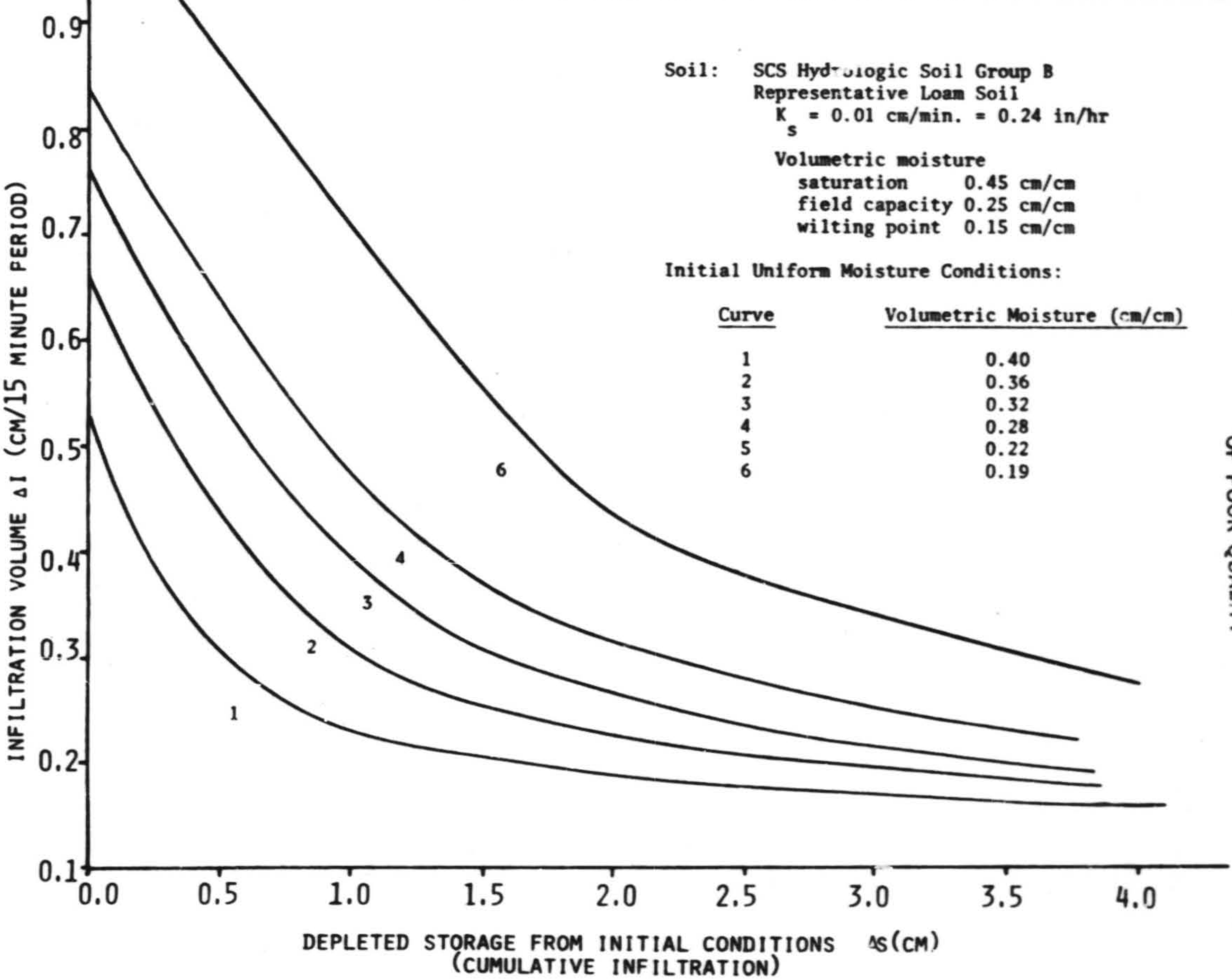
it indicates that this function, derived from numerical simulations for a specific rainfall rate, can be used to estimate infiltration capacity during any specified rainfall event. The hypothesis forms the basis for a simple procedure to be presented in the next section.

The ΔI - ΔS functions resulting from a range of initial moisture conditions in the loam soil are shown in Fig. 9. The family of curves are similar to those presented by Bloomfield et al. (1981) in Fig. 5 showing the impact of the initial moisture condition on the structure of the infiltration-storage function for a particular soil. The functions presented in this study are different in structure from those developed by Bloomfield however, because of the way that soil water storage is defined. As outlined earlier, Bloomfield used equations (4), (5) and (6) to express I as a function of used storage, US. The quantity, US, is a function of both initial moisture conditions and a specified depth of available storage and, thus, the curves are unique for the depth of storage that was assumed.

The curves shown in Fig. 9 are defined in a more general form by expressing infiltration capacity, ΔI, in terms of depleted storage which is, by definition, the cumulative infiltration volume. In this format, the curves can be applied to any specific watershed where the depth of available storage in the upper soil layer may vary.

The ΔI - ΔS functions are given in Appendix A for each of the six soils. Functions are given for time periods of 15 minutes and 5 minutes.

FIGURE 9. INFILTRATION CAPACITY AS A FUNCTION OF DEPLETED SOIL MOISTURE STORAGE



3. USE OF INFILTRATION - STORAGE FUNCTIONS

3.1 Example problem. The approach is best illustrated through the use of an example given in Table 7. The illustrated rainfall hyetograph occurs on an SCS B hydrologic soil with an initial volumetric moisture condition of 0.28 cm/cm. Therefore, curve 4 for a loam soil (Fig. 9) is used to estimate infiltration capacity throughout the event. The procedure is outlined as follows:

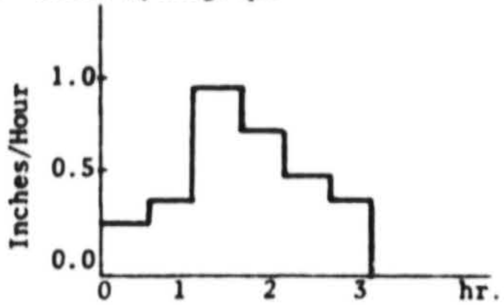
1) Columns (1) and (2) are defined by the rainfall hyetograph for the specific problem. Column (1) is the time interval by 15 minute increments. Column (2) is the volume of rainfall occurring during each 15 minute increment (for example, 0.26 in/hr x 15/60 x 2.54 cm/in = 0.165 cm).
2) ΔS, in Column (3), is the volume of available storage in the soil column that has been depleted at the beginning of each time increment. By definition, $\Delta S = 0$ at the beginning of the rainfall event because no storage has been depleted.
3) ΔI, in Column (4) represents the total volume of rainfall infiltration during the time increment. ΔI is defined as a function of ΔS(Col. (3)) by using curve 4 in Fig. 9.
4) Q in Column (5), is the rainfall excess volume over the time increment ($\Delta R - \Delta I$) and ΣQ (Col. (6) is the cumulative rainfall excess volume.
5) At the beginning of the 0.15 minute increment, no storage has been depleted. Using curve 4, the infiltration capacity for $\Delta S = 0$ exceeds 0.8 cm, so the entire volume of rainfall, 0.165 cm, infiltrates into the soil.

TABLE 7

EXAMPLE PROBLEM: USE OF ΔI - ΔS CURVES

<u>Example</u>: Estimate Infiltration for Following Conditions:

1. Storm hyetograph



2. Soil properties: SCS soil group B (assume Loam soil)

3. Initial moisture conditions: 0.28 cm/cm

4. Assume unlimited depth of storage

<u>Procedure</u>: Use representative infiltration-storage function to describe infiltration capacity for this soil and initial moisture condition.

(1) Time (min) Interval	(2) ΔR/15 min (cm)	(3) ΔS(cm)	(4) ΔI/15 min (cm)	(5) Q(cm)	(6) ΣQ(cm)	(7) ΣQ(cm) numerical
0-15	0.165	0.0	0.165	0.000	0.000	0.0
15-30	0.165	0.165	0.165	0.000	0.000	0.0
30-45	0.240	0.330	0.240	0.000	0.000	0.0
45-60	0.240	0.570	0.240	0.000	0.000	0.0
60-75	0.630	0.810	0.555	0.075	0.075	0.075
75-90	0.630	1.365	0.400	0.230	0.305	0.301
90-105	0.480	1.765	0.342	0.138	0.443	0.440
105-120	0.480	2.107	0.310	0.170	0.613	0.607
120-135	0.315	2.417	0.285	0.038	0.643	0.684
135-150	0.315	2.702	0.270	0.040	0.688	0.684
150-165	0.240	2.972	0.240	0.000	0.688	0.684
165-180	0.240	3.222	0.240	0.000	0.688	0.684
RAINFALL VOL=4.14cm=1.63 in				RAINFALL EXCESS=0.69cm		

6) At the beginning of the 15-30 minute period, 0.165 cm of storage has been depleted as a result of the infiltration during the 0-15 minute period. Again curve 4 indicates that for $\Delta S = 0.165$, the infiltration capacity exceeds the rainfall rate, and the entire rainfall volume infiltrates.

7) The above procedure is repeated for each time increment by adding the infiltration volume, ΔI, from the previous time step to the cumulative depleted storage, ΔS. The $\Delta I - \Delta S$ curve is then used to define the infiltration capacity for the current time step as a function of ΔS.

8) In this example, all precipitation enters the soil until the 60-75 minute increment. At the beginning of this increment, $\Delta S = 0.810$cm and the resulting ΔI_{max} (from curve 4) is 0.555 cm. Therefore, the infiltration rate is now limited by the soil capacity and the rainfall excess (Col.(5)) is 0.630 - 0.555 = 0.075 cm.

3.2 Comparison with Numerical Simulation. The results using this precedure were compared with the infiltration capacity resulting from a numerical solution of Eq. 7 using the specific rainfall boundary conditions of this problem. The cumulative rainfall excess volumes at the end of each time increment are given in Column (7) of Table 7. A comparison of Columns (6) and (7) indicates that the use of the $\Delta I - \Delta S$ relationship provided results in very close agreement with the numerical simulation of a specific rainfall event. The infiltration rate cruves from the numerical simulation and the infiltration-storage method are shown in Fig. 10.

TOTAL RAINFALL VOLUME = 1.630 IN.

PREDICTION OF RAINFALL EXCESS

- INFILTRATION STORAGE METHOD Q = 0.271 IN. (16.6%)
- NUMERICAL SIMULATION Q = 0.269 IN. (16.5%)

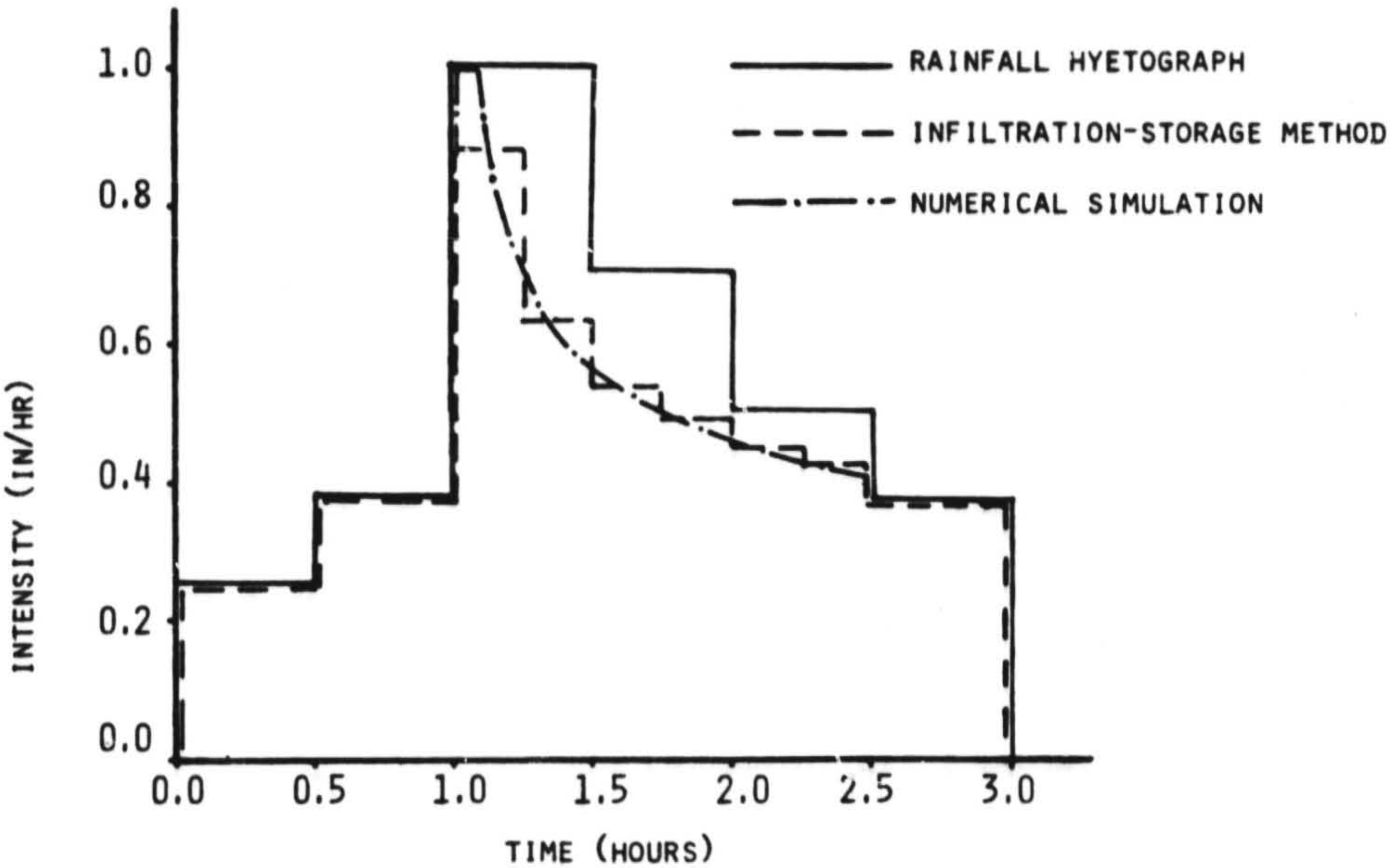


FIGURE 10. COMPARISON OF RESULTS USING INFILTRATION-STORAGE METHOD AND NUMERICAL SIMULATION

4. CONSIDERATION IN THE USE OF INFILTRATION-STORAGE CURVES

The $\Delta I - \Delta S$ curves were developed from numerical simulations for a simplified soil system. In order to apply these curves in practical watershed modeling efforts, it is necessary to consider how to adjust the curves to simulate conditions where the assumptions used in developing the curves may not be valid. Such conditions might include the following: a) some specified maximum depth of available soil storage; b) watershed moisture accounting models with varying simulation time periods; c) a non-uniform vertical distribution of initial soil moisture in the soil column; and d) variability of important soil hydraulic properties such as the saturated conductivity and porosity from the assumed representative values f a particular soil texture.

4.1 Maximum Depth of Storage. In most continuous simulation models, the infiltration capacity is related to some defined miximum available storage reflecting both the initial moisture deficit and a finite depth of soil as the estimated depth above some impeding soil layer of lower hydraulic conductivity (the A horizon in agriculture). In some models, the maximum depth is implicitly estimated in the form of an upper soil storage parameter that is defined through empirical approach.

As mentioned earlier, the $\Delta I - \Delta S$ curves were developed from numerical simulations of infiltration into soil columns of unlimited depth of available storage. Using this approach, curves developed have a more general nature instead of being structured for a specific depth of storage. However, the

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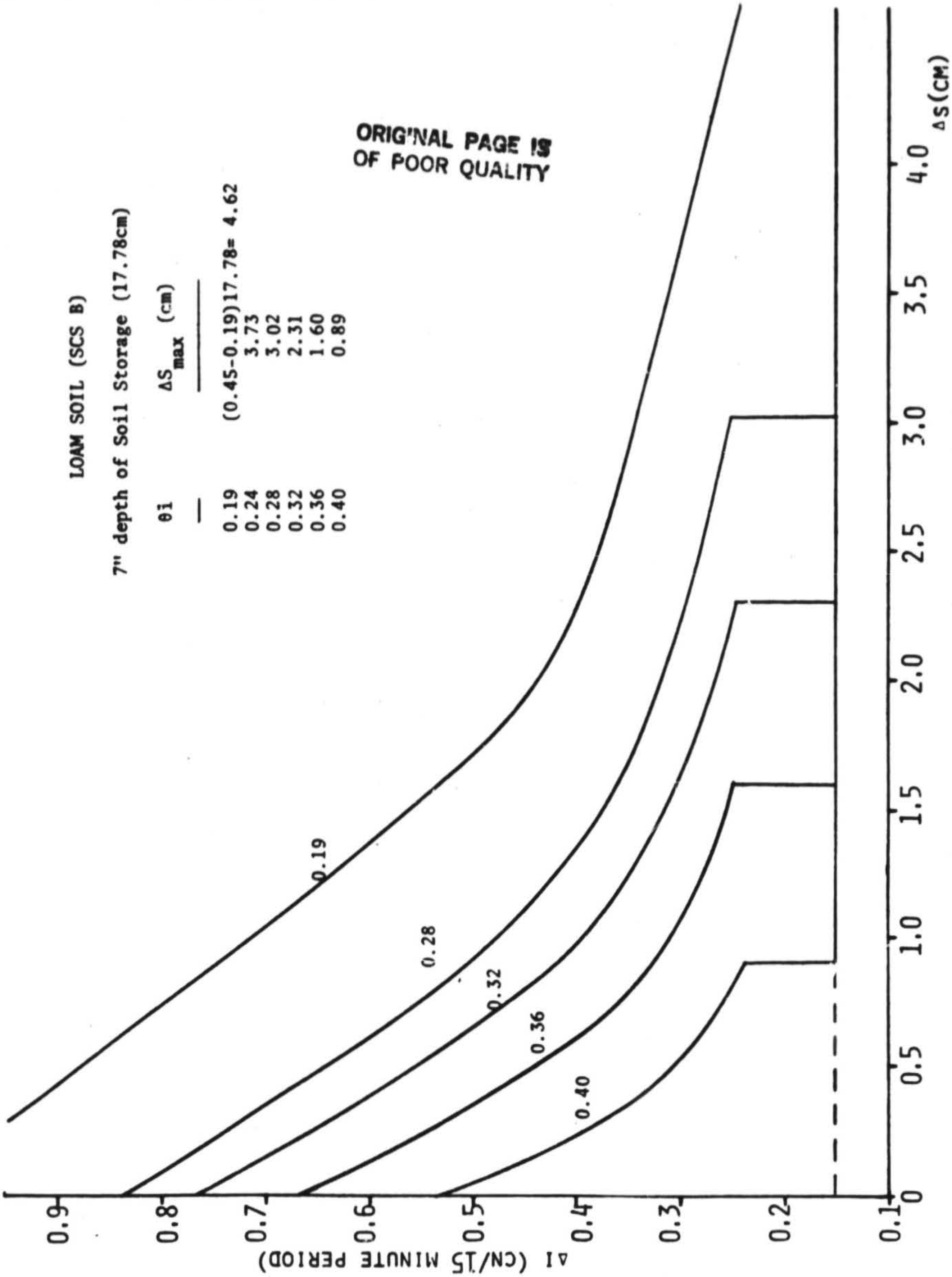
curves could be applied to a watershed condition having a specified depth of available storage by defining the maximum available storage as:

$$\Delta S_{max} = (\theta_s - \theta_i)d \qquad (17)$$

in which d is the depth of available storage to surface infiltration. The infiltration capacity of the soil is then defined by the ΔI - ΔS function until ΔS, the depleted storage, approaches ΔS_{max}. At this point, the available storage in the upper soil zone is exhausted and infiltration capacity must be defined in some other manner consistant with the structure of the overall moisture accounting model. In Holtan's equation, for instance, the infiltration capacity would be f_c, the steady state limiting infiltration rate, when surface storage becomes exhausted. In SSARR, when SMI reaches a maximum corresponding to field capacity, all rainfall becomes either surface or subsurface runoff. From a physical standpoint, once the surface soil storage is exhausted, the infiltration rate would be governed by the saturated conductivity of the surface soil layer of the impeding layer controlling drainage from the surface soil storage layer.

Figure 11 shows how the general ΔI - ΔS curves for a loam soil could be modified to represent a field condition where the available storage to a surface layer of seven inch thick soil overlaying an essentially impervious lense. When the storage of the seven inch layer of soil is exhausted, the infiltration capacity goes to zero or becomes controlled by the more impeding soil. Using this type of modification, allows for infiltration capacity to be described in terms of soil texture and the depth of available storage which considers basin geomorphology and physiography.

FIGURE 11. ΔI - ΔS CURVES MODIFIED FOR CONDITION OF DEPTH OF STORAGE = 7 IN.



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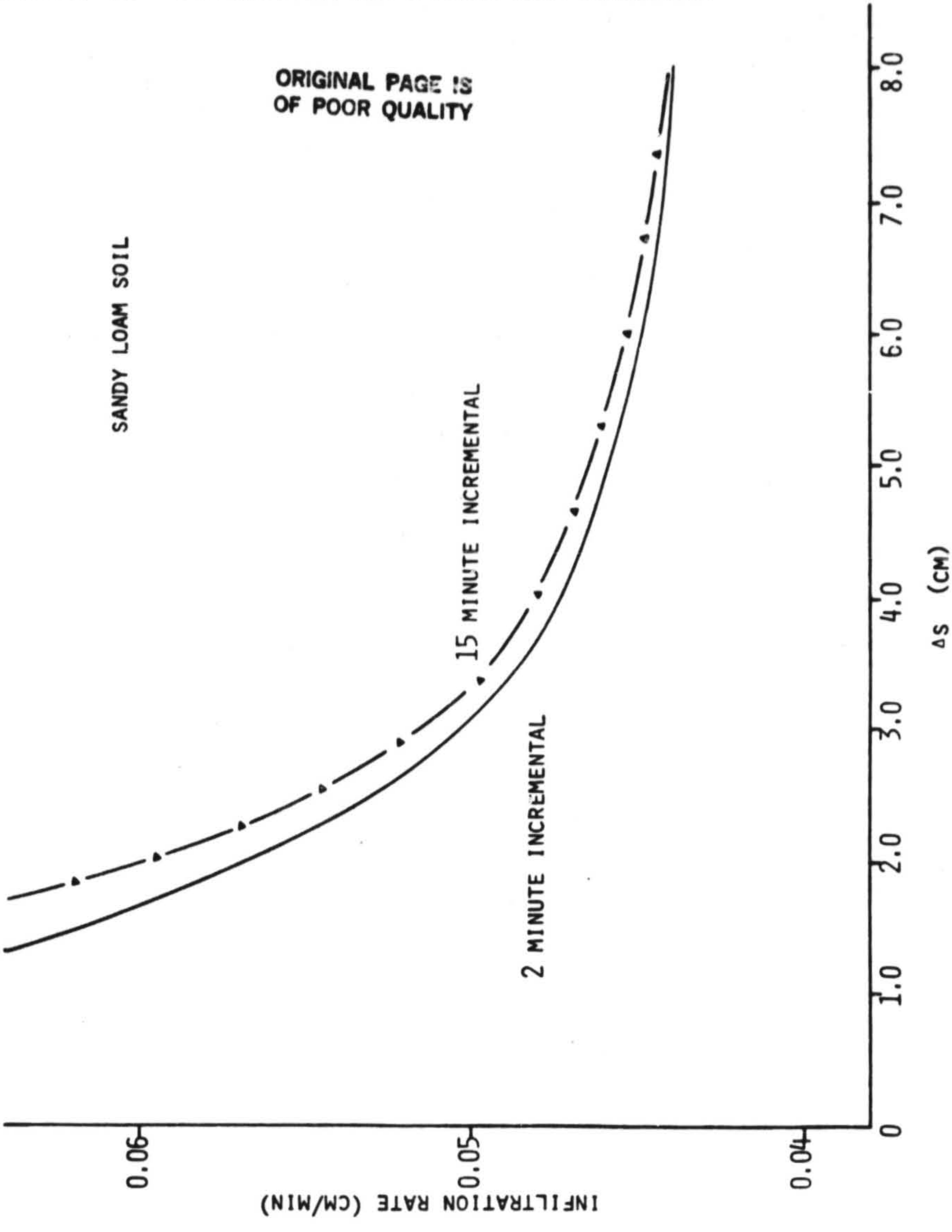
4.2 Simulation Time Period. The $\Delta I - \Delta S$ relationship is defined in terms of a specific time increment. In many instances however, it may be desirable to estimate infiltration capacities based on some other accounting period used within a particular watershed model. As Bloomfield, et al., (1981) pointed out, it would be incorrect to merely use the relationship from one accounting period to estimate infiltration volumes for another accounting period because the $\Delta I - \Delta S$ relationship is unique for the specific accounting period. The impact of the time period is illustrated in Fig. 12 where the infiltration capacity is expressed cm/hr for 15 minute and 2 minute increments.

The $\Delta I - \Delta S$ functions were expressed for two time periods of 5 and 15 minutes which were chosen to be representative of the time intervals in simulation models. In addition, time intervals of greater duration could be handled by breaking each increment into one of the two increments for which $\Delta I - \Delta S$ function was defined. For example, if the simulation model required hourly precipitation and evaporation data and performed hourly accounting of moisture movement, the rainfall hyetograph given in hourly increments could be divided into 15 minute increments and the appropriate function could be used to estimate infiltration for successive 15 minute periods throughout the rainfall event. Even if the overall simulation model uses time movements of one hour or greater, it would be advantageous to estimate infiltration capacity with more accuracy over shorter-time intervals.

4.3 Non-uniform Initial Moisture Profile. As mentioned earlier, the $\Delta I - \Delta S$ relationships were developed for initial moisture conditions vertically uniform throughout the soil regime. Without using such an assumption, an

FIGURE 12. ΔI - ΔS FUNCTIONS FOR VARYING TIME INCREMENTS



infinite number of initial conditions could have been used. Thus, the $\Delta I - \Delta S$ curves define the infiltration capacity for a particular initial moisture condition throughout the entire soil column. Moisture conditions are not necessarily uniform in the column and two questions arise. First, just how important are remotely sensed measurements of soil moisture in the top few centimeters of the soil given the uncertainty of the moisture conditions deeper in the profile which cannot easily be measured? Secondly, even if such information on the distribution of initial soil moisture conditions existed, what value would it have in a practical modeling approach?

In order to answer the above questions we must consider the sensitivity of the infiltration process to initial moisture conditions at various depths in the soil column. The problem is complex because of the variety of soil properties, initial moisture conditions and rainfall conditions. A rational examination of the theoretical infiltration process indicates the following:

1) The infiltration process is most sensitive to initial moisture conditions near the surface because: a) the time to surface ponding and initiation of runoff is controlled by the time required to saturate storage near the surface; and b) the infiltration capacity during ponded infiltration is controlled by the moisture gradient at the surface (Eq. 10).
2) The impact of the initial moisture conditions on the infiltration process decreases with depth because the impact on surface ponding and moisture gradient decreases.
3) Obviously, the deeper the wetting front moves into the soil, the greater the impact of initial moisture conditions at some depth below the surface.

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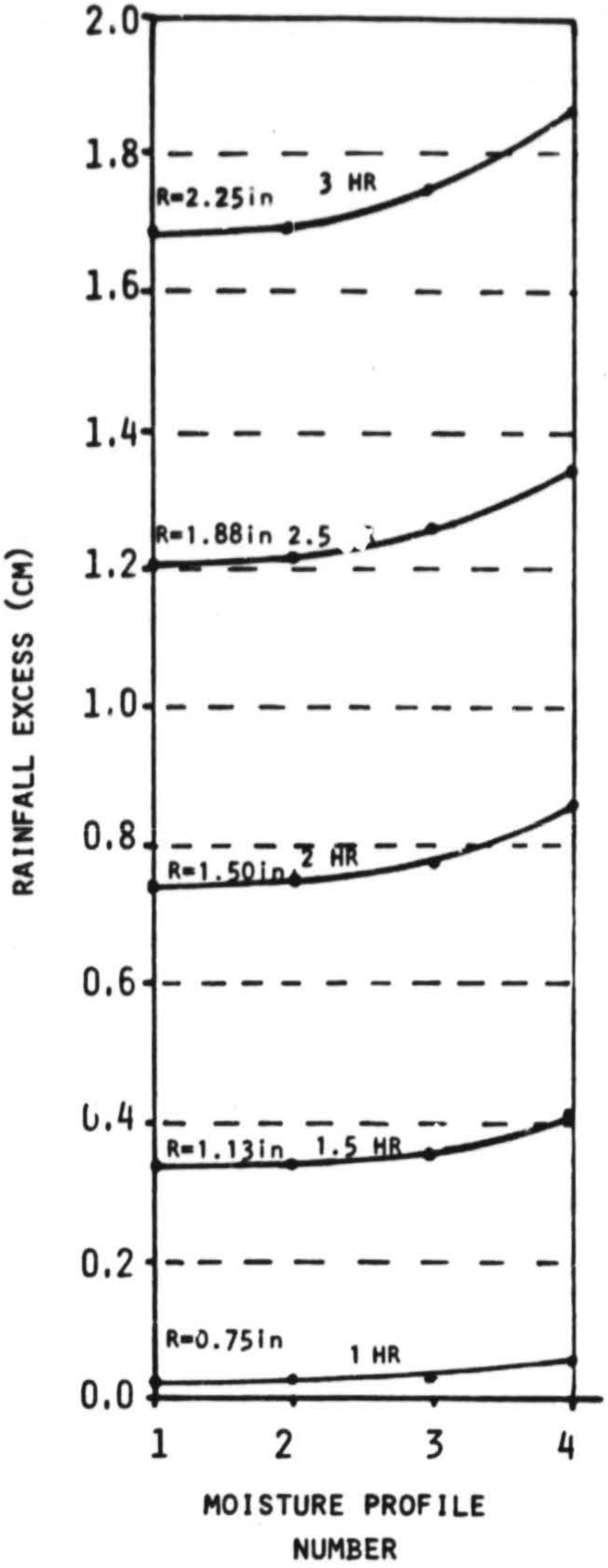
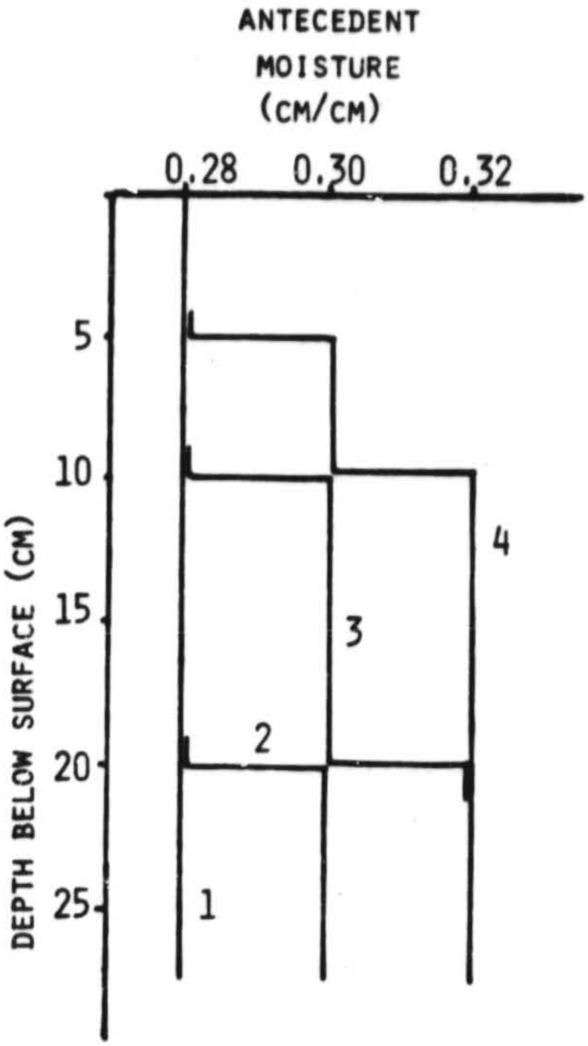
The above observations were examined for a particular soil using numerical simulations. Four initial moisture profiles all with a surface soil moisture of 0.28 cm/cm are shown in Fig. 13 a: profile 1 represents the case of a uniform initial moisture condition; profile 4 represents a more rapid variation of initial moisture condition near the surface; and profiles 2 and 3 represent intermediate cases. Infiltration capacity was simulated for each initial moisture profile from a rainfall of 0.75 in/hr. The resulting rainfall excess volumes for each initial moisture profile are shown in Fig. 13 b at various stages of the rainfall. The results indicate there is no significant difference between rainfall excess volumes resulting from profiles 1 and 2 for any of the rainfall durations. The differences resulting from profile 1 and both profiles 3 and 4 were very small for the short duration events but increased as the rainfall duration increased and the wetting front moved further into the soil. This is expected because the initial moisture content will only affect the infiltration process near the surface to a depth which has been reached by the wetting front. The longer the infiltration event progresses, the greater the depth of initial moisture will impact the process. Still, the rainfall excess produced by profile 4 for the three hour event was less than 10% greater than that estimate with profile 1.

This example represents one specific case and in no way should be used to evaluate the impact of initial moisture at various depths for the general case of all soils and initial conditions. The example discussed for a loam soil, representative of an SCS B soil, indicates that surface measurements are important even though there is uncertainty concerning moisture



13 b



13 a

FIGURE 13. RAINFALL EXCESS VOLUMES RESULTING FROM NON UNIFORM INITIAL MOISTURE PROFILES WITH SURFACE MOISTURE 0.28 CM/CM

conditions at greater depths. Obviously, additional information on conditions throughout the profile could improve infiltration estimates but these improvements may not be significant given other uncertainties in natural watersheds.

An examination of the physical principles driving the infiltration process indicates that for a variety of soil textures, a knowledge of the initial moisture conditions below a near surface region of the top 2-5 cm may not be necessary for a good estimate of infiltration and resulting runoff. In soils of high permeability, such as sand and sandy loam (SCS A or B soils), the moisture front may advance into the soil rapidly and be affected by initial moisture conditions at greater depths in the soil. However, in these soils, the hydraulic conductivity is large and dominates the process resulting in a small runoff volume regardless of the initial moisture conditions. In soils with lower permeability, clay and silt loam, for example (SCS C and D soild), the moisture front moves very slowly into the soil because of the low hydraulic conductivity. Because the moisture front advances slowly, initial moisture conditions at greater depths do not affect the process. In these soils, the runoff potential is high and greatly impacted by the initial moisture conditions near the surface.

In some cases, data may be available describing the distribution of soil moisture with depth and the question arises as to the value of such data in practical modeling. Such information as vertical distribution may be available from: 1) more sophisticated remote sensing techniques with longer wavelengths; 2) extensive surface measurements to greater depths using theoretical models driven by soil texture properites and climatic data; and 3) extensive field measurements.

The additional data could be incorporated into the infiltration-storage approach if desired. The $\Delta I - \Delta S$ relationship resulting from profile 4, (Fig. 13) is shown in Fig. 14. In the early stages of infiltration, the $\Delta I - \Delta S$ curve follows the curve corresponding to $\theta_i = 0.28$ and as ΔS increases the curve approaches the curve corresponding to $\theta_i = 0.32$. We might attempt to represent this unique curve as a function of the curves representing the two uniform moisture conditions. One simple approach is outlined in Fig. 15. Starting with an assumed initial moisture condition (Fig. 15 a), the profile is represented as a series of superimposed rectangular volumes as shown in Fig. 15 b. The maximum storage is calculated for each volume as the moisture deficit $(\theta_S - \theta_i)$ x the depth of the volume element. The $\Delta I - \Delta S$ curve corresponding to the initial moisture content of the first volume is used to calculate infiltration capacity until the maximum storage has exceeded the first volume. The $\Delta I - \Delta S$ curve for the initial moisture condition of the next volume is then used to estimate capacity until the maximum volume of storage has been reached. The $\Delta I - \Delta S$ curve is first entered at a ΔS corresponding to the volume already depleted by the preceding volume element (the area of overlap). The process could continue for as many volume elements as required. The resulting approximation of the $\Delta I - \Delta S$ function would be similar to that shown in Fig. 15 c. The method assumes that one volume element is completely filled before moisture

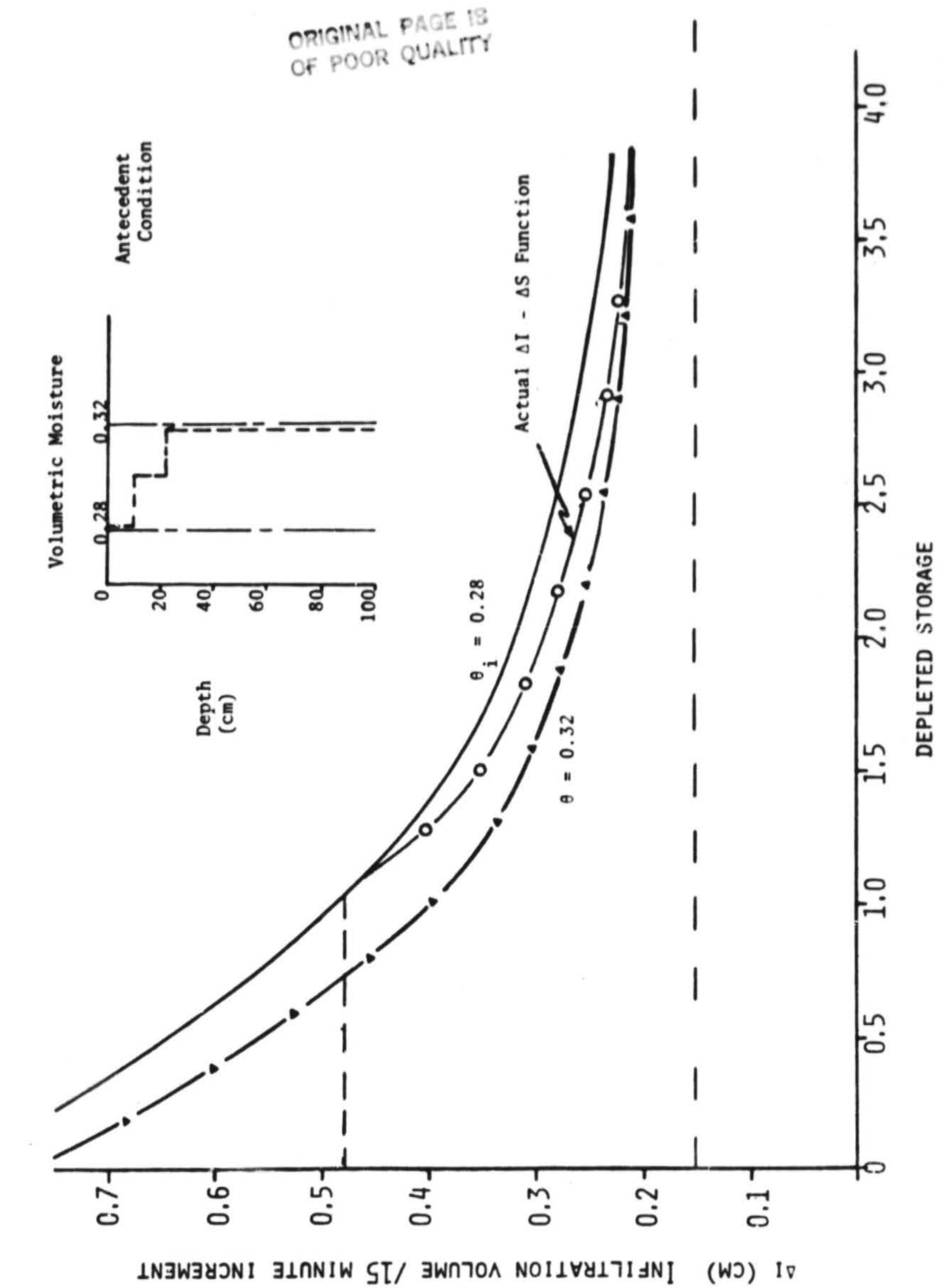
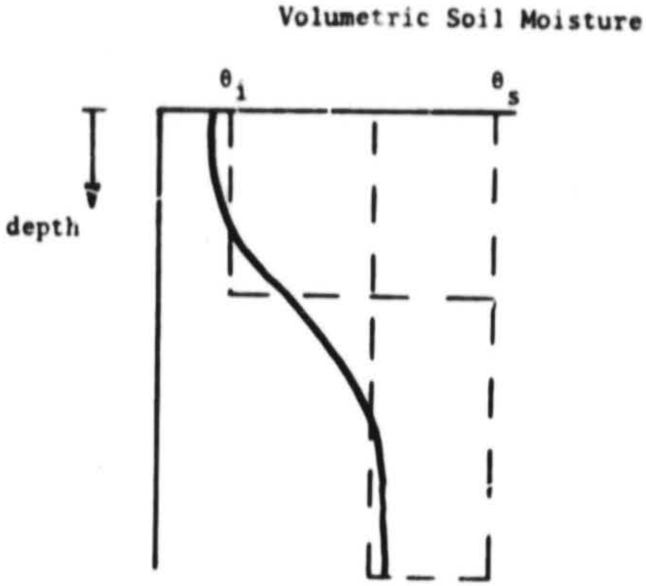
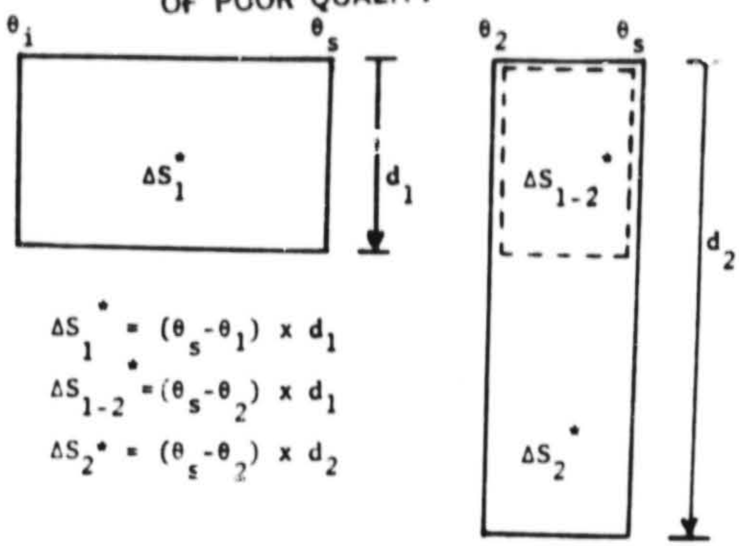


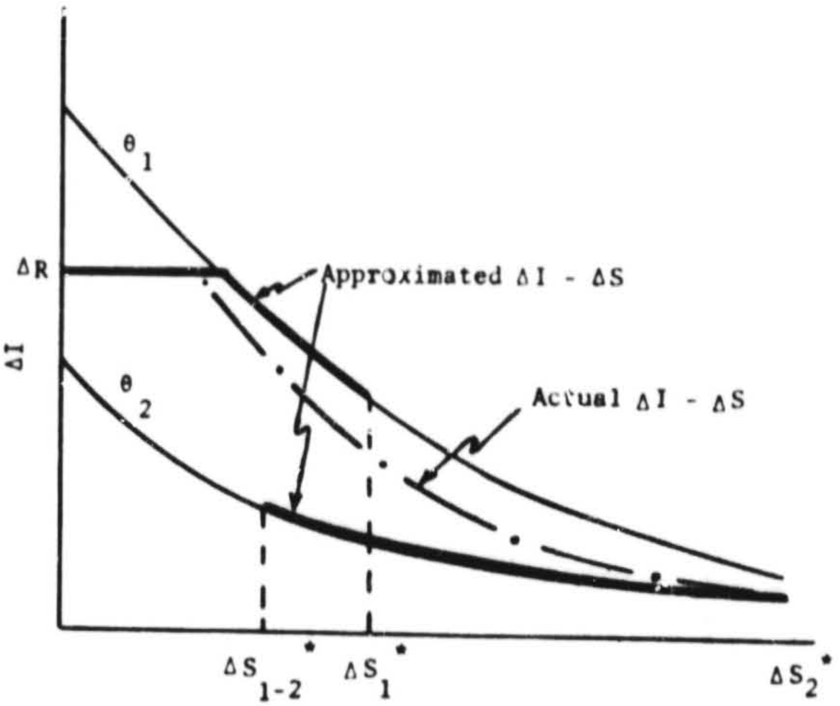
FIGURE 14. ΔI - ΔS CURVE RESULTING FROM NON-UNIFORM PROFILE



(15a) Assumed initial moisture distribution



(15b) Simplified Representatio of Storage Volume



(15c) ΔI - ΔS Function Approximated for Non-Uniform Profile

FIGURE 15. USE OF ΔI - ΔS CURVES TO REPRESENT NON-UNIFORM INITIAL CONDITIONS

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moves into the next element. This approximation would be most reasonable in soils in which the moisture wetting front is very steep, such as sand and loam, as opposed to silt loam or clay (see moisture-retention curves in Fig. 6). In addition, this approximation would be most valid in a dry initial condition where the wetting front tends to be steeper than in a wet initial condition.

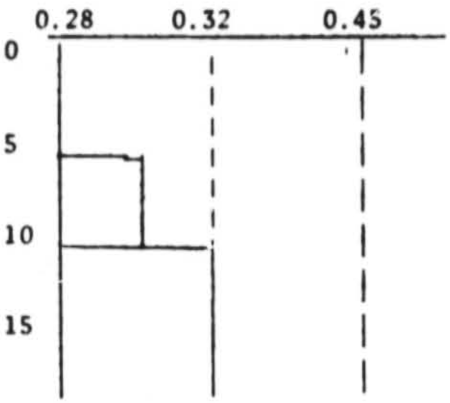
An example of the above procedure is shown in Fig. 16 for a particular non-uniform initial moisture profile. The runoff volume resulting from a 3 hour rainfall of 0.75 in/hr (2.25 in. total rainfall) was calculated as 0.77 inches. This runoff volume corresponds closely to the volume of 0.76 inches resulting from a numerical simulation of the actual initial profile. The runoff volumes resulting from only using the ΔI - ΔS curves for uniform initial moisture profiles of 0.28 and 0.32 cm/cm were 0.68 inches (11.6% error) and 0.83 inches (7.8% error), respectively. Although the example indicates that the curve adjustments will work, the differences caused by the profiles investigated were relatively small. Because of the numerous other uncertainties in a natural watershed, it is doubtful that the limited gains accomplished by the more complex adjustments are worth the effort.

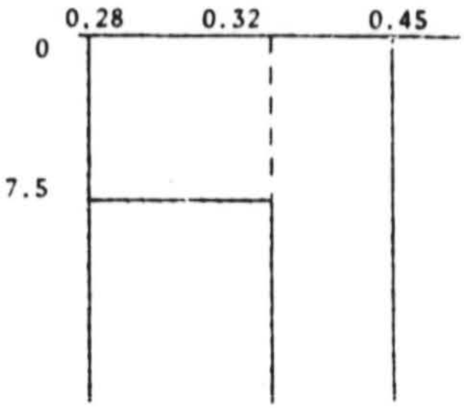
4.4 <u>Variation of Soil Hydraulic Properties</u>. The use of the I - S curves assumes that the particular hydraulic properties of the soil (K_S, θ_S, $K - \psi$, $\theta - \psi$) are similar to the representative values used in this study to derive the curves. Unfortunately, field studies and laboratory tests have indicated that important hydraulic properties such as K_S can vary significantly within soils of the same texture class. Field data presented by Nielson et al., (1973) indicated that the coefficient of variation (CV) in I_S could be 50% greater within fields of generally homogeneous soil texture. The effect of variation in K_S on the ΔI - ΔS relationship for loam and silty clay loam soils is shown

Problem:

Estimate infiltration volume

- SCS B soil
- non-uniform initial condition
- rainfall





$\Delta S_1 = (0.45-0.28)\ 7.5 - 1.28$

$\Delta S_{1-2} + (0.32-0.28)\ 7.5 = 0.30$

$\Delta S_2 =$

Time	ΔR/15 min. (cm)	ΔS (cm)	ΔI_{max} (cm)	Q (cm)	ΣQ (cm)	ΣQ (numerical) (cm)
0-15	0.476	0.0	0.476	0.0	0.0	0.0
15-30	0.476	0.476	0.476	0.0	0.0	0.0
30-45	0.476	0.952	0.476	0.0	0.0	0.0
45-60	0.476	(1.428-0.30)1.128	0.376	0.111	0.111	0.130
60-75	0.476	1.493	0.310	0.166	0.277	0.295
75-90	0.476	1.803	0.280	0.196	0.473	0.489
90-105	0.476	2.083	0.260	0.216	0.689	0.703
105-120	0.476	2.343	0.245	0.231	0.920	0.933
120-135	0.476	2.588	0.230	0.246	1.166	1.175
135-150	0.476	2.818	0.220	0.256	1.422	1.425
150-165	0.476	3.038	0.217	0.259	1.681	1.683
165-180	0.476	3.255	0.210	0.266	1.947	1.931
					ΣQ = 0.77 in.	ΣQ = 0.76 in.

Fig. 16 Example - Use of ΔI -ΔS curves for non-uniform initial moisture profile.

in Figs. 17 and 18, respectively. For each soil, K_r was varied from the representative value by $\pm$ one standard deviation assuming a normal distribution with CV of 50%. Thus, in using the representative ΔI - ΔS curves to represent a particular soil within a watershed, we might expect significant errors in the estimation of infiltration capacity due to the uncertainty of these hydraulic properties.

The variation of soil hydraulic properties within a soil texture class must always be recognized when attempting to predict runoff potential of an ungaged watershed based only on SCS soil groups or soil texture classes. However, in many instances the representative ΔI - ΔS curves may prove useful despite the problem of variability of soil hydraulic properties within soil made for large ungaged watersheds where soils are categorized according to SCS soil groups and surface moisture conditions monitored by remote sensing. When a watershed is gaged with historical rainfall-runoff data, or when more detailed information on soil hydraulic properties are available, a different method may be more feasible.

A possible approach to incorporating the variability of hydraulic parameters would be to generate a family of ΔI - ΔS curves using a Monte Carlo simulation. Using this approach, the ΔI - ΔS relation for a particular soil texture having specific initial moisture conditions would be expressed as a probability function rather than one specific curve. Ideally, such a probability function would include the effects of variation of hydraulic properties within the soil texture class. The number of simulations required to develop such a probability function for each soil texture and

FIGURE 17. EFFECT OF VARIATION ON K_S ON ΔI - ΔS CURVES FOR LOAM SOIL

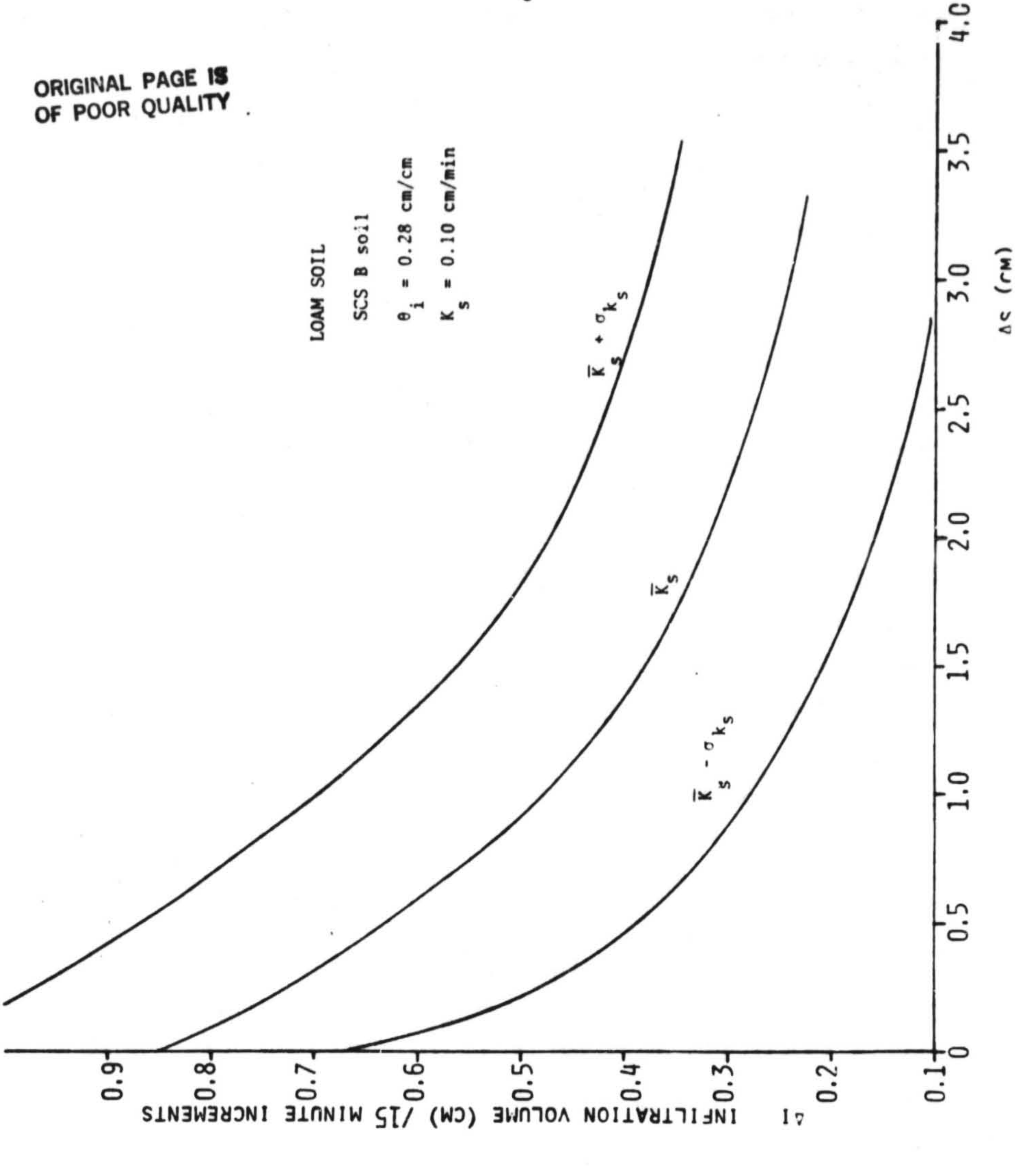


FIGURE 18. EFFECT OF VARIATION IN K_S ON $\Delta I - \Delta S$ CURVE FOR SILTY CLAY LOAM SOIL

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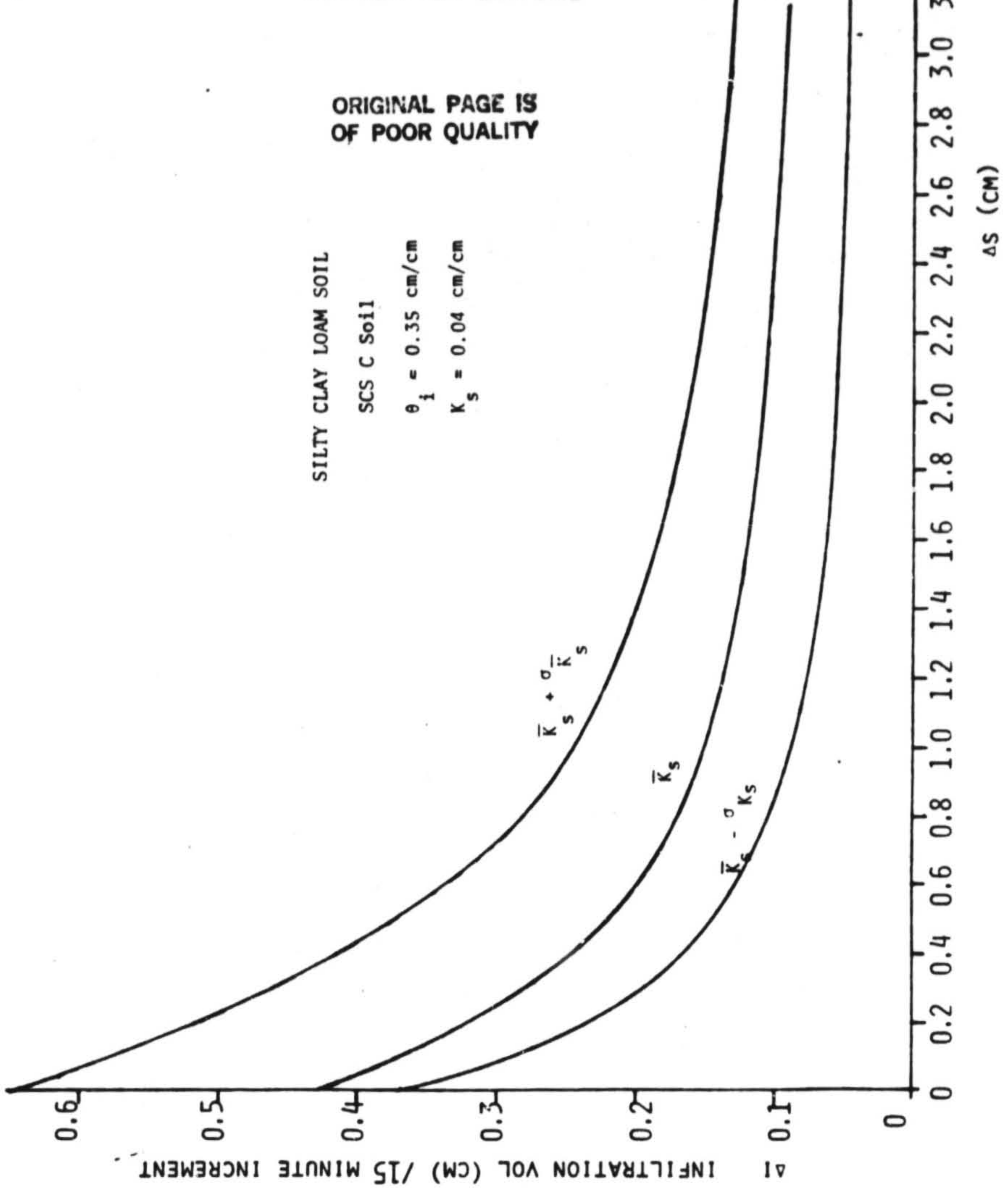

initial moisture condition would be extremely large. In addition, it would be very difficult to represent the cumulative effects of variation of many different hydraulic properties.

A more practical approach to including the effect of varying soil properties within each texture class might be to use the ΔI - ΔS curves in a form similar to that found in the Stanford Watershed Model (Crawford and Linsley, 1966). In SWM IV, areal variations of infiltration capacities are defined by plotting a cumulative frequency distribution of infiltration capacity as shown in Fig. 19. In theory, this curve would result if a large number of simultaneous infiltrometer measurements were made and plotted to show the percentage of watershed area with an infiltration capacity equal to or less than the measured values. Within a field of homogeneous soil textures, the variability could be largely attributed to variations in soil hydraulic properties. If one wished to consider the variability of K_S occurring within a texture class when estimating the infiltration capacity using the ΔI - ΔS functions, one approach could involve the following steps: (shown graphically in Fig. 20)

1) From field data, estimate a theoretical distribution of K_S values within a particular soil texture class (Fig. 20 a).
2) In addition to the ΔI - ΔS function already generated using the representational K_S value, generate at least two more ΔI - ΔS functions with K_S varied from the representative value (Fig. 20 b).
3) For a range of ΔS values, ΔI is plotted for each K_S value as a function of the fraction of the watershed unit assumed to have that particular value of K_S (Fig. 20 c).

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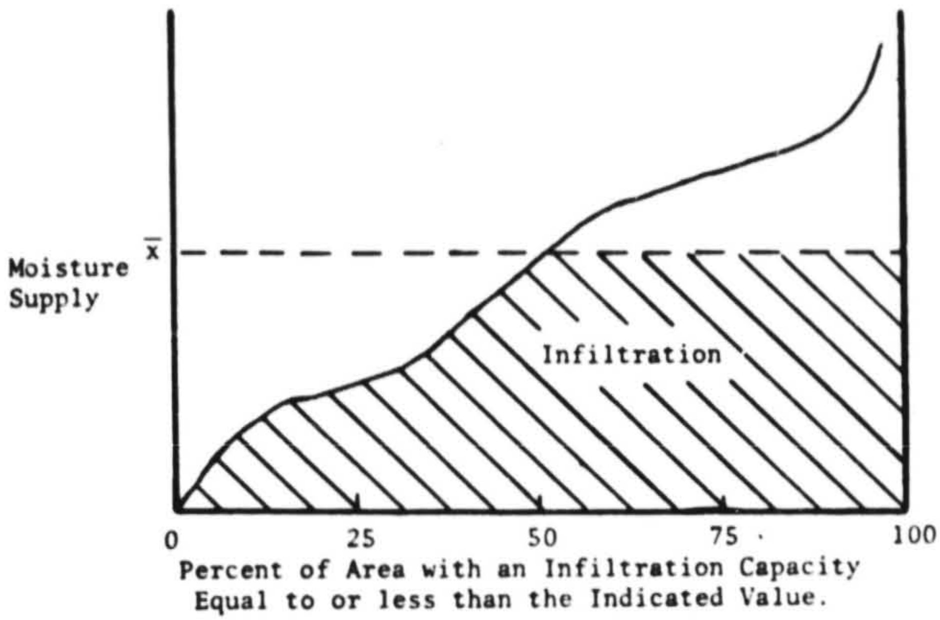

(from Crawford and Linsley, 1966)

FIGURE 19. INFILTRATION APPROACH IN STANFORD WATERSHED MODEL

(a) Assume distribution of K_s within soil texture.

(b) Generate ΔI - ΔS curves for variation of K_s.

(c) For a specific ΔS_t, express variation of ΔI over watershed unit (following approach of SMM IV).

(d) Using ΔI relationship at ΔS_t and ΔR at time t:

$$\Delta I = A_1$$
$$Q = A_2$$
$$\Delta S_{t+1} = \Delta S_t + \Delta I$$

(e) Repeat Steps (c) and (d) for ΔS_{t+1}

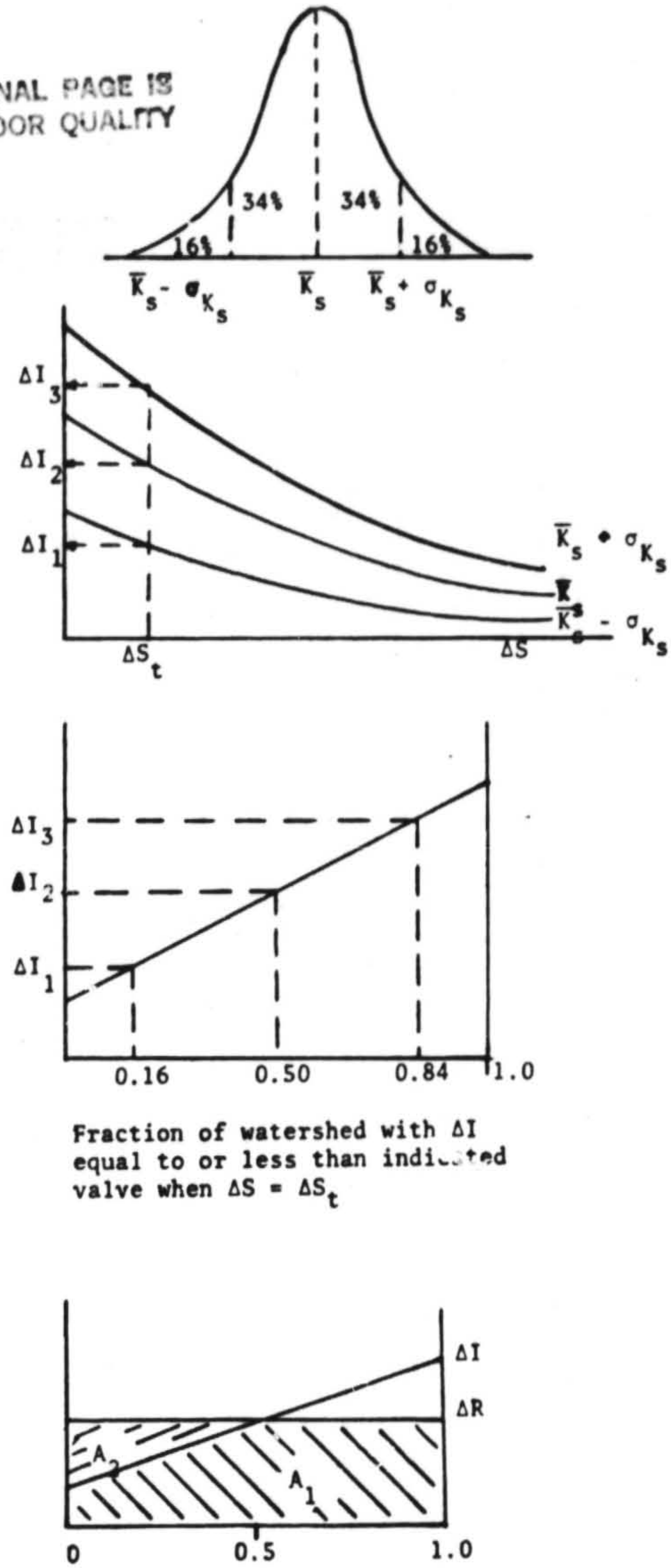

FIGURE 20. CONSIDERATION OF SPATIAL VARIABILITY OF K_S WITH UNIFORM SOIL TEXTURE

4) The rainfall moisture supply ΔR is partitioned into infiltration volume (the area under both the ΔR and ΔI curve) or runoff volume (the remaining area under the ΔR curve) (Fig. 20 d).

CHAPTER 4

APPLICATION OF INFILTRATION-STORAGE METHOD IN WATERSHED

The infiltration-storage method presented in this study could be used to improve estimates of the soil infiltration capacities which are crucial for the accounting of moisture movement throughout a comprehensive watershed model. A generalized flowchart of hydrologic processes occurring within a watershed, in Fig. 21, shows how the infiltration component affects the flow of moisture through the system. Regardless of the particular model structure, the division of precipitation into surface and subsurface processed is an important element in estimating surface runoff and updating soil moisture conditions that will affect future events.

The infiltration-storage method may be useful within a new generation of watershed moisture accounting models structured specifically to maximize the benefits of remotely sensed data. This type of model is one ultimate goal of current research efforts at RSSL and might be configured as shown in Fig. 22. The key components of the model would include:

1) A geo-referenced grid cell data base containing spatially distributed soil texture classifications obtained from soil survey maps; land-cover information from Landsat and aerial photography; and other relevant spatial data. Moisture accounting could occur at a spatial resolution equal to that of the data base.

2) A continuous moisture profile accounting model for some defined upper soil store zone affected by surface processes. The model component would predict vertical soil moisture distributions subject to ET and

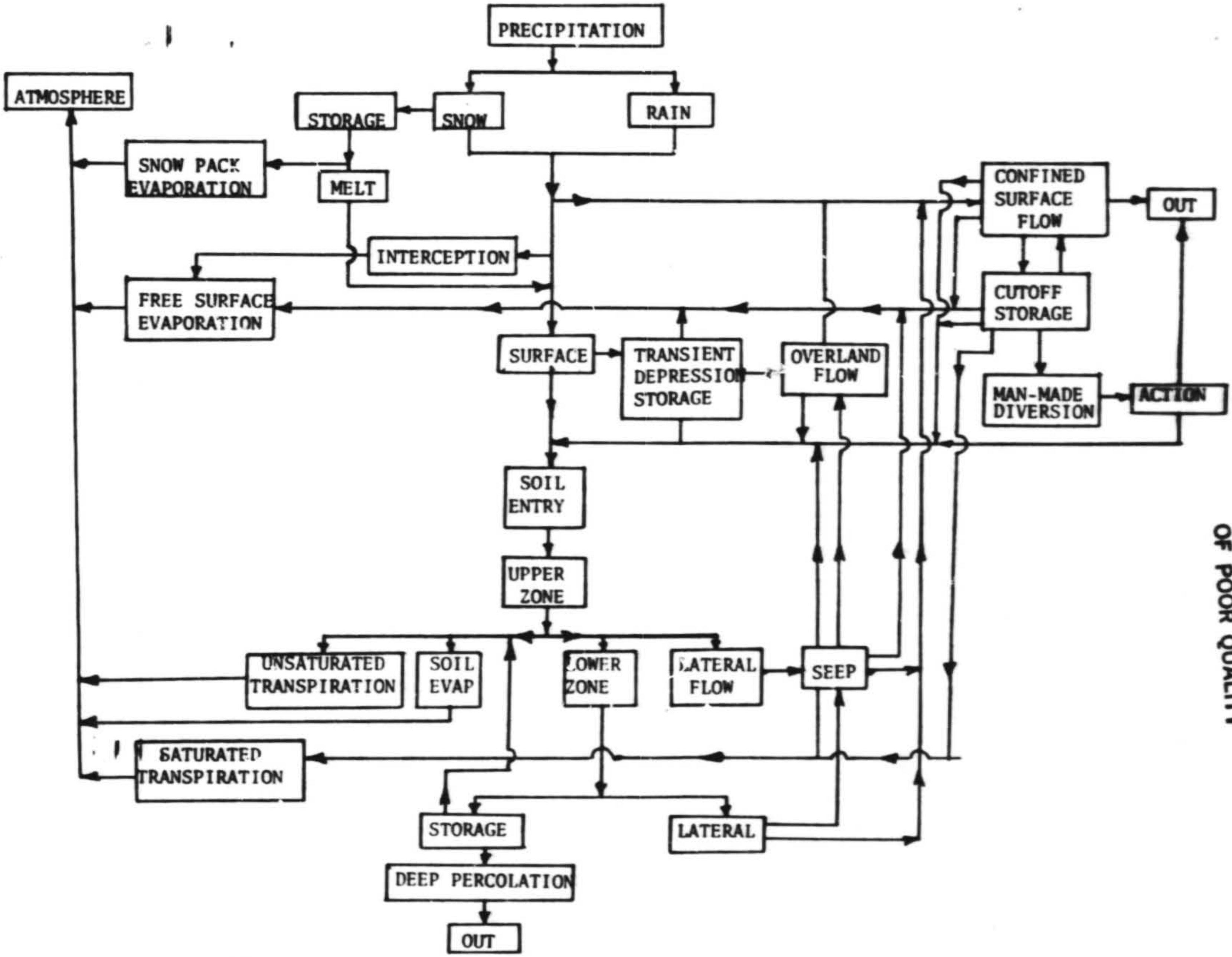

FIGURE 21. FLOWCHART OF WATERSHED HYDROLOGIC PROCESSES

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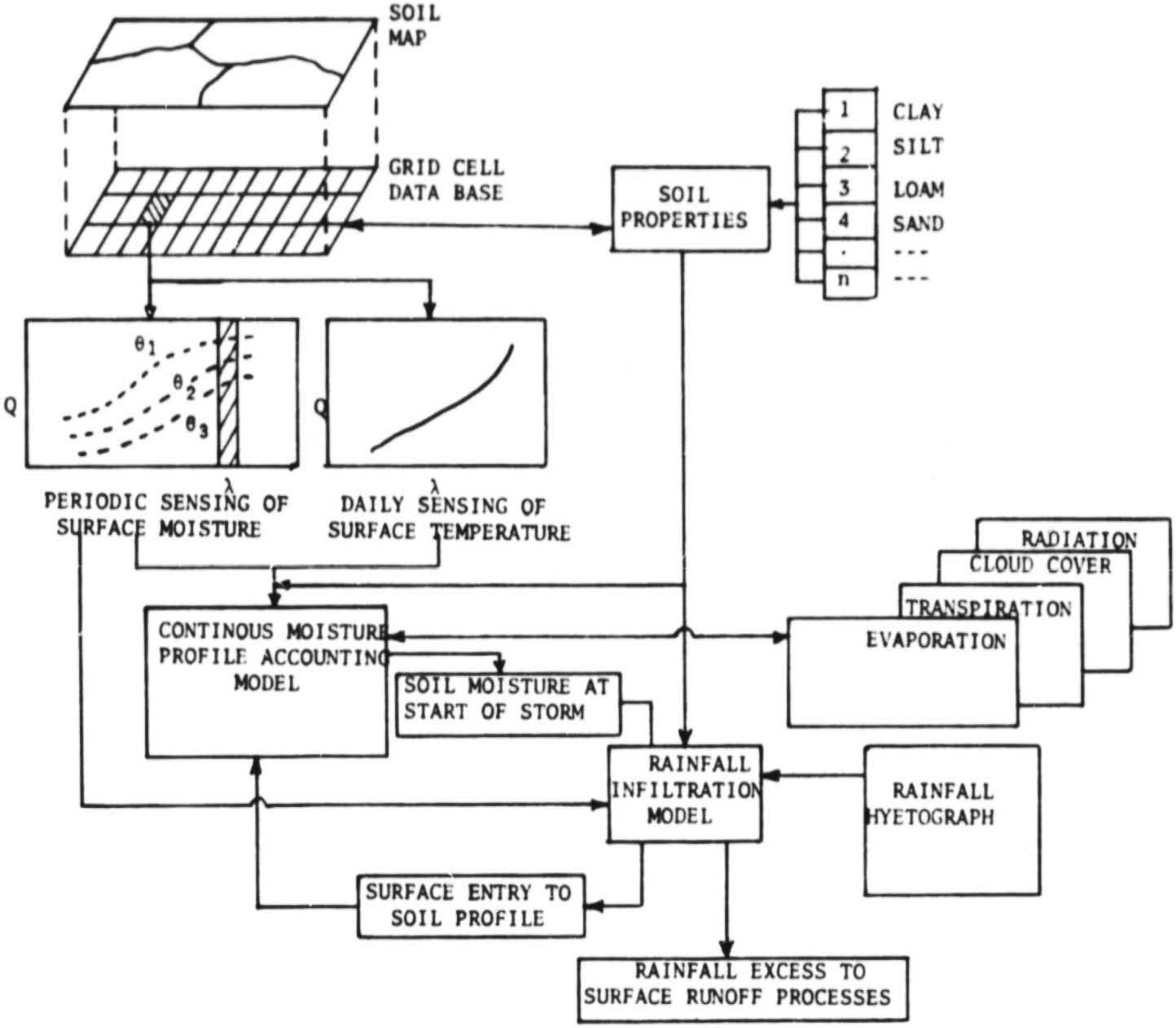


FIGURE 22. REMOTE SENSING-MODELING LINKAGES FOR CONTINUOUS MOISTURE ACCOUNTING AND INFILTRATION ESTIMATES

deep percolation and would be driven by periodic remotely sensed surface soil temperatures and moisture measurements and other available climatic data.

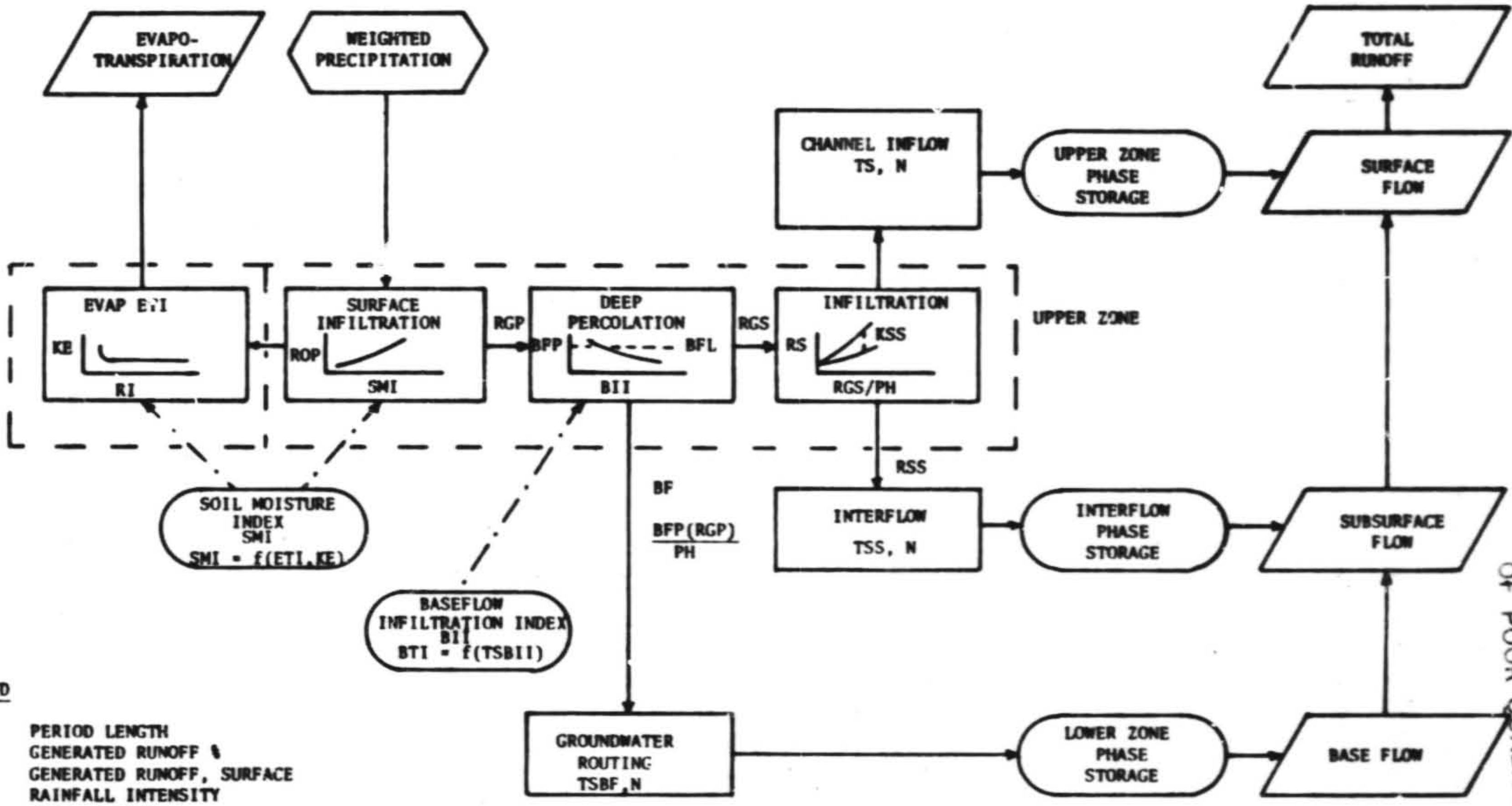
3) The rainfall infiltration component which would provide the crucial partitioning of rainfall into surface runoff and soil moisture replenishment. The model component would predict time varying infiltration capacity given a time varying rainfall hyetograph, a soil texture classification, and initial soil moisture conditions. Initial soil moisture conditions would be represented by either an initial soil moisture profile accounting model in which the initial surface moisture is updated by periodic remotely sensed measurements.

The ΔI - ΔS curves could be used to provide the infiltration component. The approach is theoretically sound yet computationally simple enough to calculate infiltration rates for many spatial units representing various soil texture and initial moisture conditions in the watershed.

It is important to consider how existing hydrologic models could be modified to gain the full benefit of remote sensing technology. These possible modifications are important because a significant effort has already been invested in the development of accepted models and the missions of many organizations revolve around their use. In addition, the acceptance and justification of a new data source such as that offered by remote sensing will most likely come as the utility of such data is demonstrated in current modeling requirements. Incorporating the ΔI - ΔS curves into current models would provide a simple, yet theoretically sound infiltration model that will logically incorporate the impact of antecedent soil moisture conditions on watershed response. The operation of this type of watershed model would interface perfectly with the anticipated capabilities of remote sensing measurements of surface soil moisture.

The effort involved in implementing the approach within existing models would vary considerably depending on the particular model structure. For example, within USDAHL and other simulations models using Holtan's infiltration equation, it would be relatively easy to substitute the representative $\Delta I - \Delta S$ curves in place of the infiltration function described by Holtan's equation. The reason is that both Holtan's equation and the representative $\Delta I - \Delta S$ curves relate infiltration capacity to the same physical quantity, the volumetric soil moisture storage in a defined upper soil zone. The $\Delta I - \Delta S$ approach could be implemented using the FORTRAN subroutine given in Appendix B, which calculates the infiltration capacity using tabular values of the representative $\Delta I - \Delta S$ curves given in Appendix A. Another approach would be to fit the function described by Holtan's equation to the representative $\Delta I - \Delta S$ curves. The parameters of Holtan's equation would then be a function of soil properties and initial moisture conditions. Bloomfield et al., (1981) demonstrated this process by fitting two other infiltration storage equations to theoretical infiltration-capacity curves for a specific soil.

The use of the representative $\Delta I - \Delta S$ curves within the SSARR model would be more difficult because of the model structure. The moisture accounting position of the model, shown within the box in Fig. 23, consists of three tabular functions which are used to separate infiltration and subsequent percolation and interflow from immediate surface runoff. The runoff percent



LEGEND

PH	PERIOD LENGTH
RGP	GENERATED RUNOFF %
RGS	GENERATED RUNOFF, SURFACE
RI	RAINFALL INTENSITY
RS	SURFACE RUNOFF
WP	WEIGHTED PRECIPITATION

Figure 23. SSARR Model Schematic Diagram (Adapted from Peck, et al., 1981)

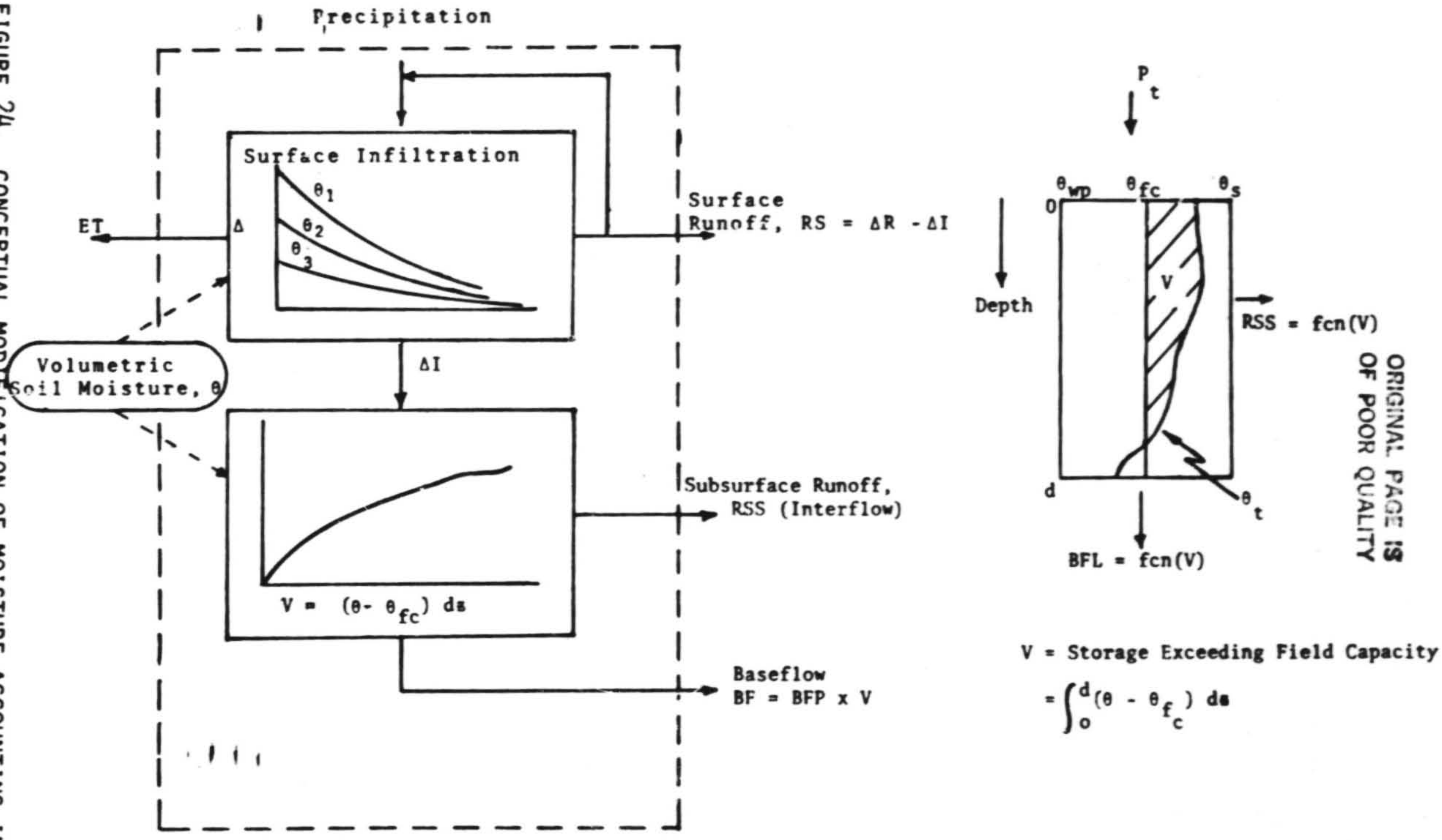
(ROP) vs soil moisture index (SMI) table defines the amount of rainfall which infiltrates to satisfy soil moisture deficits. When the soil moisture content reaches field capacity, the moisture deficits are considered replenished and SMI reaches a maximum value. The remaining rainfall, defined as runoff generated period (RGP), is then subjected to a baseflow loss considered to represent percolation from the upper soil zone to a lower storage zone. The loss is based on a table relating percent of surface runoff becoming baseflow (BFP) to a Baseflow Infiltration Index (BII). The remaining generated surface runoff (RGS) is further divided into direct surface runoff (RS) and infiltrated interflow or subsurface runoff (RSS) according to a table describing the surface-subsurface separation as a function of the total runoff, RGS. The strategy is for these tables to be adjusted by the user to fit the model response to rainfall-runoff historical data for a particular watershed.

The structure of the moisture accounting component within SSARR does not permit the direct use of the ΔI - ΔS curves. For example, the ΔI - ΔS curves could not simply be substituted for the ROP vs SMI table because the ROP vs SMI table only defines the rainfall infiltration required to replenish soil moisture deficits below field capacity. Additional surface infiltration, becoming deep percolation and interflow, is considered in other tables. The ΔI - ΔS curves, on the other hand, predict the total infiltration capacity until the surface soil moisture zone is entirely saturated. The infiltrated moisture would be subject to soil moisture replenishment, deep percolation and interflow.

Further, the SMI only reflects variation in the soil moisture condition between wilting point and field capacity. Peck et al., (1981) have noted that an empirical relationship between SMI and remotely sensed soil moisture could be developed to make use of remotely sensed measurements. However, because the SMI only reflects moisture variations between wilting point and field capacity, additional moisture variations in the range of field capacity to saturation measured by remote sensing could not be reflected in the SMI.

The use of the ΔI - ΔS curves within SSARR would require that the moisture state variable be the actual moisture content ranging from wilting point to saturation rather than the current SMI. One possible approach to revising the model structure is shown in Fig. 24. The ΔI - ΔS curves would define the separation of surface runoff, RS, from the total rainfall infiltration at the soil surface. As long as the average moisture condition was less than the field capacity, the rainfall infiltration would be considered to replenish soil moisture deficits in the surface zone. Additional infiltrated moisture exceeding the soil field capacity would be subject to deep storage percolation (BF) and near-surface lateral flow and inte (RSS). An additional table would define the division of base flow (BF) and interflow (RSS) as a function of the soil moisture storage exceeding the soil field capacity.

In using this approach, the division of rainfall into replenished near-surface soil moisture, surface runoff, interflow and baseflow is accomplished through two tabular relationships rather than the current three tables. In addition, both relationships would be a function of a single state variable,



(a) REVISED MODEL STRUCTURE

(b) ELEMENTED SOIL COLUMN

FIGURE 24. CONCEPTUAL MODIFICATION OF MOISTURE ACCOUNTING IN SSARR

volumetric soil moisture content rather than the two variables SMI and BII. The first table (ΔI - ΔS curves) would be defined by the soil texture and the current moisture conditions at the beginning of the rainfall. The second table is more empirical in nature and would need to be defined through a calibration process using rainfall-runoff data from the particular watershed.

CHAPTER 5

THOUGHTS ON PROBLEMS OF SPATIAL VARIABILITY

The technique presented in this paper and illustrated in Table 7 estimates infiltration capacities at a point. Representative soil hydraulic properties based on the SCS soil groups and a remotely sensed estimate of the surface soil moisture serve as inputs to the model. As shown in Fig. 25, taken from a parallel study conducted by Wilkening (1981), significant errors can result if the estimate of surface soil moisture is in error. Figure 25 is based on a correct surface soil moisture of .25 cm/cm. If the hydrologist incorrectly assumes the soil moisture to be higher or lower, he would get the indicated error based on experiments with numerical solutions of the Richard's equation. For example, when the moisture content is actually .25 and he incorrectly estimates it to be .28 for an analysis with the Richard's equation, he would compute the surface runoff to be approximately 60% higher than actual for a rainfall of 5.1 cm (2.0 inches). Similar patterns are encountered for improper estimates of hydraulic conductivity, porosity, etc.

It is widely recognized that even though a soil type may be considered to be relatively homogeneous, significant variations in soil moisture and hydraulic properties over an area can be expected to occur. The sensitivity illustrated by Fig. 25 and the recognized natural spatial variability in soil moisture and soil properties can be quite important in developing a spatially distributed hydrologic model. The general approach in a hydrologic model is to compute a representative infiltration rate and rainfall excess for a discrete areal unit rather than for a point. The trends exhibited by Fig. 25 can be used to estimate the spatial resolution or cell size to be used in a distributed hydrologic model.

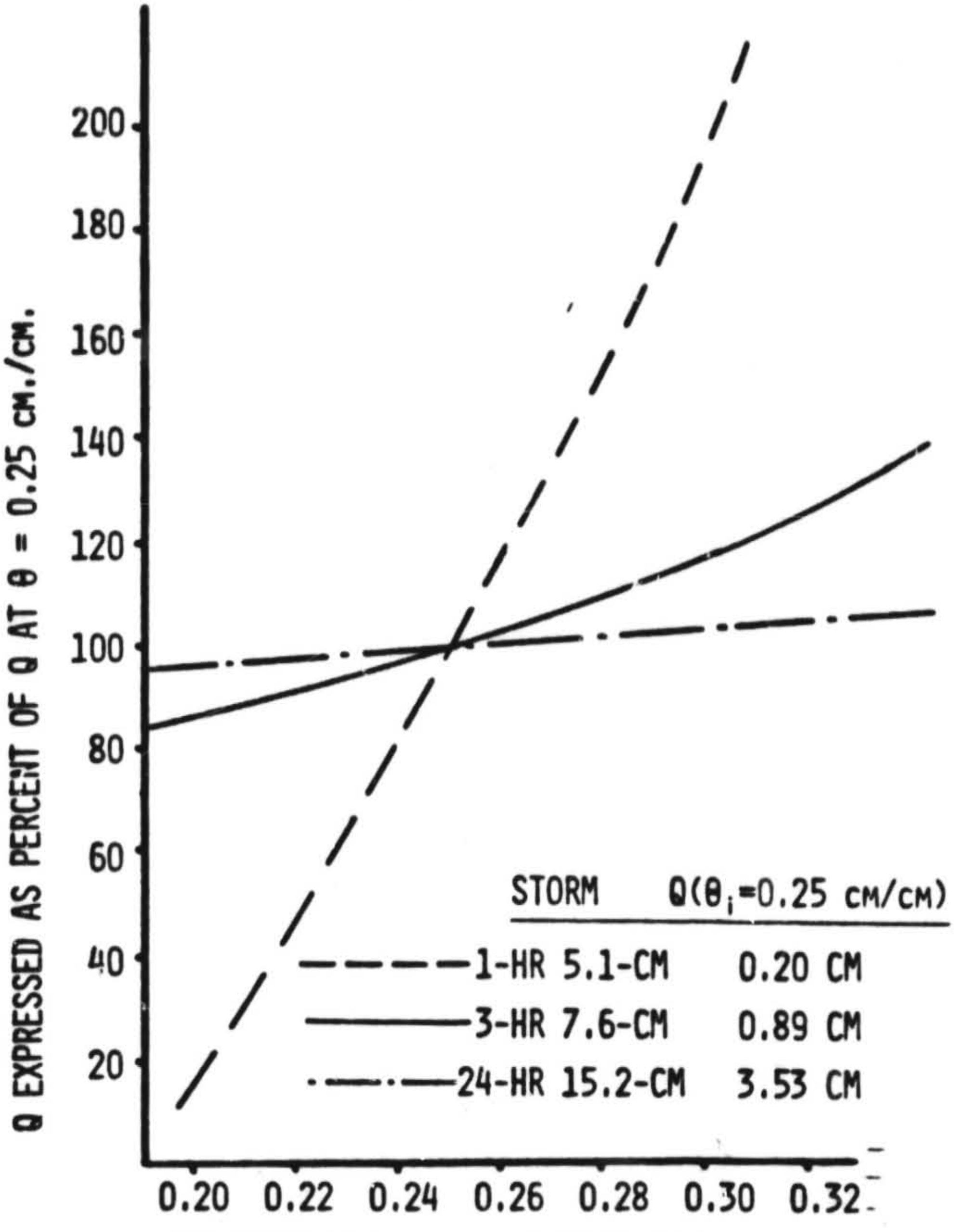


FIGURE 25. PERCENTAGE VARIATION IN EXCESS RAINFALL AS FUNCTION OF ERROR IN ESTIMATE OF ANTECEDENT MOISTURE

Figure 26 is an idealized representation of a watershed. Two tributaries merge into a main channel draining a watershed having a soil of generally uniform texture. There is, however, spatial variability in soil moisture and hydraulic properties. Because of variations in the land use and structure of a micro-drainage network, the hydrologist may want to represent the watershed as an array of cells linked in accordance with Fig. 26. However, if one uses a mean soil moisture measurement from some remote sensing platform or a mean hydraulic parameter based on the soil texture, large errors in the estimate of rainfall excess may result in many of the cells because of the inherent spatial variations of these parameters. As the cell size is increased, however, the resulting areal average of the soil moisture within the cell should approach the mean as a representative value. Thus, the desired spatial resolution may have to be increased to that shown in Fig. 26b in order to maintain some level of precision in the estimate of the rainfall excess.

This concept of defining errors in terms of spatial resolution has been examined, on a preliminary level, using field data representing an example of soil moisture variation within a 23 acre watershed instrumented by the USDA in Chickasha, Oklahoma. The watershed is well-managed pasture with soil variations occurring within one soil texture class. Soil moisture conditions in the top five centimeters were represented by the contour map shown in Fig. 27.

Using a 17 acre rectangular area within the watershed, a 48x48 retangular grid matrix was overlaid onto the contour map and encoded cell by cell to represent the moisture variation as an array of 2034 cells. Each cell was

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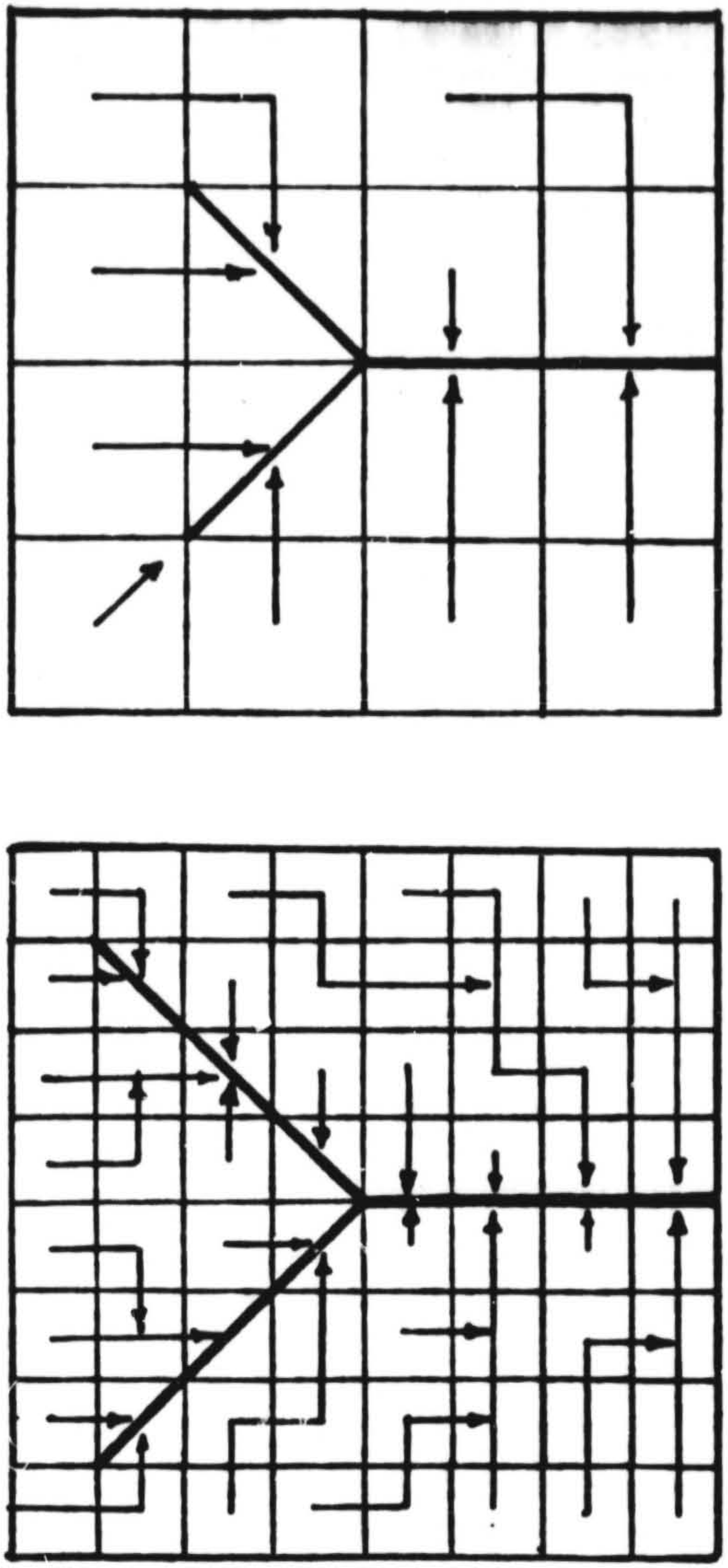


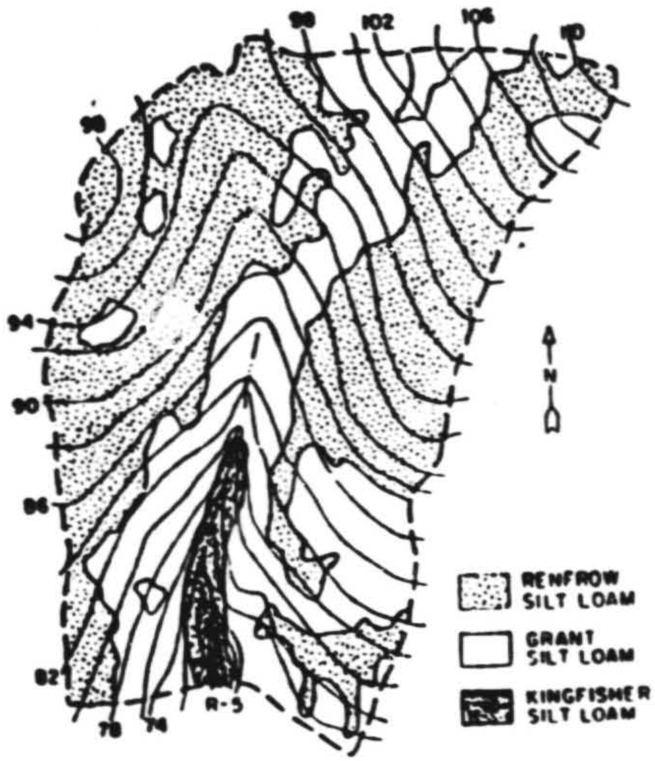
ACCEPTABLE LEVEL OF SPATIAL RESOLUTION TO ALLOW FOR PRECISION IN ESTIMATION OF RAINFALL EXCESS

b

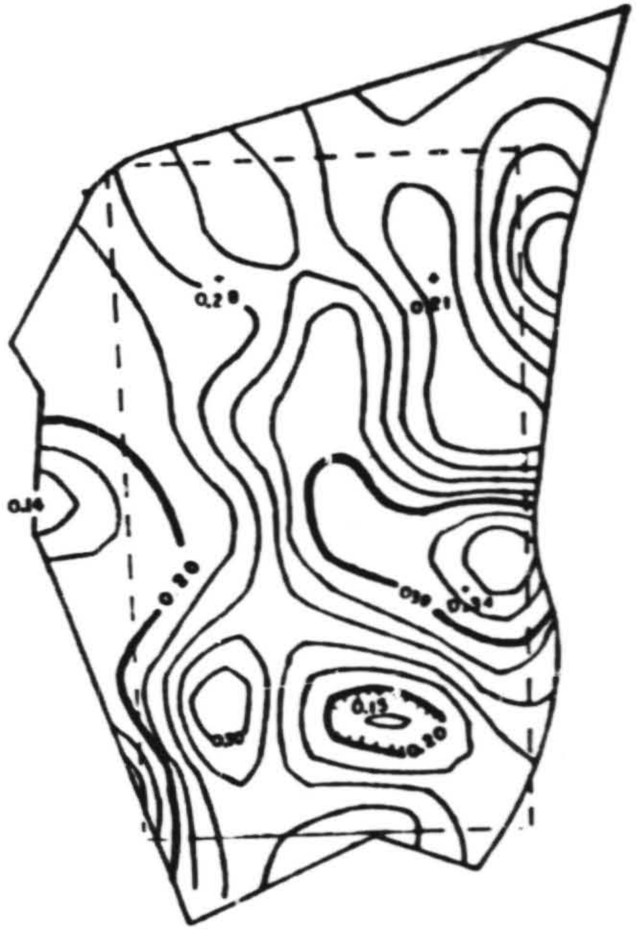
DESIRED LEVEL OF SPATIAL RESOLUTION

a

FIGURE 26. IDEALIZED SPATIALLY DISTRIBUTED WATERSHED



SOILS AND TOPOGRAPHIC MAP



SOIL MOISTURE VARIATION (TOP 5 CM.)

MAY 10, 1978

FIGURE 27. WATERSHED R5 CHICKASHA, OKLAHOMA (23 ACRES)

approximately 7 m x 4 m and represented about 0.04% of the entire 17 acre area. The mean soil moisture was calculated as .25 cm/cm and the coefficient of variation was 14% for this data set.

Because the contour map only reflects general trends in moisture variation, two additional distributions with coefficients of variation of 17% and 24% were derived from the original matrix of cells by allowing each moisture value to vary by introducing a random component through the use of a Markov chain. Figure 28 shows the symbolic maps for coefficients of variation of 14% and 17%.

Assume that the hydrologist wants to represent this watershed as a spatially distributed hydrologic model arranged as an array of cells. The rainfall excess in each cell will be computed based on the use of the mean soil moisture of .25 cm/cm. However, we do not want the computed rainfall excess to deviate from the actual theoretical rainfall excess by more than 10% in any cell. If we use the three hour storm of Fig. 25, we can accept an error of ±0.03 cm/cm (Range: .22-.28cm/cm) in the soil moisture and still estimate a rainfall excess that will be within 10% of the correct valve. Thus the question becomes, how large must our spatial resolution element be in order that the average moisture within the cell be wihtin ±.03 cm/cm of the mean soil moisture.

By aggregating the cells into new spatial units such as 2x2, 3x3, 6x6, etc., matrices some interesting insights were gained. For the area having a coefficient of variation of 14%, if we attempt to model infiltration in each of the 2304 cells based on the assumption that the mean soil moisture of .25 is representative for that cell, the computed rainfall excess would be in error by more than 10% in 40% of the cells. If the spatial resolution were increased to .13 acres (.75% of

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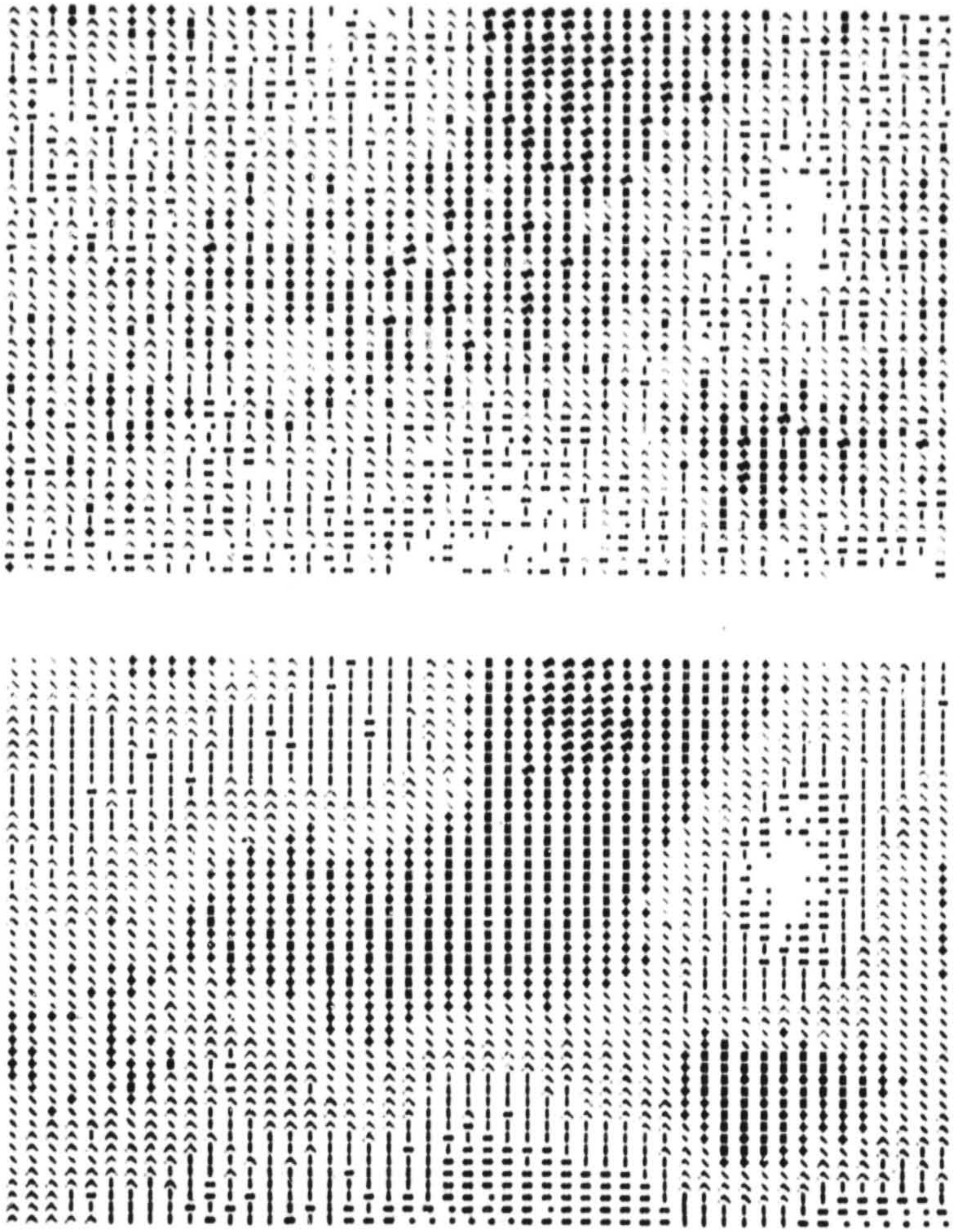

CV = 14 %

CV = 17 %

FIGURE 28. SOIL MOISTURE VARIABILITY IN CELL MATRIX

the total area) 33% of the cells would exceed the 10% error limit. If the resolution was again increased to .51 acres (3% of the overall area) 20% of the cells would exceed the limit. Finally, if the cell resolution was increased to 1.87 acres (11% of the total area) the rainfall excess would be predicted within 10% of the actual values in all of the cells. In similar examinations for the coefficients of 17% and 24% results were nearly identical.

It is recognized that this examination of spatial variability is very preliminary and that significant additional work must be undertaken. However it does illustrate an important consideration that hydrologists are going to have to make when they consider structuring spatially varied models. For example, suppose the resolution element of a sensor estimating surface soil moisture is limited by technology or economics to a .5 kilometer square. This would give an area of .25 kilometers or approximately 62 acres. Thus, the sensor would give us the mean soil moisture for a 62 acre cell. The hydrologist could use this mean soil moisture as the input to the model illustrated in Table 7 to develop the rainfall excess as for a spatially distributed model having 62 acre cells. However, he would have to look at the coefficient of variation and consider the problems outlined in this chapter if it was necessary to further subdivide the watershed into smaller cells to simulate the consequences of spatial variability in other hydrologic processes.

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CHAPTER 6

CONCLUSIONS

Rainfall infiltraion rates are sensitive to the antecedent soil moisture conditions and the hydraulic properties of the soils. These sensitivities have been investigated through the use of numerical solutions of partial differential equations that describe flow through unsaturated porous media. Although these numerical solutions provide good estimates of the infiltration rates under a variety of conditions, their use in practical hydrologic models designed to simulate watershed runoff is prohibited by the complexity of their solutions and extensive computer running times.

A series of functional relationships are presented that allow the estimation of time varying infiltration capacities for an array of initial moisture conditions and soil hydraulic properties. The procedure was developed using numerical experiments with the Richard's equation for initial moisture and soil textures that can be expected in watershed hydrology. The approach is designed to utilize remote sensing techniques to define the initial soil moisture and Soil Conservation Service soil survey maps to estimate the soil properties.

Examples are presented that show infiltration rates predicted by this relatively simple technique compare very closely with those computed with numerical sensitivity to the initial soil moisture centers on that moisture stored in the surface layer, approximately five centimeters, of the soil column. Below that, there is an influence created by variations in the soil moisture profile, but the impact of the profile variation is relatively small by comparison with other uncertainties that one would expect in the watershed.

The procedure presented has several key advantages: 1) it is straight forward and can be implemented by the practicing hydrologist using a desk top procedure or a very efficient computer program; 2) it follows the familiar infiltration-storage concept and, therefore, can be bridged into a number of current model methodologies; 3) it logically incorporates the impact of initial soil moisture conditions on the watershed response in terms of the physical principles governing the process; 4) the parameters are physically based and can be defined operationally from the SCS hydrologic soil group classifications presented in soil survey maps; and 5) it is structured to utilize near surface soil moisture that can be developed from remote sensing techniques.

CHAPTER 7

RECOMMENDATIONS

The following recommendations might be considered for the further development of the approach taken in this paper:

1) An attempt to modify the infiltration components within several current models such as USDAHL and SSARR and then apply these models to watershed applications would be worthwhile. Such an effort would indicate the amount of work required in replacing the current infiltration component with the ΔI - ΔS curves. In addition, the real benefits could be evaluated in watershed applications.

2) Finally, an attempt could be made to develop similar approaches for a more complex representation of the soil system including: stratified soil properties within the soil column; vegetative root zone and surface landcover effects; the spatial variability of moisture conditions and soil properties even with a "homogeneous" modeling unit. Attempts could be made to apply the ΔI - ΔS curves given in this study to more complex systems. For example, perhaps a combination of functions could be used to represent a stratified soil system. In addition, a more complex theoretical model might be used to incorporate the effects of surface vegetation such as the finite element solution given by Neumann et al., (1975) and Feddes et al., (1978).

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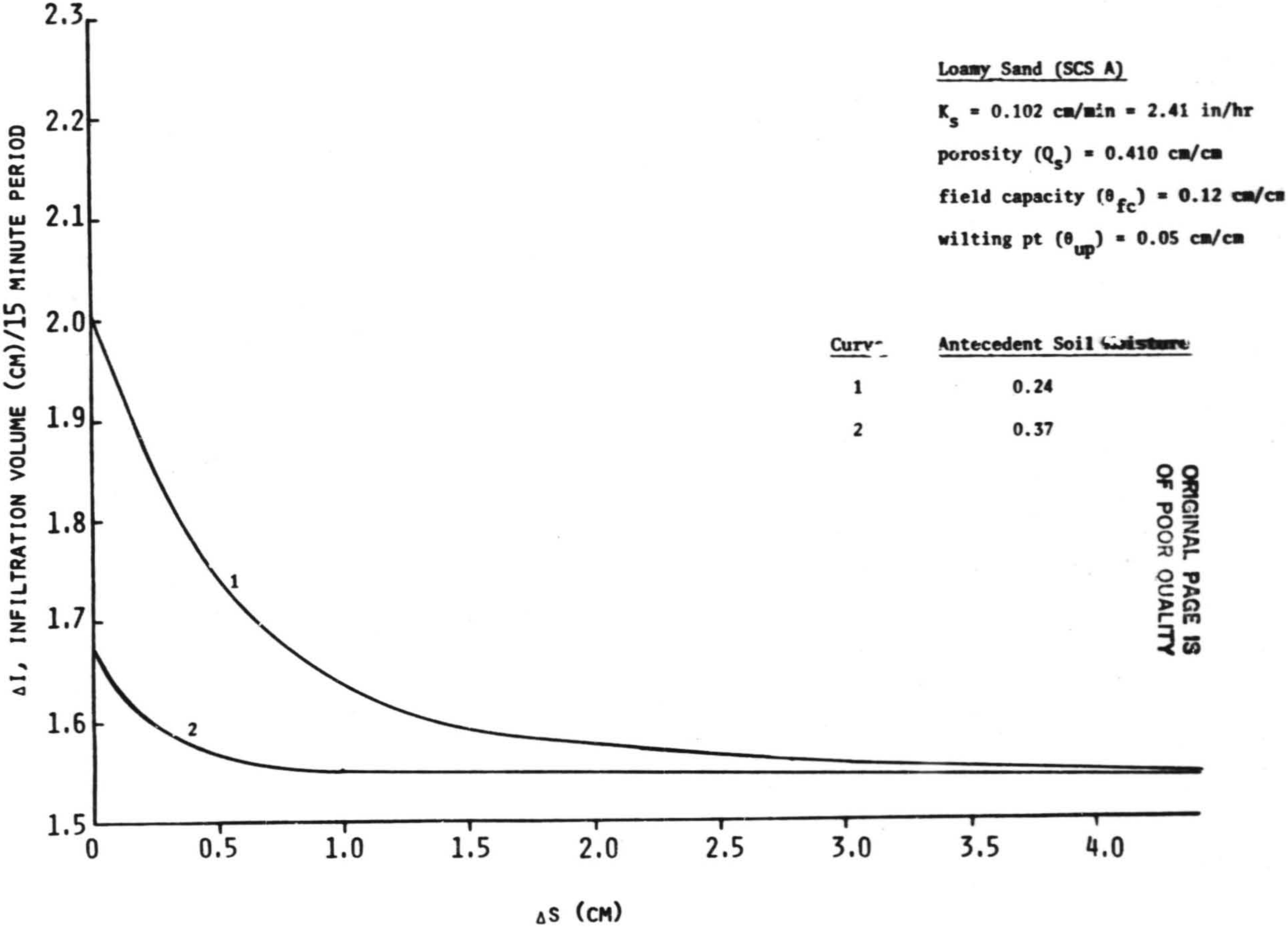
APPENDIX A:

REPRESENTATIVE ΔI - ΔS CURVES FOR THE FOLLOWING SOILS:

Texture	SCS Soil Group
Loamy Sand	A
Sandy Loam	B
Loam	B
Sandy Clay Loam	C
Clay Loam	C
Silty Clay	D



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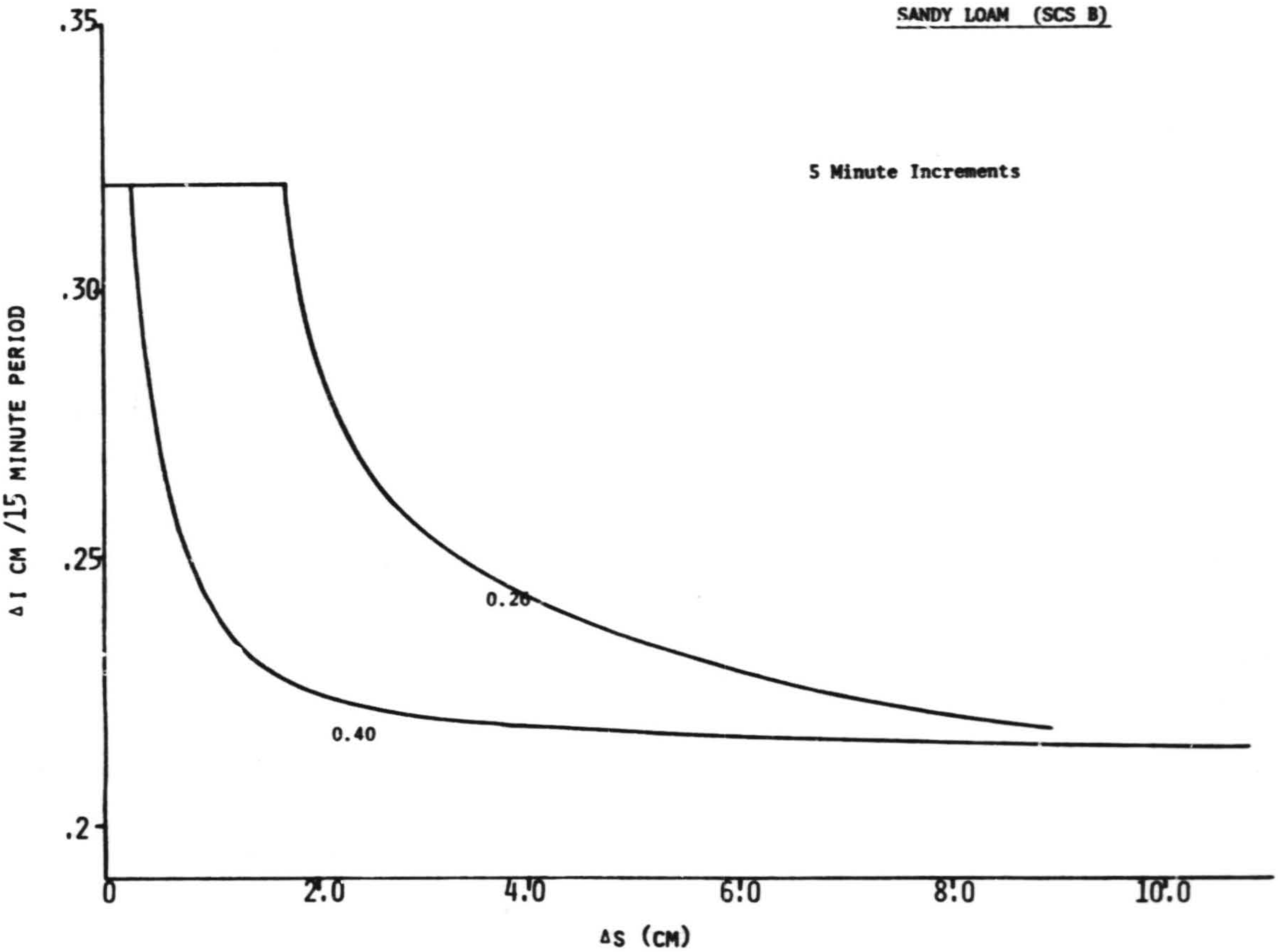
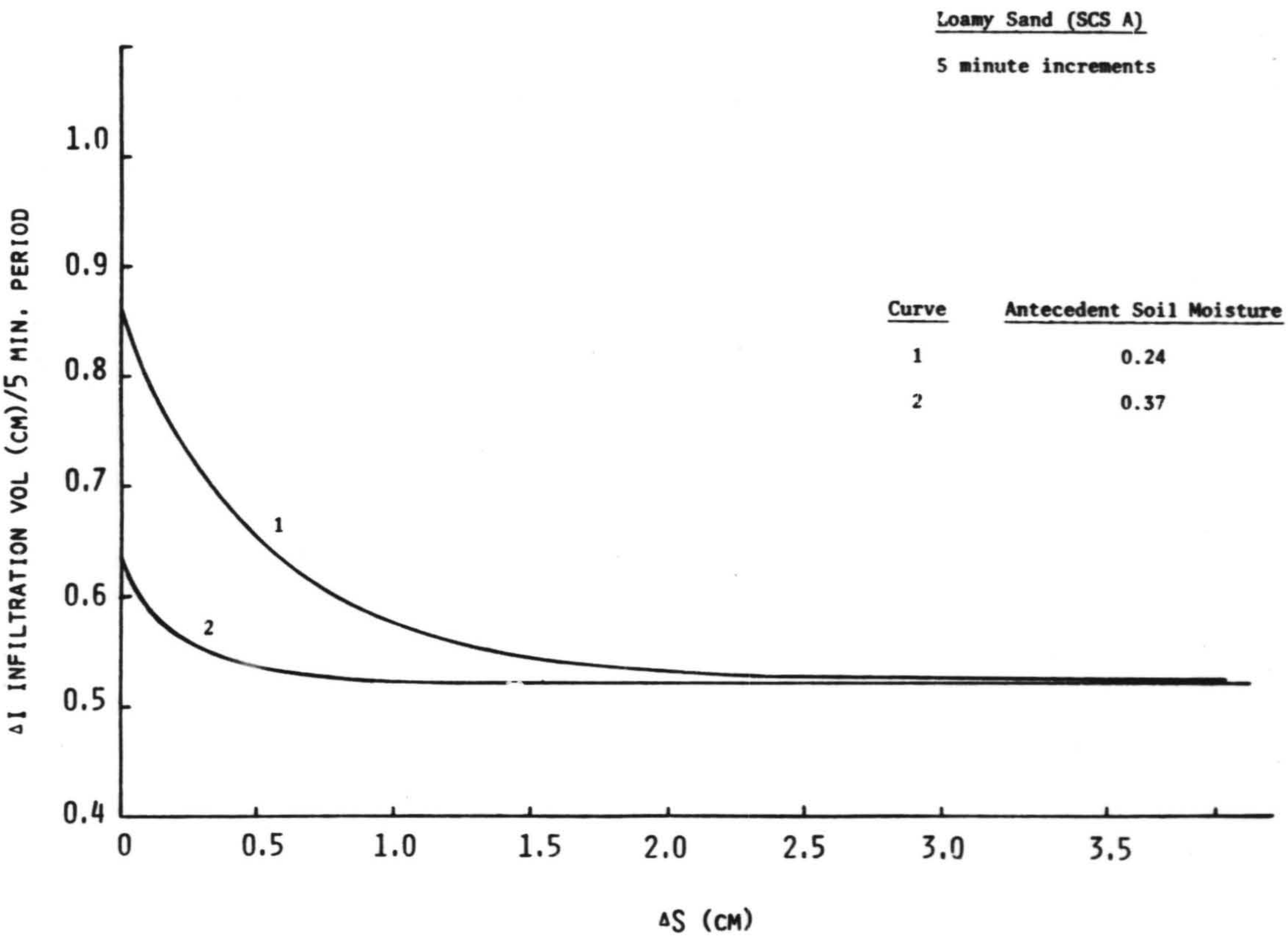
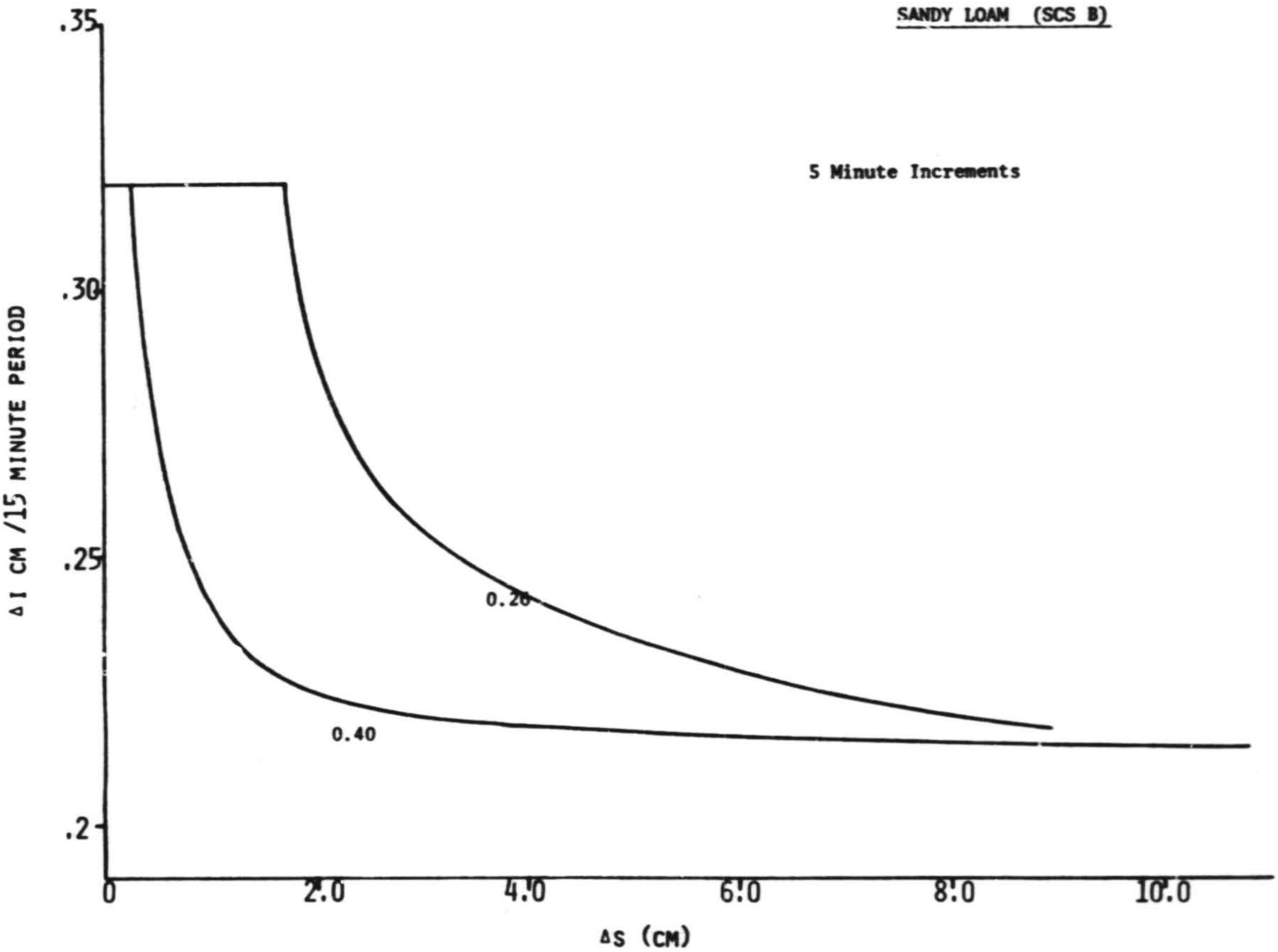
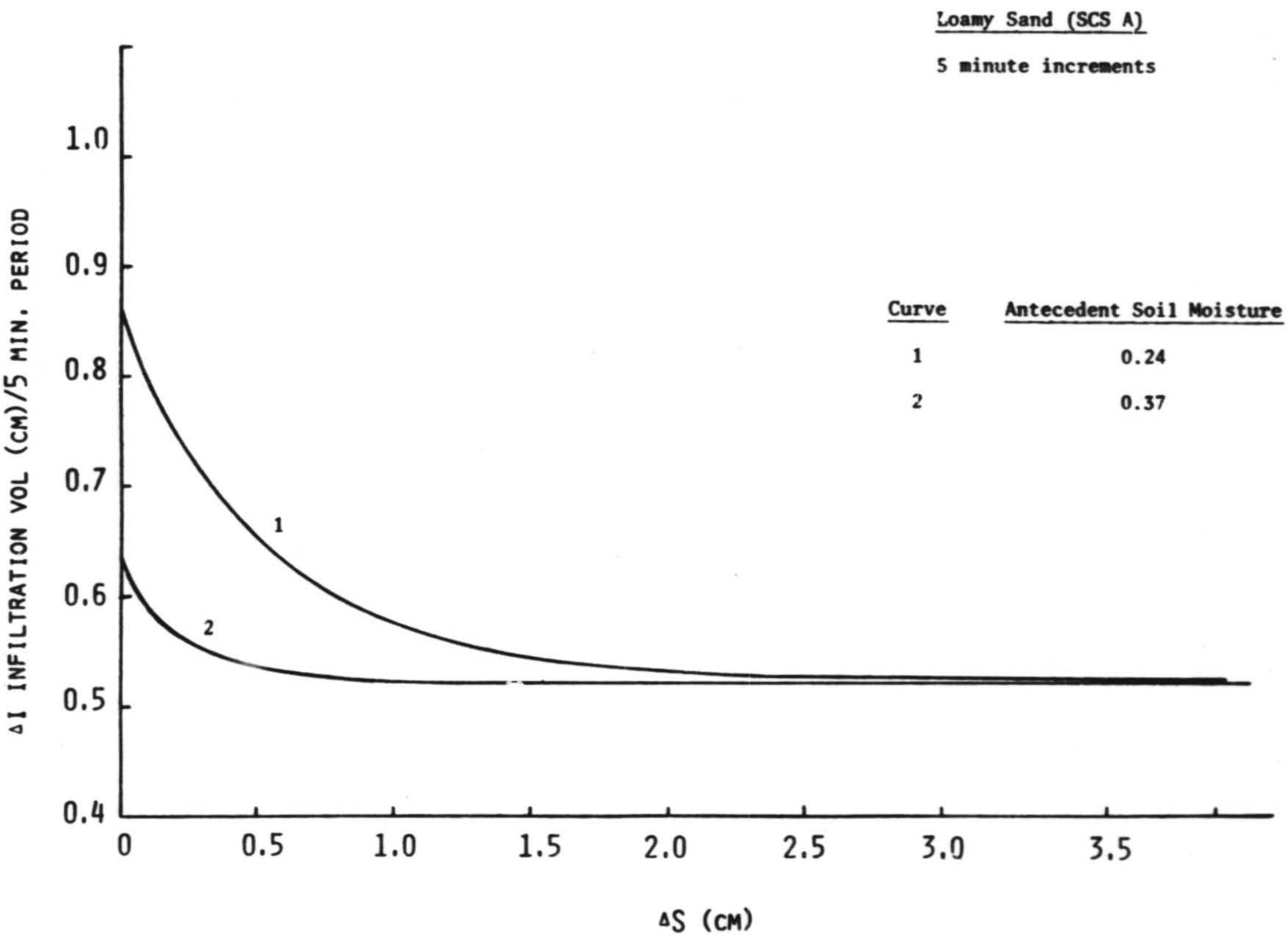
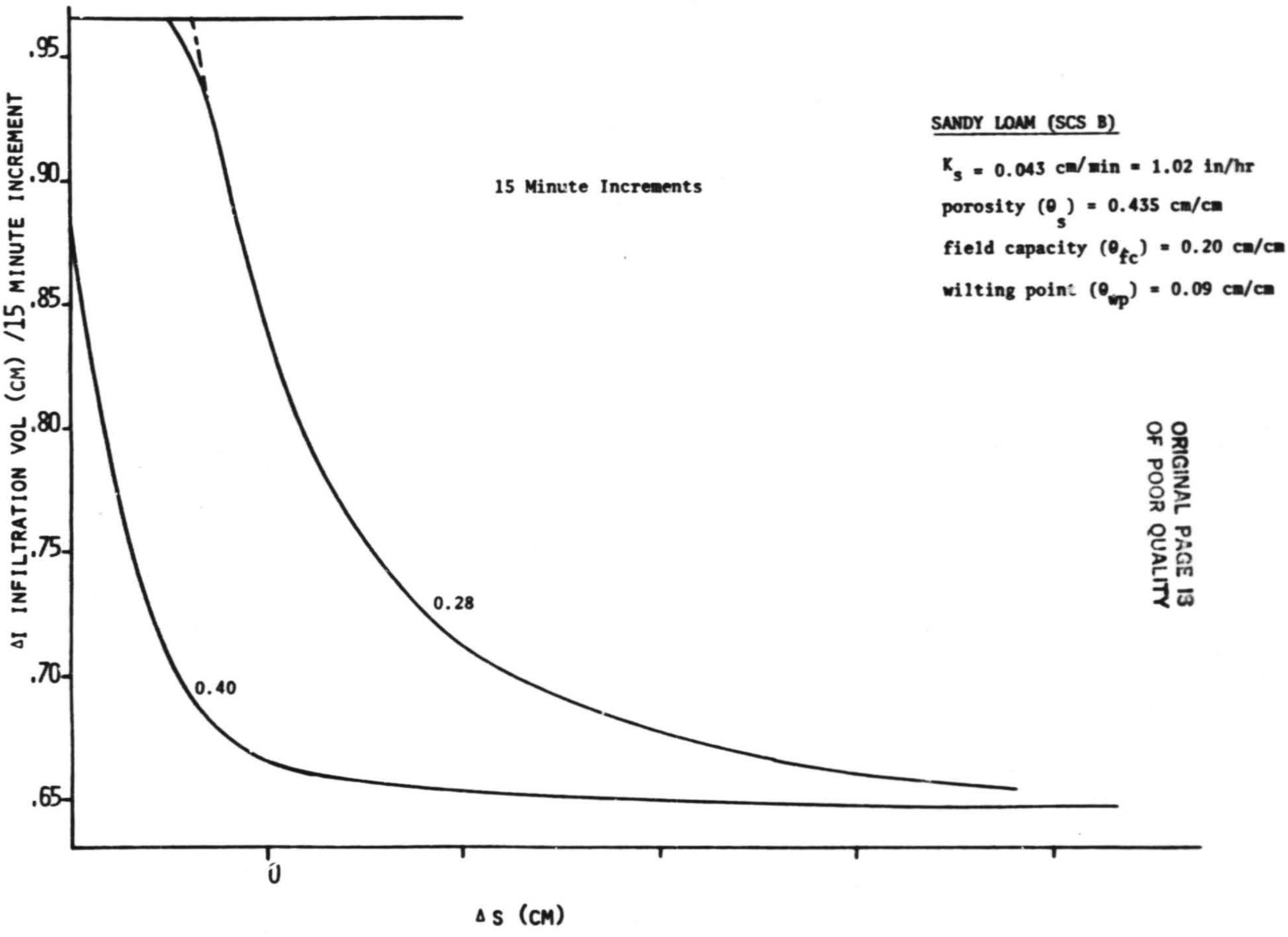

SANDY LOAM (SCS B)
5 Minute Increments
ΔI CM /15 MINUTE PERIOD
.35
.30
.25
.2
0.20
0.40
0
2.0
4.0
6.0
8.0
10.0
ΔS (CM)


Loamy Sand (SCS A)
5 minute increments
Curve
Antecedent Soil Moisture
1
0.24
2
0.37
ΔI INFILTRATION VOL (CM)/5 MIN. PERIOD
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
ΔS (CM)
1
2

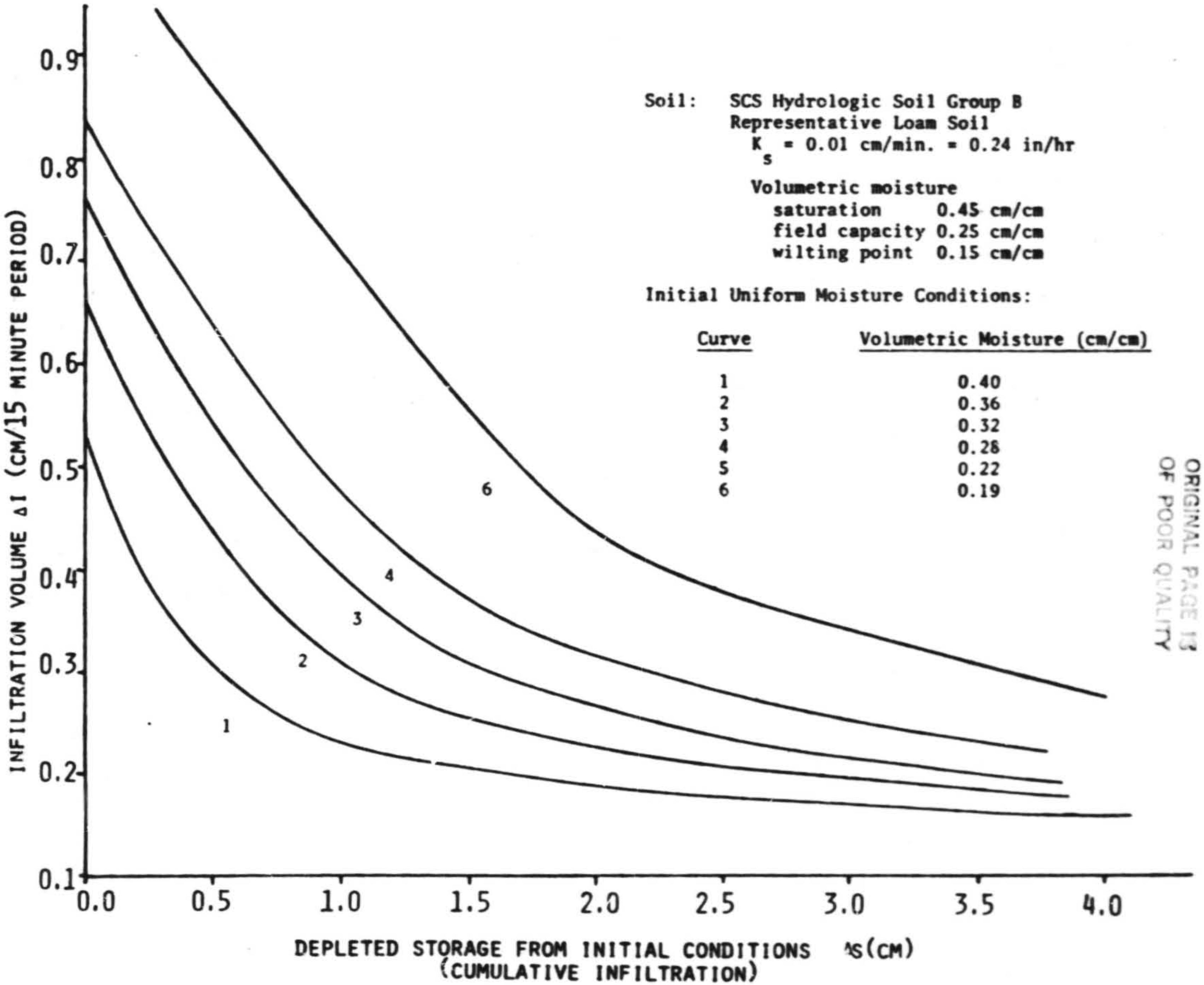

SANDY LOAM (SCS B)
5 Minute Increments
ΔI CM /15 MINUTE PERIOD
.35
.30
.25
.2
0.20
0.40
0
2.0
4.0
6.0
8.0
10.0
ΔS (CM)



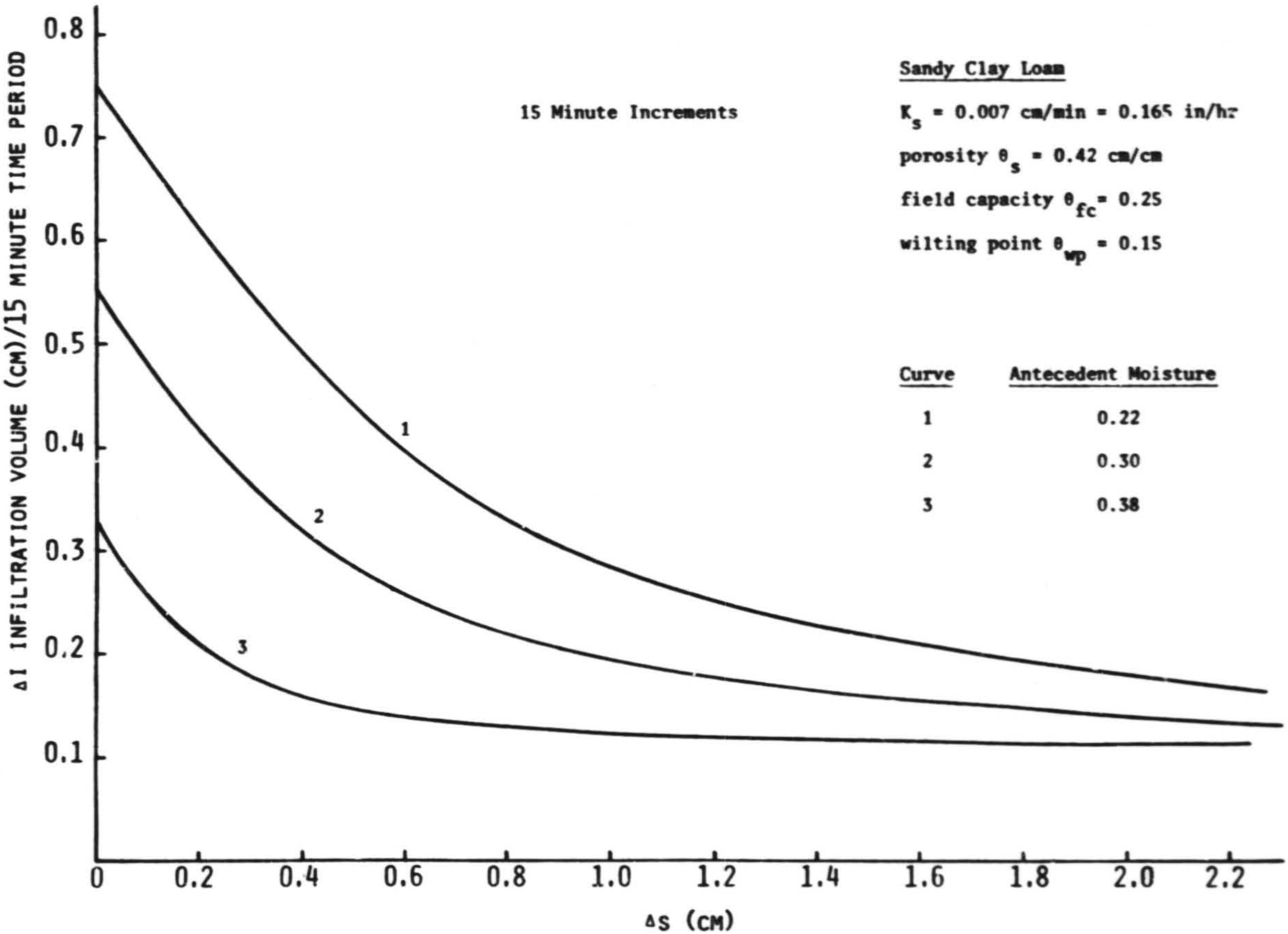
Loamy Sand (SCS A)
5 minute increments
Curve
Antecedent Soil Moisture
1
0.24
2
0.37
ΔI INFILTRATION VOL (CM)/5 MIN. PERIOD
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
ΔS (CM)
1
2



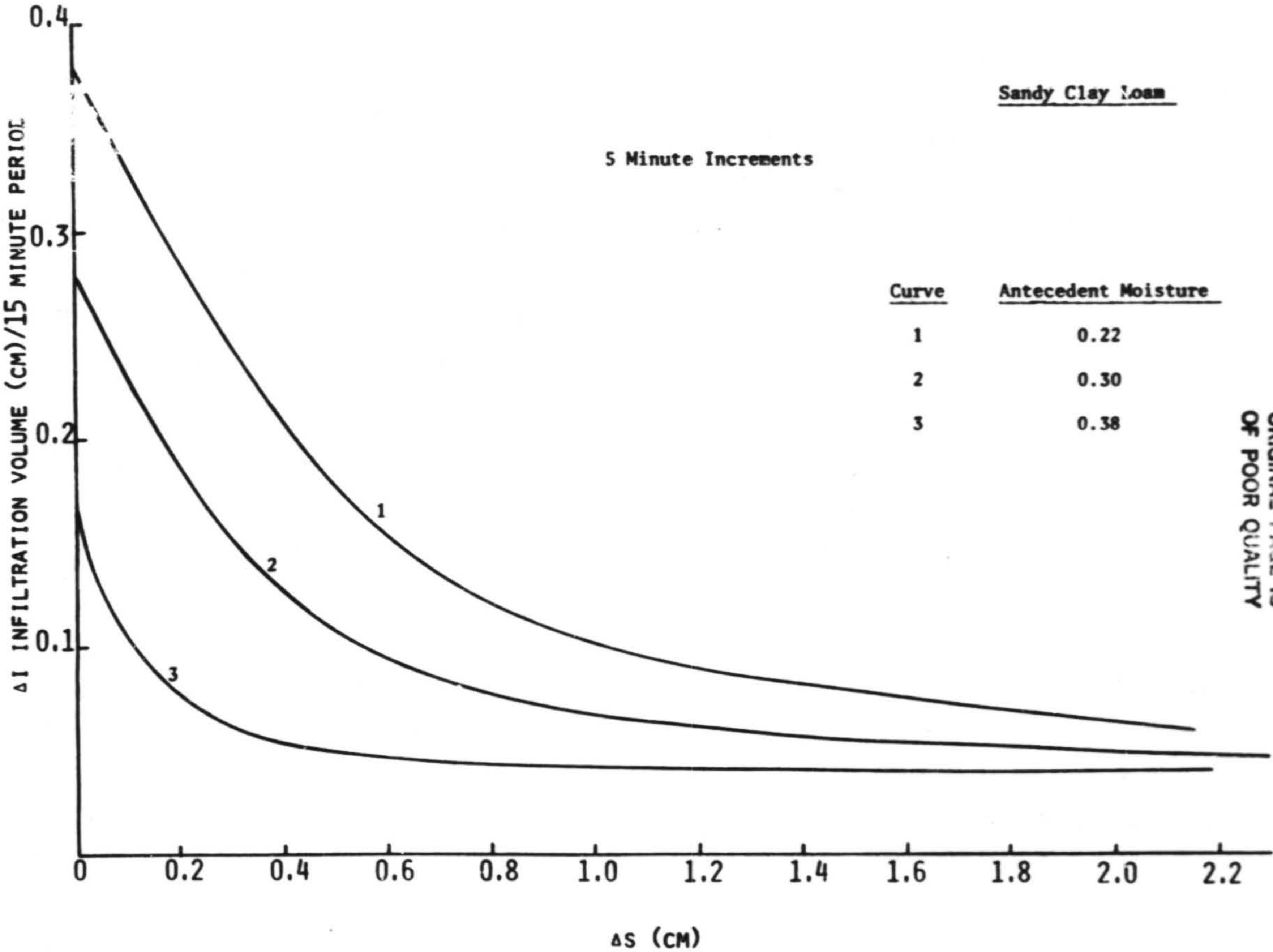
15 Minute Increments
SANDY LOAM (SCS B)
K_s = 0.043 cm/min = 1.02 in/hr
porosity (θ_s) = 0.435 cm/cm
field capacity (θ_{fc}) = 0.20 cm/cm
wilting point (θ_{wp}) = 0.09 cm/cm
0.28
0.40
.95
.90
.85
.80
.75
.70
.65
0
ΔI INFILTRATION VOL (CM) /15 MINUTE INCREMENT
ΔS (CM)



INFILTRATION VOLUME ΔI (CM/15 MINUTE PERIOD)
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0
DEPLETED STORAGE FROM INITIAL CONDITIONS ΔS(CM)
(CUMULATIVE INFILTRATION)
1
2
3
4
6
Soil: SCS Hydrologic Soil Group B
Representative Loam Soil
K_s = 0.01 cm/min. = 0.24 in/hr
Volumetric moisture
saturation 0.45 cm/cm
field capacity 0.25 cm/cm
wilting point 0.15 cm/cm
Initial Uniform Moisture Conditions:
Curve Volumetric Moisture (cm/cm)
1 0.40
2 0.36
3 0.32
4 0.28
5 0.22
6 0.19



15 Minute Increments
Sandy Clay Loam
K_s = 0.007 cm/min = 0.165 in/hr
porosity θ_s = 0.42 cm/cm
field capacity θ_fc = 0.25
wilting point θ_wp = 0.15
Curve Antecedent Moisture
1 0.22
2 0.30
3 0.38
ΔI INFILTRATION VOLUME (CM)/15 MINUTE TIME PERIOD
ΔS (CM)
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0.2
0.4
0.6
0.8
1.0
1.2
1.4
1.6
1.8
2.0
2.2



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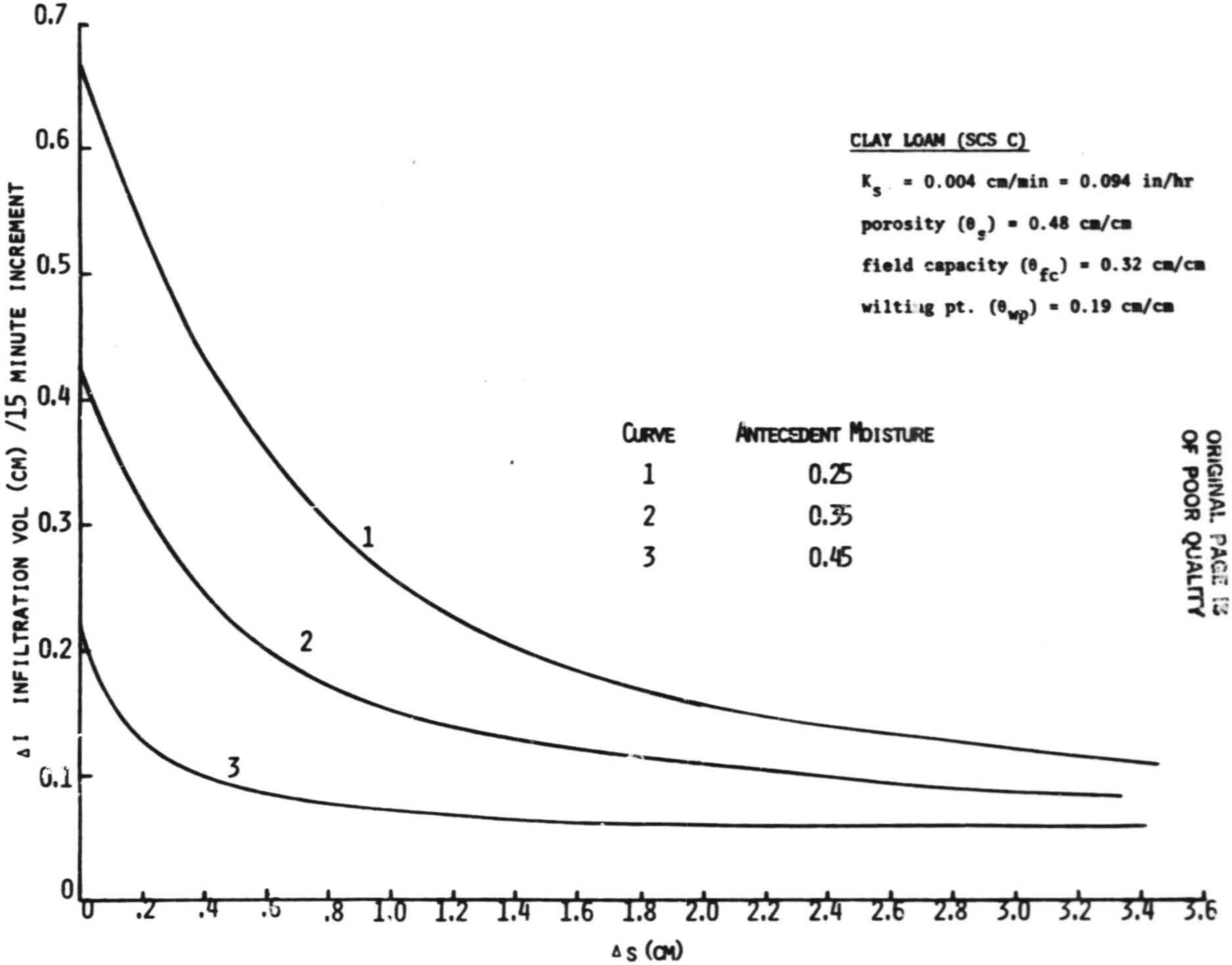

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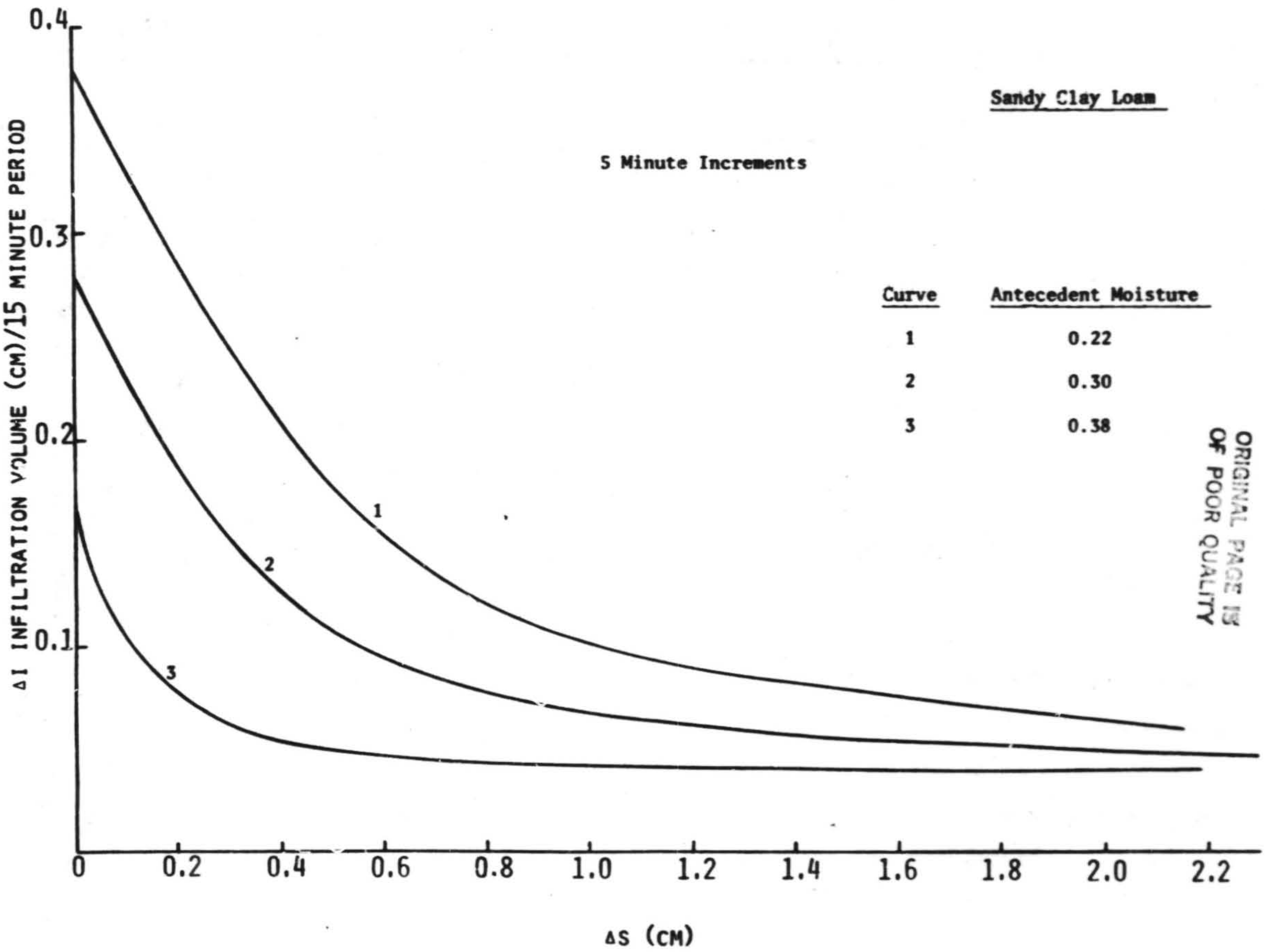
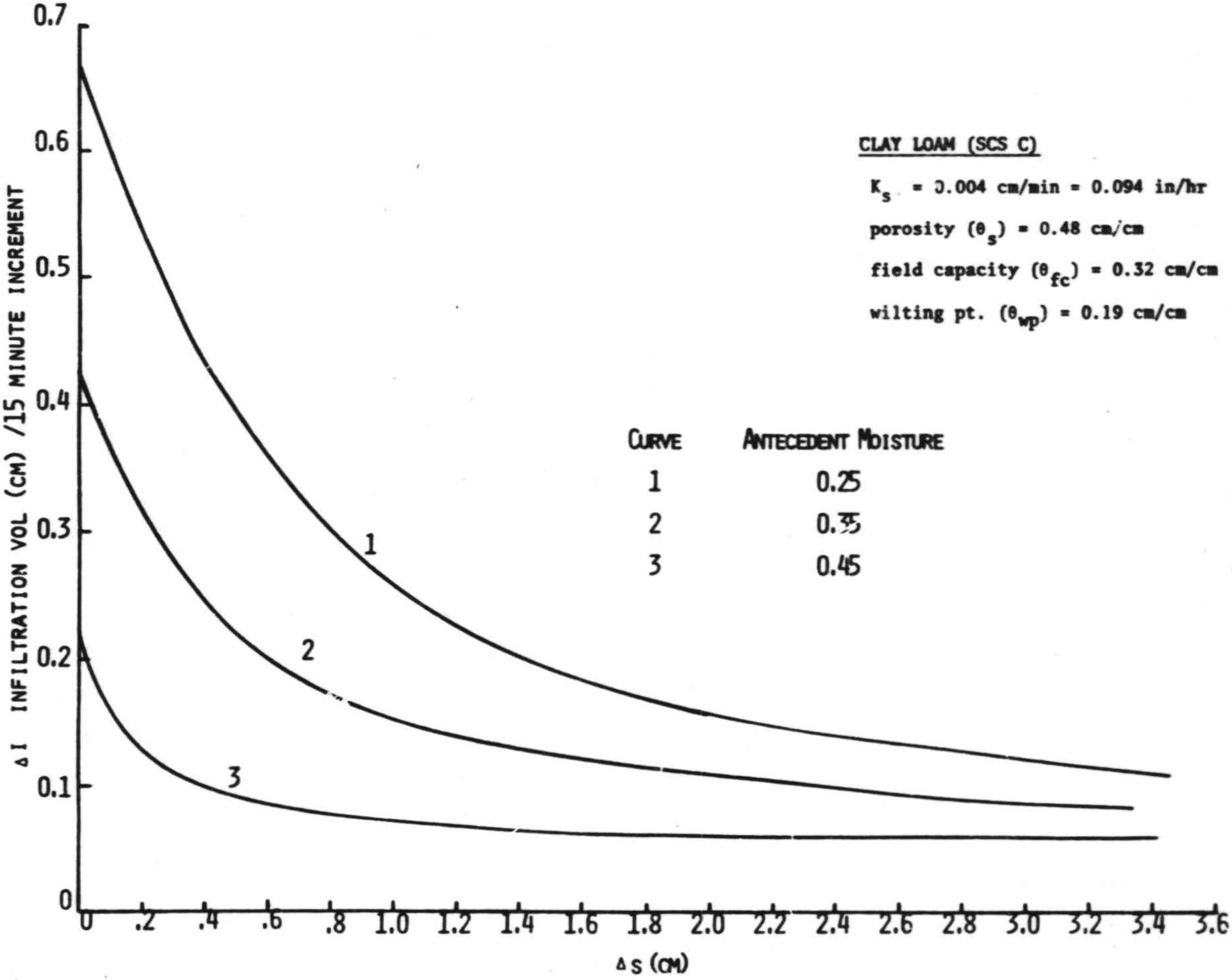
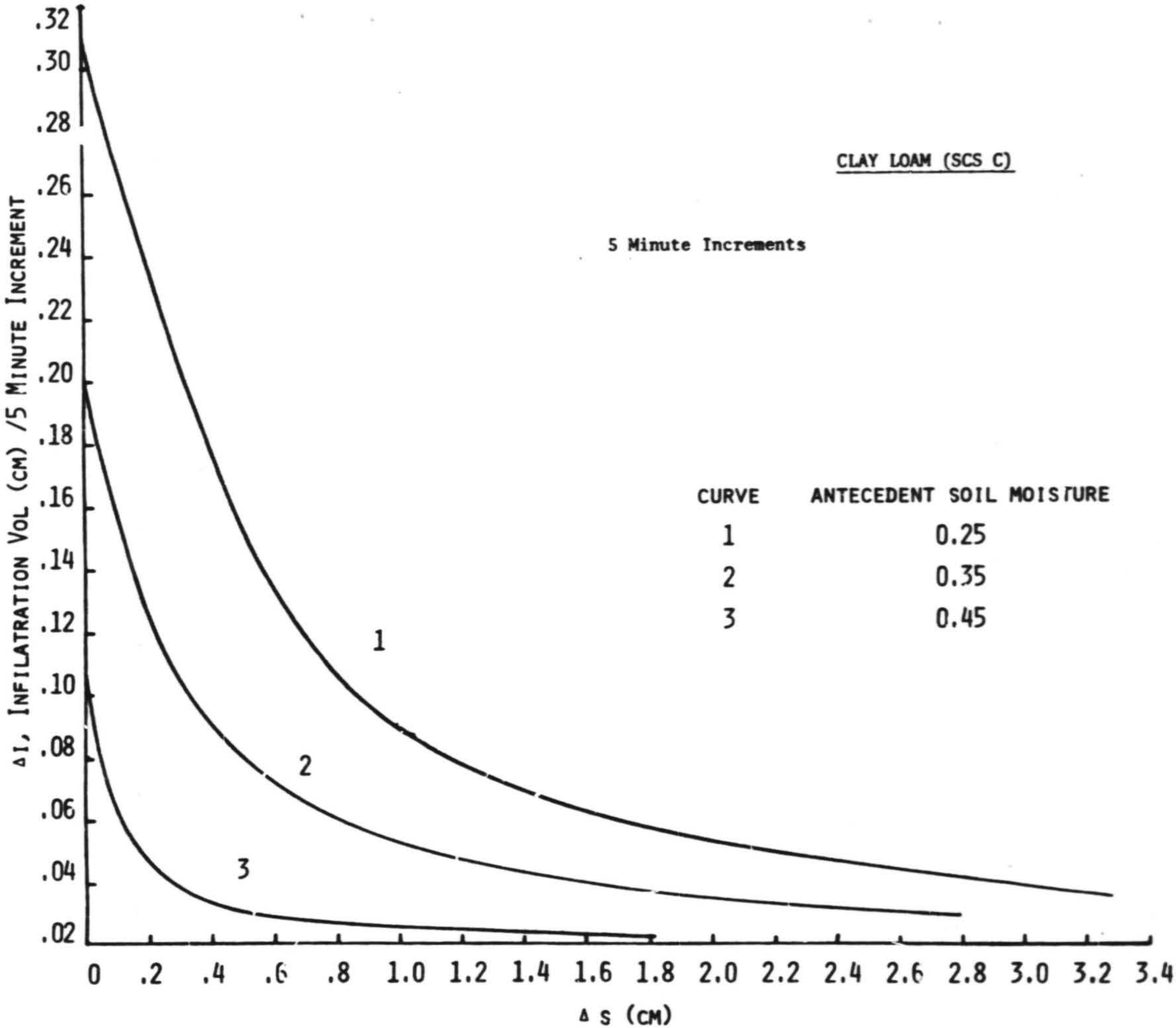
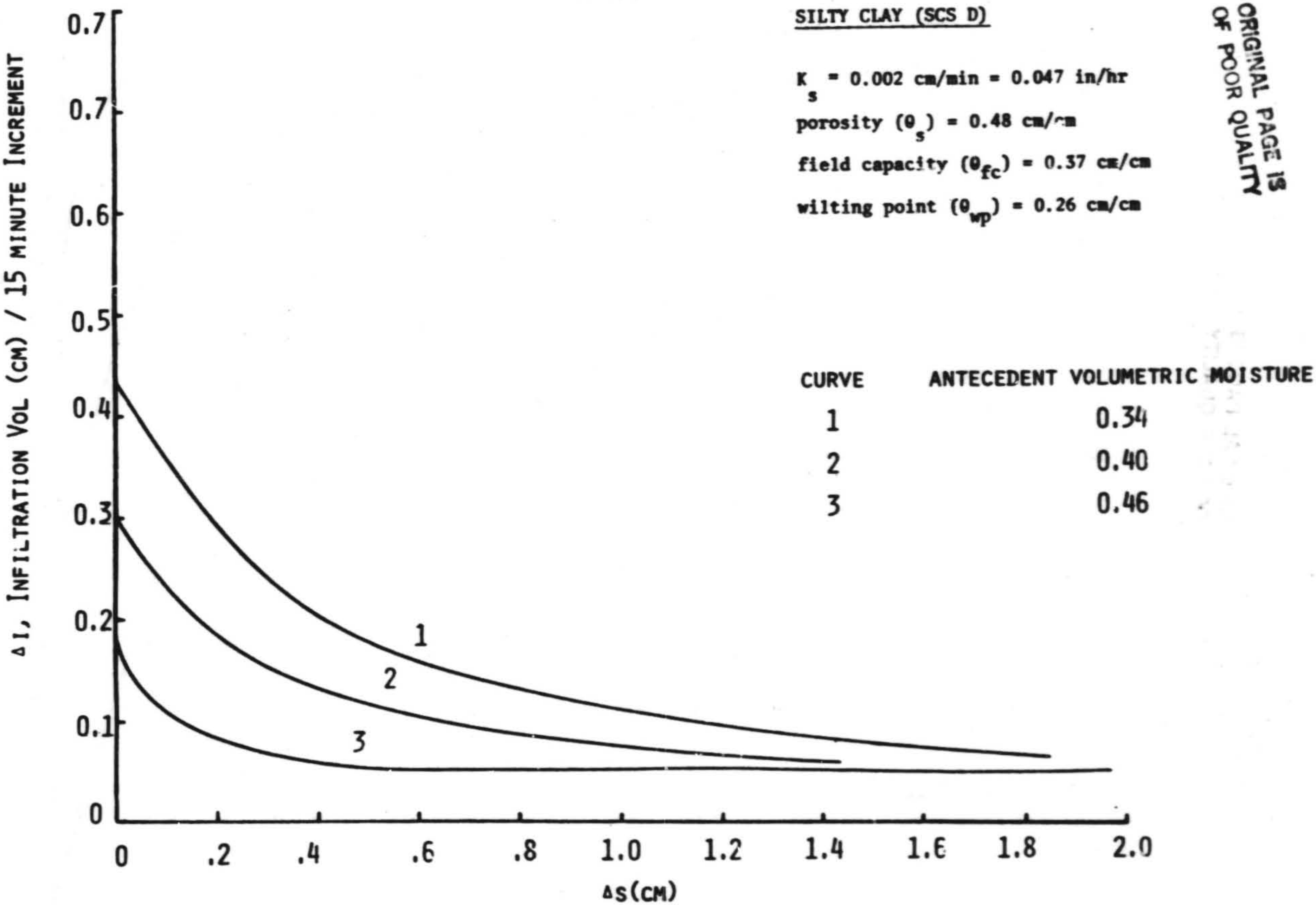
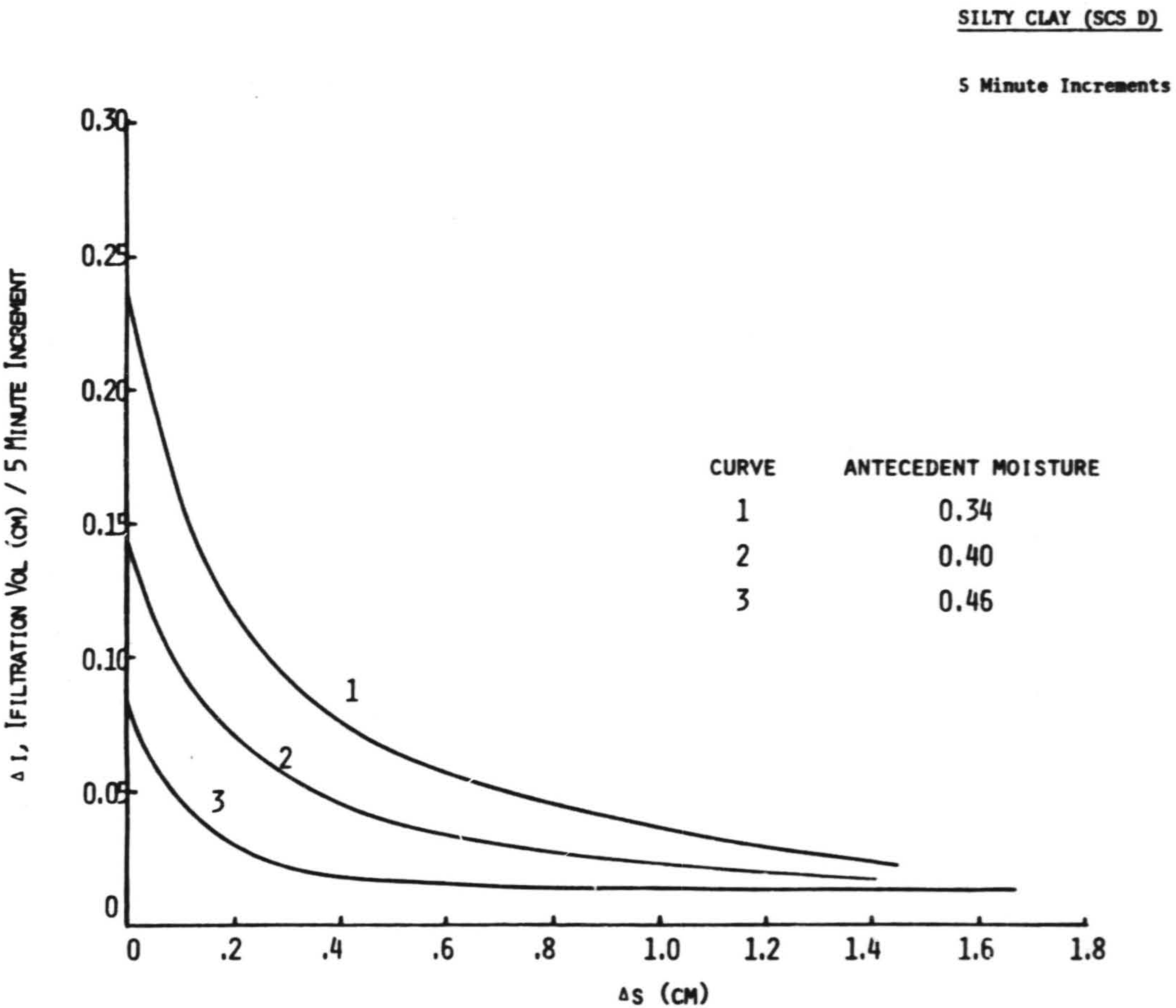

Sandy Clay Loam
5 Minute Increments
Curve
Antecedent Moisture
1
0.22
2
0.30
3
0.38
1
2
3
ΔI INFILTRATION VOLUME (CM)/15 MINUTE PERIOD
0.4
0.3
0.2
0.1
0
0.2
0.4
0.6
0.8
1.0
1.2
1.4
1.6
1.8
2.0
2.2
ΔS (CM)


CLAY LOAM (SCS C)
K_s = 0.004 cm/min = 0.094 in/hr
porosity (θ_s) = 0.48 cm/cm
field capacity (θ_{fc}) = 0.32 cm/cm
wilting pt. (θ_{wp}) = 0.19 cm/cm
CURVE ANTECEDENT MOISTURE
1 0.25
2 0.35
3 0.45
1
2
3
ΔI INFILTRATION VOL (CM) /15 MINUTE INCREMENT
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0 .2 .4 .6 .8 1.0 1.2 1.4 1.6 1.8 2.0 2.2 2.4 2.6 2.8 3.0 3.2 3.4 3.6
ΔS (CM)


CLAY LOAM (SCS C)
5 Minute Increments
CURVE ANTECEDENT SOIL MOISTURE
1 0.25
2 0.35
3 0.45
ΔI, INFILATRATION VOL (CM) /5 MINUTE INCREMENT
Δ S (CM)
.32
.30
.28
.26
.24
.22
.20
.18
.16
.14
.12
.10
.08
.06
.04
.02
0
.2
.4
.6
.8
1.0
1.2
1.4
1.6
1.8
2.0
2.2
2.4
2.6
2.8
3.0
3.2
3.4
1
2
3


SILTY CLAY (SCS D)
K_s = 0.002 cm/min = 0.047 in/hr
porosity (θ_s) = 0.48 cm/cm
field capacity (θ_fc) = 0.37 cm/cm
wilting point (θ_wp) = 0.26 cm/cm
CURVE ANTECEDENT VOLUMETRIC MOISTURE
1 0.34
2 0.40
3 0.46
ΔI, INFILTRATION VOL (CM) / 15 MINUTE INCREMENT
ΔS(CM)
0 .2 .4 .6 .8 1.0 1.2 1.4 1.6 1.8 2.0
0 0.1 0.2 0.3 0.4 0.5 0.6 0.7 0.7
1
2
3



CURVE	ANTECEDENT MOISTURE
1	0.34
2	0.40
3	0.46

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APPENDIX B:

FORTRAN PROGRAM USING TABULAR VALUES OF I - S CURVES

```
      DIMENSION RMOIST(6),R(40),RNEW(40),CUMQ(40),T(40),Q(40)
      REAL KSAT
      COMMON/FUNC/NPTS,DINF(6,40),DSTO(6,40),STOR(40)
C
C READ SOIL MOISTURE DATA
C     THI   INITIAL MOISTURE CONDITION CM/CM
C     THS   SATURATED MOISTURE CONTENT (POROSITY) CM/CM
C     DEPTH   DEPTH OF AVAILABLE STORAGE  INCHES
C     KSAT    SATURATED CONDUCTIVITY CM/MIN
C
      READ(5,1000) THI,THS, DEPTH,KSAT
C
C READ INFILTRATION-STORAGE RELATIONSHIPS
C
C     NCUR       NO. OF CURVES PROVIDED
C     NPTS       NO. OF VALUES DESCRIBING EACH CURVE (TABULAR)
C     DELT       TIME INCREMENT USED FOR SOLUTION (HOURS)
C     RMOIST(I)      INITIAL MOISTURE CONDITION CM/CM FOR CURVE I
C     DINF(I,J)      INFILTRATION VOLUME PER PERIOD - JTH POINT ON CURVE I
C     DSTO(I,J)      CHANGE IN INITIAL STORAGE - JTH POINT ON CURVE I
C
      READ(5,1000) NCUR,NPTS,DELT
      READ(5,1000) (RMOIST(I),I=1,NCUR)
      READ(5,1000) ((DSTO(I,J),DINF(I,J),J=1,NPTS),I=1,NCUR)
C
C READ RAINFALL CONDITIONS
C     N      NUMBER OF RAINFALL PERIODS
C     RDELT   DURATION OF EACH RAINFALL INCREMENT  HOURS
C     R(I)    RAINFALL RATE FOR ITH INCREMENT  INCHES/HOUR
C
      READ(5,1000) N,RDELT
      READ(5,1000) (R(I),I=1,N)
C
C
C CONVERT RAINFALL DATA SO THAT TIME INCREMENTS EQUAL DELT
C
      IF(RDELT.EQ.DELT) GO TO 100
      DIV=RDELT/DELT
      NDIV=DIV
      RN=N
      RN=DIV*RN
      N=RN
C
      DO 10 I=1,N
      DO 11 II=1,NDIV
      RNEW(II)=R(I)
11    CONTINUE
10    CONTINUE
C
      DO 12 I=1,N
      R(I)=RNEW(I)
12    CONTINUE
C
100   DO 14 I=1,N
      R(I)=R(I)*2.54*DELT
14    CONTINUE
C
C CALCULATE MXIMUM AVAILABLEE STORAGE DURING INFILTRATION EVENT
C
      SMAX=(THS-THI)*DEPTH*2.54
C
C DETERMIN CURVES TO BE USED FOR INTERPOLATION PROCEDURE
C
      DO 450 I=1,NCUR
      IF(THI.GT.RMOIST(I)) GO TO 400
      ICURV2=I
      ICURV1=I-1
      RI=I
      IF(I.EQ.1)GO TO 401
      RATIO=(RMOIST(ICURV2)-THI)/(RMOIST(ICURV2)-RMOIST(ICURV1))
      GO TO 450
401   RATIO=0.
400   IF(I.LT.NCUR) GO TO 450
      ICURV2=I
```

```
      ICURV1=0
      RATIO=0.
450   CONTINUE
C
C CALCULATE RAINFLL EXCESS FOR EACH TIME INCREMENT
C
      STOR(1)=0.
      CUMQ(1)=0.
      T(1)=0.
C
      DO 500 IT=2,N+1
      IF(STOR(IT).GT.SMAX) GO TO 502
C
C CALL SUBROUTINE TO CALCULATE INFILTRATION CAPACITY FOR
C   EXISTING SOIL STORAGE
C
C     THI   ANTECEDENT MOISTURE CM/CM
C     STOR    DEPLETED STORAGE FROM INITAIL CONDITIONS  CM
C     ICURV2  FUNCTION FOR MOISTURE CONDITION CLOSEST TO THI  ( > THI)
C     ICURV1  FUNCTION FOR MOISTURE CONDITION CLOSET TO THI   ( < THI)
C     RATIO   INTERPOLATION FACTOR IN USING BOTH FUNCTIONS
C     DELI    INFILTRATION VOLUME PER TIME PERIOD  CM
C
      CALL INTERP(THI,IT,ICURV2,ICURV1,RATIO,DELI)
C
      GO TO 503
502   DELI=KSAT*DELT*60.
      GO TO 504
503   IF(DELI.GT.R(IT-1)) DELI=R(IT-1)
504   STOR(IT)=STOR(IT-1)+DELI
      WRITE(7,2000) STOR(IT-1),STOR(IT),DELI
2000  FORMAT(3F8.3)
      Q(IT-1)=(R(IT-1)-DELI)/2.54
      CUMQ(IT)=CUMQ(IT-1)+Q(IT-1)
      T(IT)=T(IT-1)+DELT
500   CONTINUE
      DO 700 IS=1,N
      R(IS)=R(IS)/(DELT*2.54)
      STOR(IS)=STOR(IS)/2.54
700   CONTINUE
      RNEW(1)=R(1)*DELT
      DO 701 IS=2,N
      RNEW(IS)=RNEW(IS)+R(IS)*DELT
701   CONTINUE
C
      WRITE(6,799)
      WRITE(6,800) (T(I),R(I),RNEW(I),STOR(I),Q(I),CUMQ(I),I=1,N)
C
1000  FORMAT()
799   FORMAT(//1X,'TIME(HRS)',5X,'RAINFALL INTENSITY(IN/HR)',
     * 5X,'RAINFALL VOL(IN)',5X,'INFILTRATION VOL(IN)',5X,
     * 'RAINFALL EXCESS(IN)',5X,'CUMULATIVE RUNOFF(IN)')
800   FORMAT(6(F10.3,10X))
      STOP
      END
```

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```
,S STORFIL.INTERP
10R1    03/15/82-10:5 INTERP(13,)
      SUBROUTINE INTERP(THI,IT,ICURV2,ICURV1,RATIO,DELI)
      COMMON/FUNC/NPTS,DINF(6,40),DSTO(6,40),STOR(40)
      DIMENSION DEL(2),IC(2)
C
      IC(2)=ICURV2
      IC(1)=ICURV1
C
      DO 5000 I=1,2
      IF(IC(I).EQ.0) GO TO 5000
C
      DO 5001 J=2,NPTS
      IF(STOR(IT-1).GT.DSTO(IC(I),J)) GO TO 5002
      IF(STOR(IT-1).EQ.DSTO(IC(I),J)) GO TO 5003
      DEL(I)=DINF(IC(I),J-1)+(STOR(IT-1)-DSTO(IC(I),J-1))*
     *   (DINF(IC(I),J)-DINF(IC(I),J-1))/(DSTO(IC(I),J)-DSTO(IC(I),J-1))
```

```
              WRITE(7,1001) DEL(I),DSTO(IC(I),J-1),DSTO(IC(I),J),
             * DINF(IC(I),J-1),DINF(IC(I),J)
      1001   FORMAT(5F8.3)
             GO TO 5000
      5003   DEL(I)=DINF(IC(I),J)
             GO TO 5000
      5002   IF(J.EQ.NPTS) DEL(I)=DINF(IC(I),J)
      5001   CONTINUE
      5000   CONTINUE
             IF(ICURV1.EQ.0) GO TO 5100
             DELI=DEL(2)-RATIO*(DEL(2)-DEL(1))
             GO TO 5120
      5100   DELI=DEL(2)
      5120   WRITE(7,1000) DELI,DEL(1),DEL(2),RATIO
      1000   FORMAT(4F10.3)
      5121   RETURN
             END
RNING 2608 DUMMY ARGUMENT 'THI' IS NEVER REFERENCED

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