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Planning Committee
L. R. Ahuja, GPSR, Fort Collins, CO (coordinator)
David Bosch, SEWRL, Tifton, GA
Gerald Flerchinger, NWWRC, Boise, ID
Gary Frasier, RRR, Fort Collins, CO
Sally Logsdon, NSTL, Ames, IA
Fred Pierson, NWWRC, Boise, ID
Walter Rawls, HL, Beltsville, MD
Roger Smith, WMR, Fort Collins, CO

Compiled and edited by:
L. R. Ahuja and Amy Garrison
USDA-ARS-GPSR
Fort Collins, CO

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For further technical information:

USDA-ARS-NPA
Great Plains Systems Research
P. O. Box E
Fort Collins, CO 80522

Preface

The science of hydrology embraces the process of infiltration which determines the movement of water into and within the shallow soil mantle that covers the terrestrial surface of our planet. The infiltration process has attracted the scientific attention of geophysicists, hydrologists, and soil scientists because of the vital role that water plays in our biological world. Considerable progress has been made in identifying and defining the physical laws that govern the movement of water within soils and other porous materials, and in characterizing the hydraulic properties of these materials. However, many major scientific challenges, related to the effects of spatial and temporal changes in these hydraulic properties on the movement of water, have defied the efforts of some of the most brilliant minds in the international scientific community for decades. Increased public awareness and concern about the potentially adverse long-term economic, health and environmental impacts of many agricultural and industrial chemicals has provided the stimulus for improving our knowledge and understanding of the infiltration process. An improved understanding of the infiltration process at field and larger scales, which encompass large spatial and temporal variabilities due to both natural causes and management effects, will provide society with the knowledge and information needed to make optimal decisions on chemical use and resource management. The challenges are great, the benefits substantial.

This workshop was organized by ARS to review the state-of-science and identify critical further research needs to address the above challenges. The outcomes of this workshop documented in these Proceedings will serve as a framework for the ARS's strategic plan for research on infiltration in the near future.

David A. Farrell
National Program Leader, Hydrology
USDA-ARS, Beltsville, MD

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Executive Summary

In today's agriculture, infiltration of water into soils is a fundamental hydrological process that plays a dominate role in almost all important natural resource problems or issues, such as runoff, erosion, irrigation, water conservation, groundwater and surface water quality, and global environmental change. Research during the last half century has established a fundamental theory of infiltration into soils at a given point. Attempts to apply this theory to infiltration at a field scale have had only a limited success, mainly due to the large spatial and temporal variability of the governing parameters that exist at field scale. Lumped parameter models have failed to provide the accuracy we need. The research on spatial variability during the last quarter century has enlightened us about the nature of this variability, but the application of this knowledge to develop a field-scale infiltration theory is still in its infancy. Major hurdles are in approximating, measuring, and quantifying the large and complex patterns of spatial and temporal variabilities, and implementing these into a model for field-scale infiltration. Also, very little research has been conducted on characterizing the effects of land use and management practices on the temporal and spatial variabilities; these effects are often more important than the natural variabilities in a given area. These major issues call for new innovative methodologies in characterizing variabilities, as well as new scale-dependent infiltration theories. At the suggestion of the ARS National Program Staff, the ARS Workshop on "Real World" Infiltration was held at Pingree Park, Colorado, July 22-25, 1996, to address the above issues by reviewing current ARS research and identifying future research needs. Planning for the Workshop started a year earlier. A notice for the Workshop was sent to all ARS research units dealing with natural resources, the National Program Staff, and several cooperating agencies. Thirty-five scientists attended the Workshop, including an invited speaker and two NRCS scientists.

The vision of the Planning Committee was to have a strategic plan for a coordinated ARS infiltration research effort that will accommodate large spatial and temporal variabilities of relevant soil properties at different scales. The following Specific Goals were delineated:

1) Review and analyze the knowledge gained from the past experimental studies on infiltration at plot, field, and small watershed scales, separately for crop- and rangelands. Assess how parameter variabilities were determined and incorporated in modeling.

2) Analyze old and explore new concepts for identifying and quantifying spatial variability and patterns of relevant soil infiltration parameters of crop- and rangelands, and practically feasible experimental methods of determining these variabilities at different scales.

3) Analyze, identify, and quantify temporal variability of relevant soil infiltration parameters of crop- and rangelands, and practically feasible experimental methods of determining these variabilities at different scales.

4) Explore possibilities of new scale-dependent infiltration equations that may indirectly account for large variabilities in relevant infiltration parameters, as well as a framework for incorporating spatial and temporal variabilities in existing equations and the use of GIS technology.

Teams were organized before the Workshop to prepare written reports on the state-of-the-science and future research needs pertaining to the above goals. These reports were consolidated and sent to all participants before the Workshop.

The first day of the Workshop was devoted to giving of a charge to the participants by the National Program Leaders, followed by presentations by the Specific Goal teams of the state-of-the-science and future research needs identified. Feedback was encouraged after each team presentation. In the evening, Dr. Vijay Gupta, University of Colorado, Boulder, gave an invited talk on the potential use of Cascade Theory in simple- and multi-fractal scaling of infiltration for different spatial scales.

On the second day, eight Cross-Goal Interest groups discussed, modified, added, and ranked the research needs identified by the Specific Goal teams relevant to the subject assigned to each group. The eight Cross-Goal Interest Groups were:

Group 1: Mechanical/management/tillage effects - process knowledge, space-time-causal factor relationships, modeling . . .
Group 2: Experimental methods/measurements for field-scale quantification of spatial and temporal variabilities
Group 3: Biological effects --roots, canopy, worms, . . .process knowledge, space-time-causal factor relationships, modeling, . . .
Group 4: Physical effects--freezing-thawing, hydrophobicity, swelling-shrinking, . . .
Group 5: Parameterization/estimation
Group 6: Stochastic statistical characterizations and modeling
Group 7: Spatial characterization needs for precision farming
Group 8: Special topics: Minimum data set, methodologies, amending infiltration, and some general items

In the afternoon, the group leaders presented their group's conclusions to the entire workshop. In the evening, the two National Program Leaders and the two NRCS cooperators gave their opinion on the prioritized research needs. On the third day, the participants as a whole suggested to National Program Leaders that they appoint a small committee to take the prioritized list of research needs and develop them into a strategic plan for a coordinated ARS research on infiltration. The participants also decided that the proceedings of this Workshop should be published soon.

Prioritized research needs identified by each of the eight Cross-Goal Interest groups are attached as an Appendix. The twelve highest ranked needs overall, not in any order of priority, are as follows:

- Develop consolidated database of existing plot and watershed infiltration data. Similarities and differences in data sets need to be identified and documented. Need one FTP site for all data, in standardized format and with information such as georeference, size, soil, methods, etc.

- A new book on standardized methods of measurement and analysis; revise Hydrology Handbook and include a data collection protocol - minimum data set, appropriate methodology, and data format.

- Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.

- Incorporate terrain attributes into relationships for surface soil attributes, soil hydraulic properties and soil textures to obtain regional-scale prediction equations.

- Test multifractal techniques for modeling spatial dependence of properties/processes including methods to incorporate parameter variability into parameter estimation.

- Develop improved predictive models which include:
    - b) seasonally varying infiltration rates due to plant growth and worm activity, shrink/swell;
    - c) systematic small-scale spatial variability such as crop-row position effects.

- Conduct a comprehensive review and evaluation of existing data and creation of new data to quantify/estimate the temporal/spatial character of seal/crust as influenced by distributed residue cover, soil type, wetting/drying, and crops.

- Measure and quantify spatial variability of infiltration parameters between plant bases and interspaces, as a function of soil type, grazing intensity, and other factors.

- Quantify the changes in infiltration behavior from changes in soil parameters as a result of mechanical modifications, including modeling the aggregate behavior of an area containing internal infiltration variability, with application to "management" modeling.

- Improve our knowledge of disaggregation statistics and rainfall intensity distributions.

- At larger scales (e.g., 10 ha +), modeling a really variable infiltration should not be done independently of the surface runoff, which has considerable organized and random heterogeneity itself, nor should it be modeled without consideration of small-scale rainfall rate heterogeneities.

- Describe the mean and variance of soil water stored in the root zone as a function of size of the area to help in decision support for precision farming.

The Planning Committee

Specific Goal 1

Knowledge Gained from Past Plot, Field and Small Watershed Studies

Infiltration and Runoff Plot Studies on Rangelands: Rainfall Simulator Experiments

Jeffry Stone and Ginger Paige[1]

Abstract

Plot studies on rangelands have been used since the 1930's to investigate fundamentals of the rainfall/runoff/erosion process and the impacts of grazing management and land characteristics on these processes. The vast majority of studies have reported final infiltration rates on small plots as affected by either grazing intensity or vegetation and soil surface characteristics. Most of the studies are consistent with each other on a qualitative basis. That is, interspace areas have lower infiltration rates than under canopy areas and infiltration rates under high intensity grazing are lower than under low intensity grazing. Ascribing reasons for these differences is more problematic and the studies are less consistent. For example, although many studies have found positive correlations between final rates and increasing vegetation and litter cover, some have not. Parameterization of infiltration models has also been less than successful. The reasons for the lack of success are probably because infiltration is not measured directly and none of the experiments were expressly designed to measure infiltration parameters. Considerations for future research include defining rainfall characteristics in the Western U.S., determining a correspondence between natural and simulated plot response, determining a correspondence among simulators at point, small plot, and large plot scales, quantifying partial area response, conducting interior plot measurements, and examining all components of the hydrograph.

Introduction

Plot studies on rangelands have been used since the 1930's to investigate fundamentals of the rainfall/runoff/erosion process and the impacts of grazing management and land characteristics on these processes. Although some of the studies have monitored plot response under natural rainfall, the vast majority have used some variation of an infiltrometer or rainfall simulator. Rainfall simulation provides a relatively easy and economical way of obtaining a large amount of data under controlled conditions in a short period of time. In addition, controlled application rates allow for the comparison of steady state infiltration response to alternative management systems, to differences in vegetation and soil characteristics, and facilitates model parameter identification. The two basic components of a rainfall simulator experiment are the type of rainfall simulator used and the experimental design. The types of rainfall simulators used for rangeland infiltration experiments have been drop formers on small plots and stationary and intermittent sprinklers on both small and large plots. The experimental design generally has consisted of an initial wetting of the plot and a data run the following day with the primary data collected being the runoff at the end of the plot.

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[1]USDA-ARS, Southwest Watershed Research Center, 2000 E. Allen Rd., Tucson, AZ 85719

Problem Statement

The key factors of infiltration rainfall simulator experiments on rangelands have been; 1. measuring the process, 2. relating infiltration rates to land use or vegetation/soil characteristics, 3. identifying infiltration parameters for specific infiltration models, and 4. estimating those parameters from easily measured soil and/or vegetation characteristics. The impact of these factors varies with plot size and thereby influence the manner in which data are obtained, reduced, and presented.

Small plot experiments have the advantages of minimizing the natural heterogeneity of rangeland vegetation, soil, and surface characteristics, reaching steady state relatively quickly, and being easy to implement. The disadvantages are that the absolute amount of runoff during the early portion of the simulated event is small so that measurement error becomes a factor and there can be a fair amount of variability among plots.

Large plot experiments have the advantages that the heterogeneity is averaged and the runoff hydrograph is better defined because the absolute rates are higher. However, spatial variability effects are harder to measure directly and the cost and effort are greater than with the small plot experiments. In addition, given the spatial variability, it is uncertain if the entire plot is contributing to runoff if only one application intensity is used.

Experiments at both scales do not measure the infiltration rate directly but compute it from the runoff rate. On small plots, generally infiltration rates are computed as the rainfall rate minus the runoff rate. On large plots, an infiltration model is needed to compute the infiltration rate which presents an inverse problem where the runoff hydrograph has to be de-routed in order to compute the rainfall excess rate and then infiltration.

Rainfall Simulator Types

Rainfall simulators are designed to apply a controlled amount and rate of water on a plot. Among the characteristics which are important in the design of simulators are 1. drop size and impact velocities near that of natural rainfall, 2. drop characteristics and intensity uniform over the plot, 3. application on a plot size sufficient to represent the conditions to be evaluated, 4. application rate representative of the storm size of interest, 5. impact angle near vertical, and 6. portability, ease of operation, and cost (Meyer, 1979). The selection of which type of simulator to use is a tradeoff among the factors listed above.

Rainfall simulators can be classified either as drop formers or sprinklers (Mutchler and Hermsmeier, 1965). Drop formers apply a uniform drop size over a small area. Sprinklers apply a distribution of drop sizes over an area limited only by the number of sprinkler heads used. Bubenzer (1979) classified sprinkler simulators as continuous or intermittent application and as stationary, lateral, oscillating, or rotating. Most rangeland infiltration studies on small plots have used either drop formers or stationary sprinkler simulators. For large plot studies, stationary or rotating boom simulators have been predominantly used. A list of some of the simulators and their characteristics are presented in Tables 1 and 2.

Table 1. Drop former simulator characteristics

Name	Intensity mm/hr	Plot size m	Comments
Mobile Infiltrometer (Barnes and Costel, 1957)	25-152	0.36	
Double-Tower (Rhodes, 1961)	25-150	1<	
Drop former (Meeuwig, 1969)		1	40% natural kinetic energy
Mobile Infiltrometer (Blackburn et al., 1974)	5-200	1	
Portable Simulator (Malkuti and Gifford, 1978)	28-250	0.36	40% natural kinetic energy

Rotating Boom Simulator

Since the 1980's, rotating boom experiments of relatively uniform experimental design have been conducted by the United States Department of Agriculture-Agricultural Research Service (USDA-ARS) at many rangeland sites. These have included the Reynolds Creek Experimental Watershed, ID; Walnut Gulch and Santa Rita Experimental Watersheds, AZ; over 50 rangeland sites as part of the rangeland Water Erosion Prediction Project (WEPP) and the Interagency Rangeland Water Erosion Team (IRWET) field programs; Department of Energy sites Los Alamos National Laboratory, NM, Nevada Test Site, NV, and Idaho National Laboratory, ID; US Forest Service sites in Arizona and California, and several sites in northern Mexico. As such, the rotating boom experiment represents one of the largest rainfall simulator data bases available, much of it in electronic format. The rotating boom nozzle, the VeeJet 80100, was first used by Meyer and McCuen (1958) on a Purdue simulator. Meyer and Harmon (1979) compared the impact energy, terminal velocity, and raindrop distribution of the VeeJet 8070, 80100, and 80150 nozzles with characteristics of natural rainfall. They obtained a good correspondence between the 80100 and 80150 nozzles and natural rainfall energy data from northern Mississippi (McGregor and Mutchler, 1977) and with drop size data taken by Carter et al. (1974) in the south United States for storms of intensities of 2-13 and 26-51 mm/hr respectively.

Table 2. Sprinkler simulator characteristics

Name	Type	Intensity mm/hr	Plot Size m	Comments
Type D (Beutner et al., 1940)	Continuous Stationary		3 x 4	
Type F (Kincaid et al., 1964)	Continuous Stationary		3 x 4	low intensity
Modified Purdue (Bertrand and Parr, 1961)	Continuous Stationary	64 -115	1	Spraying Systems Full Jet 5b, 5d, 7LA
Mobile Infiltrometer (Rauzi and Smith, 1973)	Continuous Stationary		1	
Rocky Mountain Infiltrometer (Dortignac, 1951)	Continuous Stationary	127	variable	Type F
Wilcox (Wilcox et al., 1988)	Continuous Stationary	70-200	1	Spraying Systems Full Jet 1/4G10
USGS (Lusby, 1977)	Continuous Stationary	50		Rainjet 78C
Purdue Oscillating (Meyer and Harmon, 1979)	Intermittent Oscillating	10-140	1	VeeJet 80100, 80150
Rotating Boom (Simanton and Renard, 1982; Simanton et al., 1991)	Intermittent Rotating	60,120, 180	3 x 10	VeeJet 80100

Rainfall Simulator Comparisons

Very little research has been conducted to compare plot response due to the simulator type thus making it difficult to draw other than qualitative conclusions among the various studies. One of the first comparisons of simulators was done by Kelly (1940) who compared the Type F, a stationary sprinkler simulator, with an oscillating intermittent Colorado simulator and found that the Type F gave final infiltration rates which were 20-30% lower. Rhodes et al. (1964) compared final intake rates measured using a tower drop former (Rhodes, 1961) with a double ring infiltrometer and found that the ring infiltrometer gave rates that were 2 to 3 times higher. Aboulabbes et al (1985) found similar results when comparing sprinkler infiltrometers and ring infiltrometers. On croplands, Neibling et al. (1981) found slightly less soil loss and runoff when a shorter oscillation time was used on a Purdue type simulator. Ward (1986) compared the Purdue small area simulator (1 m^2) with a modified Colorado standpipe sprinkler large area simulator (2000 ft^2) and found that both gave similar final infiltration rates at three former ARS watersheds in New Mexico.

Natural runoff events rarely reach steady state so direct comparisons of infiltration rates with those of simulated rainfall are difficult. Kincaid et al. (1964) computed steady state infiltration using the Type F infiltrometer on 1.8 x 3.7 m^2 plots at Walnut Gulch and Kincaid and Williams (1966) measured natural rainfall-runoff on the same plots. Infiltration rates were not computed for the natural events because the steady state assumption was not met. Lusby and Lichty (1983) compared optimized Green-Ampt Mein-Larson (GAML) infiltration model (Mein and Larson, 1973) saturated hydraulic conductivity obtained from rainfall simulation with those computed from naturally occurring rainfall on Willow Gulch near Denver CO. They found that the conductivity term for the observed data was lower than that obtained from rainfall simulation at two upland ponderosa sites but was about the same for two low land grass sites.

Experimental Designs

The important factors in a rainfall simulator experimental design are 1. water application characteristics (rate, duration, number of runs), 2. plot size, 3. treatments, and 4. measurements. Tables 3 and 4 list design characteristics of selected small and large plot rangeland experiments.

Table 3. Small plot rainfall simulator plot experimental designs

Simulator Type	Intensity (mm/hr)	Number of Runs	Measurements	Treatments	Reference
Blackburn drop former	208	pre-wet, wet	30 minute sample	grazing	McCalla et al., 1984
	153	-do-	5 minute grab sample	grazing	McGinty et al., 1978
	140	-do-	average of 28 and 30 minute samples	canopy interspace	Blackburn et al., 1990
Malekuti	64, 127	pre-wet, wet, very wet	variable	grazing	Devaurs and Gifford, 1986
Meyer and Harmon	110	pre-wet, wet	30 minute average	vegetation	Pluhar et al., 1987
Wilcox	103	pre-wet, wet	5 minute grab samples	vegetation	Wilcox et al., 1986

Table 4. Large plot rainfall simulator plot experimental designs

Simulator Type	Intensity (mm/hr)	Number of Runs	Measurements	Treatments	Reference
Type D1	76-86	antecedent, wet	variable	natural	Beutner et al., 1940 Sharp and Holtan, 1939
Type F	102	antecedent	variable	natural	Kincaid et al., 1964
Rotating Boom	60	dry, wet, very wet	flume with FW1 recorder	natural, clipped, bared, tilled	Simanton and Renard, 1982
USGS				natural	Lusby and Litchy, 1983
Rotating Boom	60, 120	dry, wet, very wet	flume with bubble gage	natural, clipped, bared two small plots bared covered and uncovered	Simanton et al., 1991

Water application characteristics

The water application rate for sprinkler simulators typically depends on the nozzle type, operating pressure, number of sprinklers, and the duration of intermittent application. Drop formers generally can vary the application rate through a wide range through the use of flow meters. Sprinklers vary application rate through pressure, number of nozzles, or by intermittent application on the plot. Typical application intensities have varied between 60 - 150 mm/hr although some studies have used higher rates (Blackburn et al., 1980). Some studies have attempted to simulate a specified return period storm. Most of the studies have used a single application rate. However, it has been observed that the apparent infiltration rate increases as the rainfall application rate increases (Cook, 1946; Moldenhauer et al., 1960; Hawkins, 1982; Dunne et al., 1991). The proposed explanation is that as the application rate increases, portions of the plot which have a high infiltration capacity begin to contribute to runoff and thus the apparent infiltration rate becomes greater. This implies that for the typical single application rate of simulator experiments, only a portion of the plot may be contributing to runoff. Sharp and Holtan (1940), Lusby and Litchy (1983), and the WEPP rangeland field experiments (Simanton et al., 1991) used multiple application rates within the same run but did not attempt to quantify the relationship between intensity and infiltration capacity.

The duration of the application is most often fixed; common durations have been 30 and

60 minutes which has been generally sufficient for the runoff to reach steady state. A few studies were run until the runoff rate was constant.

Most of the experiments have used a simulator run at a single application rate for one hour one day after a pre-data run. The pre-data run is used to obtain uniform moisture on the plot. The WEPP experiments (Simanton et al., 1991) used three runs; a dry run at a single intensity followed 24 hours later by a wet run at a single intensity followed 30 minutes later by a very wet run at two intensities.

Plot size

Plot sizes have varied as indicated in Tables 1 and 2. The most common plot sizes have been on the order of a square meter used in many grazing and canopy/interspace studies, 1.8 x 3.7 m used with the Type F infiltrometer, and 3 x 10 m used with the WEPP experiments.

Treatments

The majority of small plot experiments used to evaluate grazing management have not treated the plot, but have placed the plot within a treated area of study. Most of the grazing studies have used a control or no grazing treatment and ranges of grazing intensity. The standard WEPP experiment had three treatments on 3x10 m plots, natural, clipped, and bared. The clipped and bared treatments are to isolate the effects of vegetative cover and ground cover respectively on runoff and erosion. Two additional treatments were used on 1x.6 m plots, bared and bared/covered. The small bared plot and covered bare plot were intended to separate the effects of soil crusting on infiltration in addition to identify the WEPP interrill erosion parameter.

Measurements

As mentioned in the introduction, the infiltration rate is not measured on the plot scale but is computed from the runoff hydrograph either as a steady state infiltration rate or computed from an optimized infiltration equation. Ancillary measurements include initial soil moisture conditions, vegetation and soil surface characteristics, and microtopography. These are used either in developing regression equations relating final infiltration rates and plot characteristics or in parameter estimation of infiltration and runoff models.

Most of the experiments compute the runoff rate by taking a volume of water as a grab sample and dividing it by the sample time duration. The sample is taken either at the end of the plot or pumped or gravity fed to a tank. A common sample interval for small plot experiments is every 5 minutes. Beutner et al. (1940) and Kincaid et al. (1964) took grab sample on large plot studies and took more samples during the rise of the hydrograph than during the steady state period. Flumes either coupled with water level recorders or bubble gages have been used on large plot studies by Lusby and Litchy (1983) and by the WEPP rangeland experiment (Simanton et al., 1991).

Plot characteristics such as soil surface and vegetation properties have been measured primarily to either develop regression relationships between final rates of infiltration and plot characteristics or to develop parameter estimation equations for infiltration models such as the GAML model (Alberts et al., 1995) for WEPP. Initial soil moisture conditions have been measured generally by gravimetric samples, canopy and surface cover characteristics have been measured by point frames (i.e. Wood et al., 1986), the line intercept method (Gamougoun et al., 1984), and the gridded sampling quadrat method (Blackburn et al., 1980).

Almost all the experiments can be classified as steady state and lumped because 1. the infiltration rate is assumed to be equal to the difference between the application rate and steady state runoff rate and 2. the runoff rate is measured only at the end of the plot. Inherent in all the experiments is the assumption that at steady state, the entire plot is contributing to runoff.

Results

Results of simulator experiments on rangelands have been reported as 1. final rates as influenced by grazing intensity, vegetation characteristics, and chemical or mechanical treatment of vegetation, 2. regression relationships between final rates and vegetation/soil characteristics and 3. parameter estimation of infiltration equations.

Grazing effects

Gifford and Hawkins (1978) reviewed the rangeland infiltration literature of studies conducted to examine the impact of grazing on infiltration. Grazing treatments were classified into ungrazed, light, moderate, or heavy grazing, and good, fair, or poor range condition. The initial soil moisture conditions ranged from dry to field capacity. The infiltration rates presented range from averages over the entire run to final rates. Gifford and Hawkins (1978) concluded that there was a significant difference in final infiltration rates between ungrazed and grazed and between heavy and moderate/light grazing but that it was difficult to determine differences between moderate and light grazing. Studies carried out after Gifford and Hawkin's review are consistent their report. Statistically significant differences between final infiltration rates for heavy and light to moderate grazing have been reported on the Texas Experimental Station (McGinty et al., 1978; Blackburn et al., 1980; McCalla et al., 1984; Knight et al., 1984), Fort Stanton, NM (Gamougoun et al., 1984; Weltz and Wood, 1986), Utah (Merzougui and Gifford, 1987; Devaurs and Gifford, 1986) and Arizona (Tromble et al., 1974).

Cover Effects

Grazing intensity and range condition are both qualitative terms so that many studies beginning in the late 1970's began to attempt to quantify the effects of grazing on measurable characteristics such as vegetation and soil surface cover. Positive and negative correlations between final infiltration rate and vegetation and soil characteristics found in the literature are listed in Table 5.

Vegetation: Among the variables listed in Table 5, the effect of vegetation has been questioned. Weltz and Wood (1986) and Wood et al. (1987) found positive correlations with total above ground biomass, grass standing crop, and litter accumulation. However, Johnson and Niederhof (1941) and Marston (1952) found no strong relationship between vegetative cover and infiltration. Smith and Leopold (1942) and Dortignac and Love (1966) found large changes in infiltration with only small changes in vegetation density. Busby and Gifford (1981) found that clipping crested wheatgrass and compacting the soil in southeastern Utah had no significant effects on infiltration. They concluded that because the cover was less than 50% and the clipping did not reflect long term conditions, that there was no impact. They also found no single set of variables which could explain differences in infiltration across all treatments. Johnson and Blackburn (1989) reported an 18% increase in runoff for clipped plots on only the very wet run on sagebrush sites in Utah, while Simanton et al. (1991) found that clipping grass canopy cover had no significant effect on final infiltration rate or the runoff/rainfall depth ratio using the WEPP data. Kincaid et al. (1964) found a non-linear relationship between increasing canopy cover and increasing infiltration rates but that below a certain percent cover that there was no relationship on a brush dominated site at Walnut Gulch. Lane et al. (1987) found significant positive correlations between final infiltration rates and vegetative and ground cover on large plots in Arizona and Nevada. Bolton et al. (1990) found that on the Jornada Range in New Mexico, vegetation did not affect runoff depth on 4 m^2 natural rainfall plots but had a significant effect on 1 m^2 rainfall simulator plots. Dunne et al. (1991) examined the vegetative cover effects on final infiltration rates in Kenya using a sprinkler rainfall simulator. They found little influence of vegetative cover postulating that the root system had more of an influence on infiltration rates. They also found a relationship between application rate and apparent final infiltration rate that was independent of the percent vegetative cover.

Table 5. Correlations between final infiltration rate and vegetation and soil characteristics

Positive	Negative
Canopy cover	Clay
Live biomass	Gravel cover
Litter cover	Rock cover
Basal cover	Bulk density
Soil organic carbon	Surface horizon structure
Roughness coefficient	Bare ground
Porosity	
Number of depressions	
Total ground cover	

Canopy Interspace: On many rangelands, there are discrete areas of shrubs or trees and interspaces without vegetation. Blackburn et al. (1975) found a significant difference between 30 minute infiltration rates for the coppice dune and the interspace areas. Infiltration rates were positively correlated with the extent and surface morphology of dune interspace areas and negatively correlated with vesicular horizons. The negative correlation with bare ground compared with results of Duley and Domingo (1949) and Branson and Owen (1970). The positive correlation of plant and litter cover was not as strong as reported by Dortignac and Love (1966), Rauzi et al. (1968), and Meeuwig (1969). Balliette et al. (1986) found that average final infiltration rates were greater under sagebrush canopy than in the interspace areas. Rostagno (1989) found that eroded shrub interspace areas had lower infiltration rates than non-eroded in northeastern Patagonia, Argentina. Blackburn et al. (1990) studied the temporal and spatial variation of infiltration under and outside of sagebrush canopy from February to May at Reynolds Creek. Interspace rates were significantly lower than the canopy areas and the February-March rates were significantly lower that the remainder of the simulations. Tromble (1980) studied the effects of rootplowing creosote dominated rangeland on the Jornada Experimental Range in New Mexico and found that final infiltration rates were greater on creosote plots than the plowed plots. Johnson and Gordon (1984) found that the interspace area produced 2.5 times the runoff as the under sagebrush canopy area.

Rock Effects: Poesen et al. (1990) contrasted the results of authors who reported positive (Tromble et al. 1974, Blackburn et al. 1975) with those who reported negative (Kincaid et al., 1966; and Tromble, 1976) effects of rock fragments on the soil surface with the amount of runoff volume on small rainfall simulator plots. They postulated that imbedded rock fragments increase runoff while if they lay on the soil surface they decrease runoff volume.

Rangeland Treatments: Brock et al. (1982) examined the effects of herbicides and rootplowing for brush control on infiltration in north central Texas. They found that regardless of treatment, final infiltration rates were higher within the canopy than within the interspace areas but that there was no significant difference between the control and the treatments. Bedunah and Sosebee (1985) found that the vibratill and shred treatments significantly increased the infiltration rate while all other brush control treatments were not significantly different from the control. Contrary to studies by Knight et al. (1984) and Brock et al. (1982), they found no difference between the infiltration rates under and outside mesquite canopies. Wood et al. (1986) found that final infiltration rates on fertilized and unfertilized pasture were different than the control. Knight et al. (1986) studied oak mottes on the Edwards Plateau in Texas and found that 30 minute infiltration rates were higher for undisturbed conditions than for areas where mulch and organic layers were removed.

Infiltration Parameters

Small Plot Experiments

Sabol et al. (1982) used the modified Purdue simulator to develop runoff ratios, Curve Numbers, and GAML parameters for 10 sites, developed and undeveloped, in the Albuquerque area. They used a least squares procedure to obtain the conductivity and capillary terms of the GAML model and obtained values which were in the same range as Rawls et al. (1982).

Devaurs and Gifford (1986) used a modular drop forming device on .37 m^2 plots located within 3 m x 10.5 m plot which were used with the rotating boom rainfall simulator on Reynolds

Creek. The treatments were grazed, ungrazed, and tilled. They compared GAML parameters obtained by a least squares fitting of the data to those computed from soil texture (McCuen et al., 1981). The data were fitted by plotting the infiltration rate versus the reciprocal of the cumulative infiltration depth so that the intercept is the GAML conductivity term and the slope is the hydraulic conductivity times the effective matric potential. This method gave negative values of the GAML parameters for some cases. The texture derived parameter values worked best for the tilled rangeland soils but did poorly for the control plots. Hutten and Gifford (1988) compared the observed infiltration rates with those predicted from soil characteristics (McCuen et al., 1981) and found that the observed rates were much higher than the soil predicted rates on native rangeland and plowed sites.

Large Plot experiments

Large plot experiments differ not only in the plot size but also in the information which can be obtained from the experiment. Because of the size, processes which may be negligible on a smaller scale may become significant on the larger scale. Sharp and Holtan (1940) stated

> ...only during those portions of the hydrograph when runoff is constant, and after satisfaction of depression- and surface-storage, can infiltration rates be determined directly, and with any degree of accuracy. During any period of the hydrograph when the rate of runoff is changing, three other factors, rate of infiltration, and amounts of depression- and surface-storage may or may not be changing also.

The point is that the only data available from large plots are runoff hydrographs so that to identify time varying infiltration rates or model the process, the entire hydrograph has to be analyzed. The Sharp and Holtan statement is true for small plots, but because of the small absolute amounts of runoff, it is harder to accurately define the rise and recession of the hydrograph.

Some of the first and still most complete analyses of the runoff hydrograph were performed in the late 1930's. Sharp and Holtan (1940) analyzed hydrographs for detention storage and depression storage from rainfall simulator experiments on the Concho River Watershed, TX. A graphical method was used to compute detention storage from the recession of the hydrograph and depression storage as the residual of the plot water balance at steady state. Beutner et al. (1940) computed Horton infiltration (Horton, 1939) parameters, stage-discharge relationships, and roughness coefficients similar to Manning's n for 14 sites in Arizona.

Lusby and Litchy (1983) used trial and error to fit GAML and the kinematic wave model Manning's n using simulated and natural rainfall plot data and natural rainfall watershed data in Colorado with inconsistent results. Kuczera and Patterson (1993) used a stationary rainfall simulator on 2 by 8 m plots to fit parameters for a coupled Horton-kinematic wave model. The wet run was used to fit Manning's n, the final infiltration parameter of the Horton equation, and depression storage. The dry run was used to fit the remainder of the Horton equation parameters. They obtained good fits of the hydrograph but were uncertain what the results meant in regards to the kinematic wave model.

Alberts et al. (1995) used the WEPP and IRWET rangeland data to develop optimized GAML conductivity terms and estimation procedures based on vegetation and soil characteristics.

The methodology used the WEPP model to estimate the matric term, adjusted that term with the site soil porosity and initial soil water conditions, and adjusted the conductivity until the simulated runoff volume matched the observed volume for the wet run. Multiple regression analysis was then used with the fitted conductivity terms to develop equations which predicted conductivity as a function of vegetation and soil properties.

Roughness coefficient

The roughness coefficient, expressed as Manning's n or Chezy C, has been used with the kinematic wave model to route rainfall excess. Two approaches which have been used to evaluate the roughness coefficient using rainfall simulator data are to directly measure the local flow depth or velocity on the plot and to use the hydrograph at the outlet of the plot with the kinematic wave model. Studies which have used the first approach are Emmett (1970) and Abrahams et al. (1986) who measured flow depths at regular intervals downslope on large plots (9 to 14.4 m) on natural rangeland hillslopes. Engman (1986) used the second approach which consisted of a method that minimized the difference squared between the observed and predicted hydrograph as computed by a finite difference solution of the kinematic wave model. However, Woolhiser (1975) suggested that, because the flow depth is so small on rangelands, the roughness coefficient be computed using the recession portion of the hydrograph with the equation for kinematic storage on the flow surface at steady state.

The majority of studies on rangelands have related optimized values of the roughness coefficient (generally Manning's n) to qualitative descriptions of vegetation (Emmett, 1970; Woolhiser, 1975), vegetative or surface condition (Ree et al., 1977; Abrahams et al., 1986), or simply as a broad class termed "range" (Engman, 1986). The most comprehensive list of Manning's n and Chezy C values for rangelands was compiled by Weltz et al. (1991) who used Engman's hydrograph fitting method to identify Darcy-Weisbach friction factors for natural, clipped, and bared WEPP rangeland rainfall simulator plots. They estimated the total friction factor as a summation of friction sub-factors associated with grain roughness, random roughness, ground surface cover, and canopy cover.

Summary

The vast majority of studies have reported final infiltration rates on small plots as affected by either grazing intensity or vegetation and soil surface characteristics. Most of the studies are consistent with each other on a qualitative basis. That is, interspace areas have lower infiltration rates than under canopy areas and infiltration rates under high intensity grazing are lower than under low intensity grazing. Ascribing reasons for these differences is more problematic and the studies are less consistent. For example, although many studies have found positive correlations between final rates and increasing vegetation and litter cover, some have not.

Parameterization of infiltration models, primarily the GAML model, has also been less than successful. The reasons for the lack of success are probably because none of the experiments were expressly designed to measure the GAML parameters and could be an indication that infiltration formula such as the GAML do not perform well under rainfall excess conditions.

Considerations for Future Research

1. Rainfall characteristics in the Western U.S.: In much of the western U.S., rainfall intensity is the controlling factor in the initiation and rates of runoff. However, the relationship between important natural rainfall characteristics (kinetic energy, drop size distribution) occurring in the western US and rainfall simulator characteristics has not been extensively studied. For example, Tracy et al. (1984) found that thunderstorm rainfall in southeastern Arizona had a higher kinetic energy as measured using a distrometer than the rainfall energies found by Carter et al. (1974) and McGregor and Mutchler (1977) in other parts of the country.

2. Correspondence between natural and simulated plot response: There have been very few studies relating infiltration rates or parameter estimates obtained from simulation to those obtained from natural rainfall. Several plots used for the USLE experiments at Walnut Gulch were monitored for natural rainfall (Simanton et al., 1984) for a year but were discontinued because of equipment problems. Results indicated that both runoff and sediment yield were greater for the natural events than for the rainfall simulation events. There is a potential for some comparisons with existing watershed data. Both the rangeland USLE and WEPP experiments were done at the Walnut Gulch and Reynolds Creek Experimental Watersheds. The WEPP experiment also had plots at watersheds R5 and R7 at Chikasha, OK, Cottonwood, SD, and Los Alamos National Laboratories.

3. Correspondence among simulators at point, small plot, and large plot: Point measurements, such as ring infiltrometers and disk permeameters, are popular methods of characterizing infiltration because they are easy to use, quick, and economical. Studies have shown that these methods yield higher infiltration rates than plot scale measurements using rainfall simulators. If point measurements and small and large plot simulators are going to be used in the future, there is a need to relate infiltration rates and parameter values obtained at the three scales.

4. Partial area response: At both the small and large plot scale, partial area response can be a significant process controlling the rates and amounts of infiltration. Because most experiments are run at a single intensity and the runoff is computed assuming the entire plot area is contributing, extending results to natural rainfall is difficult. Experiments must be designed to take into account the observed increase in apparent infiltration rate with increasing application rate.

5. Interior plot measurements: Spatial variability of infiltration includes the runon-runoff process, that is areas which have runoff flowing onto areas where the infiltration capacity is greater than the rainfall intensity. Quantification of this process will involve routing models which can account for dynamic infiltration and rainfall excess routing. In order to validate these models, runoff measurements must be taken not only at the end of the plot but also within the plot. If point measurements are made within the plot, then as stated in point 3

above, a correspondence must be made between point and plot scale measurements of infiltration.

6. Examine all components of the hydrograph: The only data available at the plot scale is the hydrograph. Progress in infiltration research is dependent on being able to define the change in depression and surface storage with time as well as being able to compute runoff.

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Predicting WEPP Effective Hydraulic Conductivity Values on Rangelands

Fred Pierson[1], Ken Spaeth[2], Mark Weltz[3], Steven Van Vactor[1], and Mary Kidwell[3]

Abstract

The Agricultural Research Service (ARS) and the Natural Resource Conservation Service (NRCS) cooperatively conducted rainfall simulation experiments at 26 sites in 10 western states for a total of 444 plot-runs. The data was combined with other similar rainfall simulation data from an additional 21 sites collected as part of the original ARS Water Erosion Prediction Project (WEPP) to create a database containing a total of 820 plot-runs. Subsets of this database were then used to estimate WEPP Green-Ampt effective hydraulic conductivity values for rangelands. This paper provides site-specific summaries of the soil, vegetation and hydrology data collected from all sites and presents regression equations for estimating time-invariant WEPP effective hydraulic conductivity values on rangelands.

Introduction

In 1990, the Agricultural Research Service (ARS) and the Natural Resource Conservation Service (NRCS) entered into a cooperative effort to specifically address the development of the Water Erosion Prediction Project (WEPP) model for use on rangelands. As a result of this cooperation, the National Range Study Team (NRST) and Interagency Rangeland Water Erosion Team (IRWET) were created. The NRST conducted rainfall simulation experiments at 26 sites in 10 western states for a total of 444 plot-runs. IRWET combined the NRST data with other similar rangeland rainfall simulation data from an additional 21 sites collected by the WEPP Team to create a database containing a total of 820 plot-runs. Subsets of this database were then used to develop, calibrate, and validate rangeland specific components of the WEPP model. This paper outlines the database and methodology IRWET used to estimate WEPP Green-Ampt effective hydraulic conductivity values for rangelands.

Rainfall Simulation Experiments

Rainfall Characteristics

A rotating boom simulator (Swanson, 1965; Swanson, 1979; Simanton et al., 1987, 1990) was used at all locations. It is trailer-mounted and has ten 7.6 m booms radiating from a central stem. The 30 nozzles on each boom spray continuously downward from an average height of 3 m. The boom movement is circular over the plots and applies rainfall intensities of approximately

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[1]USDA-ARS, NWRC, 800 Park Blvd, Plaza 4, Suite 105, Boise, ID 83712

[2]USDA-NRCS, NWRC, 800 Park Blvd, Plaza 4, Suite 105, Boise, ID 83712

[3]USDA-ARS, SWRC, 2000 East Allen Road, Tucson, AZ 85719

65 or 130 mm/hr with drop size distributions similar to natural rainfall. Simulation was done at 65 mm/hr on each plot for 60 minutes or until steady state runoff was achieved for a 'dry run' (initially dry), for 30 minutes or until steady state runoff occurred for the 'wet run' (initially at field capacity i.e. 24 hours after the 'dry run'), and finally 130 mm/hr of rainfall was applied until steady state runoff was achieved for the 'very wet run' (i.e. 30 minutes after the 'wet run').

Runoff Plots

Rainfall was simulated uniformly over three pairs of 3.05 by 10.67 m plots. Distribution of rainfall within each plot was determined by both non-recording and recording rain gauges. Runoff was determined by using pressure transducer bubble gauges calibrated to the flume positioned at the plot headwall (Simanton et al., 1987, 1990). Runoff samples were collected on timed intervals to measure sediment concentration and estimate total sediment yield. For the WEPP data, six plots were sampled at each site where paired plot treatments consisted of natural, clipped (vegetative material was clipped to a 20 mm height) and bare soil (all soil cover removed) treatments. Data from the natural plots were used to develop erosion, runoff, and infiltration relationships whereas data from the clipped plots were used to separate canopy cover effects on runoff and erosion. For the NRST data, all six plots sampled were undisturbed replicates of native vegetation where soil characteristics and slope were nearly constant.

Site Characteristics

Thirty-four of the 47 locations sampled were used in this analysis (sites and plots were removed from analysis due to missing data or runoff equilibrium problems). All sites were representative of common rangeland soils and plant cover types that contribute to variation in rangeland hydrology. Thirty unique combinations of rangeland cover type, range site, soil family, and soil surface texture were represented (Tables 1 and 2).

Soil Properties

Twenty-two pedons around each study site were examined, five representative pedons were selected and described, and a detailed profile description and characterization was done on one representative pedon. Soil descriptions and characterizations were performed by NRCS personnel and the NRCS National Soil Survey and Soil Mechanics Laboratories in Lincoln, Nebraska. Antecedent soil moisture condition of each plot and bulk density were determined using open-ended core and compliant cavity methods, respectively. Selected soil properties for each study site are shown in Table 1.

Vegetation Characteristics

Canopy and ground cover were determined by point-sampling (Mueller-Dombois and Ellenberg, 1974). The point center quarter method (Dix 1961) was used to determine plant parameters such as height, canopy cover, geometric shape, density, and mean distance of shrub, bunchgrass, sod, and annual grasses. Estimates of standing biomass of current year's growth by species and previous year's growth plus decumbent litter were collected utilizing SCS double sampling techniques (SCS, 1976) and by clipping and separating all biomass within five sub-plots located in each runoff plot. Biomass was determined by oven-drying and weighing each sample. Plant composition was determined by the weight method (SCS, 1976). A general description of vegetation characteristics for each site is given in Table 2.

Table 1. Abiotic mean site characteristics and optimized effective hydraulic conductivity (K_e) (mm hr^{-1}) values from USDA-IRWET [1] rangeland rainfall simulation experiments used to develop the baseline effective hydraulic conductivity equations for the WEPP model.

Location	Soil family	Soil series	Surface texture	Slope (%)	Organic matter (%)	Bulk density (g cm^{-3}) [2]	Mean optimized K_e	Range in optimized K_e Min.	Max.
1) Prescott, Arizona	Aridic argiustoll	Lonti	Sandy loam	5	1.3	1.6	7.0	4.1	9.8
2) Prescott, Arizona	Aridic argiustoll	Lonti	Sandy loam	4	1.3	1.6	5.6	3.4	6.9
3) Tombstone, Arizona	Ustochreptic calciorthid	Stronghold	Sandy loam	10	1.8	9.8	28.7	24.5	32.9
4) Tombstone, Arizona	Ustollic haplargid	Forest	Sandy clay loam	4	1.5	6.9	8.7	3.6	13.8
5) Susanville, California	Typic argixeroll	Jauriga	Sandy loam	13	6.4	32.9	16.7	15.3	18.7
6) Susanville, California	Typic argixeroll	Jauriga	Sandy loam	13	6.4	1.2	17.2	13.9	20.3
7) Akron, Colorado	Ustollic haplargid	Stoneham	Loam	7	2.5	1.5	7.3	1.5	15.0
8) Akron, Colorado	Ustollic haplargid	Stoneham	Sandy loam	8	2.4	1.5	16.5	8.4	23.0
9) Akron, Colorado	Ustollic haplargid	Stoneham	Loam	7	2.2	1.5	8.8	4.8	14.0
10) Meeker, Colorado	Typic camborthid	Degater	Silty clay	10	2.4	1.5	8.0	5.2	10.8
11) Blackfoot, Idaho	Pachic cryoborall	Robin	Silt loam	7	7.5	1.3	7.0	4.7	9.7
12) Blackfoot, Idaho	Pachic cryoborall	Robin	Silt loam	9	9.9	1.2	7.8	6.6	9.7
13) Eureka, Kansas	Vertic argiudoll	Martin	Silty clay loam	3	6.0	1.4	2.9	1.1	4.6
14) Sidney, Montana	Typic argiboroll	Vida	Loam	10	5.2	1.2	22.5	18.4	26.5
15) Wahoo, Nebraska	Typic argiudoll	Burchard	Loam	10	5.1	1.3	3.3	2.0	4.4
16) Wahoo, Nebraska	Typic argiudoll	Burchard	Loam	11	4.8	1.3	15.3	13.1	17.5
17) Cuba, New Mexico	Ustollic camborthid	Querencia	Sandy loam	7	1.5	1.5	16.5	14.5	18.5
18) Los Alamos, New Mexico	Aridic haplustalf	Hackroy	Sandy loam	7	1.4	1.5	6.3	5.2	7.3
19) Killdeer, North Dakota	Pachic haploborall	Parshall	Sandy loam	11	3.6	1.3	23.2	21.2	25.4

Table 1. *Continued.*

Location	Soil family	Soil series	Surface texture	Slope (%)	Organic matter (%)	Bulk density ($g\ cm^{-3}$)[2]	Mean optimized K_e	Range in optimized K_e Min.	Max.
20) Killdeer, North Dakota	Pachic haploborall	Parshall	Sandy loam	11	3.5	1.3	22.4	17.9	26.9
21) Chickasha, Oklahoma	Udic argiustoll	Grant	Loam	5	4.0	1.3	17.8	9.4	27.7
22) Chickasha, Oklahoma	Udic argiustoll	Grant	Sandy loam[3]	5	2.3	1.5	13.6	8.8	18.8
23) Freedom, Oklahoma	Typic ustochrept	Woodward	Loam	6	3.1	1.4	14.9	13.0	16.8
24) Woodward, Oklahoma	Typic ustochrept	Quinlan	Loam	6	2.3	1.5	20.4	15.5	25.9
25) Cottonwood, South Dakota	Typic torrert	Pierre	Clay	8	3.2	1.5	9.3	8.6	10.0
26) Cottonwood, South Dakota	Typic torrert	Pierre	Clay	12	3.7	1.4	3.6	2.7	4.4
27) Amarillo, Texas	Aridic paleustoll	Olton	Loam	3	3.0	1.5	8.4	6.5	9.7
28) Amarillo, Texas	Aridic paleustoll	Olton	Loam	2	2.5	1.5	5.8	2.4	10.4
29) Sonora, Texas	Thermic calciustoll	Perves	Cobbly clay	8	8.9	1.2	2.2	0.8	3.7
30) Buffalo, Wyoming	Ustollic haplargid	Forkwood	Silt loam	10	2.8	1.5	5.9	4.2	8.8
31) Buffalo, Wyoming	Ustollic haplargid	Forkwood	Loam	7	2.4	1.5	4.6	1.7	11.5
32) Newcastle, Wyoming	Ustic torriothent	Kishona	Sandy loam	7	1.7	1.5	21.7	14.8	26.3
33) Newcastle, Wyoming	Ustic torriothent	Kishona	Loam	8	2.2	1.5	23.1	20.0	28.6
34) Newcastle, Wyoming	Ustic torriothent	Kishona	Sandy loam	9	1.4	1.5	9.0	6.3	12.4

[1] Interagency Rangeland Water Erosion Team is comprised of USDA-ARS staff from the Southwest and Northwest Watershed Research Centers in Tucson, AZ and Boise, ID, and USDA-NRCS staff members in Lincoln, NE and Boise, ID.

[2] Bulk density calculated by the WEPP model based on measured soil properties including percent sand, clay, organic matter and cation exchange capacity.

[3] Farm land abandoned during the 1930's that had returned to rangeland. The majority of the 'A' horizon had been previously eroded.

Table 2. Biotic mean site characteristics from USDA-IRWET [1] rangeland rainfall simulation experiments used to develop the baseline effective hydraulic conductivity equation for the WEPP model.

Location	MLRA[2]	Rangeland cover type[3]	Range site	Dominant species by weight (descending order)	Canopy Cover (%)	Ground Cover (%)	Standing Biomass (kg ha^{-1})
1) Prescott, Arizona	35	Grama-Galleta	Loamy upland	Blue grama Goldenweed Ring muhly	48	47	990
2) Prescott, Arizona	35	Grama-Galleta	Loamy upland	Rubber rabbitbrush Blue grama Threeawn	51	50	2,321
3) Tombstone, Arizona	41	Creosotebush-Tarbush	Limy upland	Tarbush Creosotebush	32	82	775
4) Tombstone, Arizona	41	Grama-Tobosa-Shrub	Loamy upland	Blue grama Tobosa Burro-weed	18	40	752
5) Susanville, California	21	Basin Big Brush	Loamy	Idaho fescue Squirreltail Wooly mulesears Green rabbitbrush Wyoming big sagebrush	29	84	5,743
6) Susanville, California	21	Basin Big Brush	Loamy	Idaho fescue Squirreltail Wooly mulesears Green rabbitbrush Wyoming big sagebrush	18	76	5,743
7) Akron, Colorado	67	Wheatgrass-Grama-Needlegrass	Loamy plains #2	Blue grama Western wheatgrass Buffalograss	54	96	1,262
8) Akron, Colorado	67	Wheatgrass-Grama-Needlegrass	Loamy plains #2	Blue grama Sun sedge Bottlebrush squirreltail	44	86	936
9) Akron, Colorado	67	Wheatgrass-Grama-Needlegrass	Loamy plains #2	Buffalograss Blue grama Prickly pear cactus	28	82	477
10) Meeker, Colorado	34	Wyoming big sagebrush	Clayey slopes	Salina wildrye Wyoming big sagebrush Western wheatgrass	11	42	1,583

Table 2. *Continued.*

Location	MLRA[2]	Rangeland cover type[3]	Range site	Dominant species by weight (descending order)	Canopy Cover (%)	Ground Cover (%)	Standing Biomass (kg ha^{-1})
11) Blackfoot, Idaho	13	Mountain big sagebrush	Loamy	Mountain big sagebrush Letterman needlegrass Sandberg bluegrass	71	90	1,587
12) Blackfoot, Idaho	13	Mountain big sagebrush	Loamy	Letterman needlegrass Sandberg bluegrass Prairie junegrass	87	92	1,595
13) Eureka, Kansas	76	Bluestem prairie	Loamy upland	Buffalograss Sideoats grama Little bluestem	38	58	526
14) Sidney, Montana	54	Wheatgrass-Grama-Needlegrass	Silty	Dense clubmoss Western wheatgrass Needle & thread grass Blue grama	12	81	2,141
15) Wahoo, Nebraska	106	Bluestem prairie	Silty	Kentucky bluegrass Dandelion Alsike clover	27	80	1,239
16) Wahoo, Nebraska	106	Bluestem prairie	Silty	Primrose Porcupinegrass Big bluestem	22	87	3,856
17) Cuba, New Mexico	36	Blue grama-Galleta	Loamy	Galleta Blue grama Broom snakeweed	13	62	817
18) Los Alamos, New Mexico	36	Juniper-Pinyon Woodland	Woodland community	Colorado rubberweed Sagebrush Broom snakeweed	16	72	1,382
19) Killdeer, North Dakota	54	Wheatgrass-Needlegrass	Sandy	Clubmoss Sedge Crocus	69	96	1,613
20) Killdeer, North Dakota	54	Wheatgrass-Needlegrass	Sandy	Sedge Blue grama Clubmoss	71	88	1,422
21) Chickasha, Oklahoma	80A	Bluestem prairie	Loamy prairie	Indiangrass Little bluestem Sideoats grama	60	46	2,010

Table 2. *Continued.*

Location	MLRA[2]	Rangeland cover type[3]	Range site	Dominant species by weight (descending order)	Canopy Cover (%)	Ground Cover (%)	Standing Biomass (kg ha^{-1})
22) Chickasha, Oklahoma	80A	Bluestem prairie	Eroded prairie	Oldfield threeawn Sand paspalum Scribners dichanthelium Little bluestem	14	70	396
23) Freedom, Oklahoma	78	Bluestem prairie	Loamy prairie	Hairy grama Silver bluestem Perennial forbs Sideoats grama	39	72	1,223
24) Woodward, Oklahoma	78	Bluestem-Grama	Shallow prairie	Sideoats grama Hairy grama Western ragweed Hairy goldaster	45	62	1,505
25) Cottonwood, South Dakota	63A	Wheatgrass-Needlegrass	Clayey west central	Green needle grass Scarlet globemallow Western wheatgrass	46	68	2,049
26) Cottonwood, South Dakota	63A	Blue grama-Buffalograss	Clayey west central	Blue grama Buffalograss	34	81	529
27) Amarillo, Texas	77	Blue grama-Buffalograss	Clay loam	Blue grama Buffalograss Prickly pear cactus	23	97	2,477
28) Amarillo, Texas	77	Blue grama-Buffalograss	Clay loam	Blue grama Buffalograss Prickly pear cactus	10	87	816
29) Sonora, Texas	81	Juniper-Oak	Shallow	Buffalograss Curly mesquite Prairie cone flower Hairy tridens	39	68	2,461
30) Buffalo, Wyoming	58B	Wyoming big sagebrush	Loamy	Wyoming big sagebrush Prairie junegrass Western wheatgrass	53	59	7,591
31) Buffalo, Wyoming	58B	Wyoming big sagebrush	Loamy	Western wheatgrass Bluebunch wheatgrass Green needlegrass	68	60	2,901

Table 2. *Continued.*

Location	MLRA[2]	Rangeland cover type[3]	Range site	Dominant species by weight (descending order)	Canopy Cover (%)	Ground Cover (%)	Standing Biomass (kg ha^{-1})
32) Newcastle, Wyoming	60A	Wheatgrass-Needlegrass	Loamy plains	Prickly pear cactus Needle-and-thread Threadleaf sedge	11	77	1,257
33) Newcastle, Wyoming	60A	Wheatgrass-Needlegrass	Loamy plains	Cheatgrass Needle-and-thread Blue grama	56	81	2,193
34) Newcastle, Wyoming	60A	Wheatgrass-Needlegrass	Loamy plains	Needle-and-thread Threadleaf sedge Blue grama	32	47	893

1 Interagency Rangeland Water Erosion Team is comprised of USDA-ARS staff from the Southwest and Northwest Watershed Research Centers in Tucson, AZ and Boise, ID, and USDA-NRCS staff members in Lincoln, NE and Boise, ID.

2 USDA - Soil Conservation Service. 1981. Land resource regions and major land resource areas of the United States. Agricultural Handbook 296. USDA - SCS, Washington, D.C.

3 Definition of Cover Types from: T.N. Shiflet, 1994. Rangeland cover types of the United States, Society for Range Management, Denver, CO.

WEPP Effective Hydraulic Conductivities for Rangelands

When using WEPP on rangelands, the user should only use the time-invariant effective hydraulic conductivity (K_e) by setting the flag in line 2 of the soil file to 0. No provisions have been put in the model for changing K_e over time on rangelands. Therefore, users are advised against setting the flag in line 2 of the soil file to 1 when simulating rangeland conditions. Using a flag equal to 1 will allow the model to alter K_e based on cropland conditions. The selected input value for time-invariant K_e on rangelands must represent both the soil type and the management practice. This method differs from the curve number method in that no soil moisture correction is necessary since WEPP accounts for moisture differences via internal adjustments to the wetting front matric potential term of the Green and Ampt equation.

Baseline default equations for predicting WEPP optimized K_e values on rangelands were developed using rainfall simulation data collected on 150 plot-runs from 34 locations across the western United States (Table 1). K_e for each of the 150 plot-runs was obtained by optimizing the WEPP model based on total runoff volume (mm). Multiple regression procedures were then used to develop predictive equations for optimized K_e based on both biotic and abiotic plot-specific properties (Table 3). The resulting equations are as follows:

If rill surface cover (cover outside the plant canopy) is less than 45%, K_e is predicted by:

$$K_e = 57.99 - 14.05(\ln CEC) + 6.21(\ln ROOT10) - 473.39(BASR)^2 + 4.78(RESI)^2 \qquad (1)$$

where CEC is cation exchange capacity (meq/100 ml), ROOT10 is root biomass in top 10 cm of soil (kg m^{-2}), BASR is the fraction of basal surface cover in rill (outside the plant canopy) areas based on the entire overland flow element area (0-1), and RESI is the fraction of litter surface cover in interrill (under plant canopy) areas based on the entire overland flow element area (0-1). BASR is the product of the fraction of basal surface cover in rill areas (FBASR, expressed as a fraction of total basal surface cover) and total basal surface cover (BASCOV). RESI is the product of the fraction of litter surface cover in interrill areas (FRESI, expressed as a fraction of total litter surface cover) and total litter surface cover (RESCOV).

If rill surface cover is greater than or equal to 45%, K_e is predicted by:

$$\begin{aligned} K_e = & -14.29 - 3.40(\ln ROOT10) + 37.83(SAND) + 208.86(ORGMAT) \\ & + 398.64(RROUGH) - 27.39(RESI) + 64.14(BASI) \end{aligned} \qquad (2)$$

where SAND is the fraction of sand in the soil (0-1), ORGMAT is the fraction of organic matter found in the soil (0-1), RROUGH is soil surface random roughness (m), and BASI is the fraction of basal surface cover in interrill areas based on the entire overland flow element area (0-1). BASI is the product of the fraction of basal surface cover in interrill areas (FBASI, expressed as a fraction of total basal surface cover) and total basal surface cover (BASCOV).

The user is cautioned against using equations 1 and 2 with data falling outside the ranges of data values upon which the regression equations were developed. Ranges of values for each variable used in equations 1 and 2 are given in Table 3. Equations 1 and 2 have not been independently validated, however, they performed well at predicting K_e compared to the data set from which the equations were derived (Figure 1). The residuals plotted in Figure 2 show no bias and are similarly distributed between the two equations. Predictions of K_e were used in the model to predict runoff volume and peak runoff rate with the results shown in Figures 3 and 4.

Table 3. Ranges of values for variables used to develop equations 1 and 2.

Equation 1			
Variable	Mean	Minimum	Maximum
CEC (meq/100 ml)	20	7	45
ROOT10 (kg m^{-2})	0.45	0.09	0.99
BASR (0-1)	0.06	0.00	0.27
RESI (0-1)	0.34	0.05	0.84
Equation 2			
Variable	Mean	Minimum	Maximum
ROOT10 (kg m^{-2})	0.69	0.12	1.95
SAND (0-1)	0.43	0.02	0.71
ORGMAT (0-1)	0.04	0.02	0.10
RROUGH (m)	0.013	0.005	0.045
RESI (0-1)	0.16	0.02	0.41
BASI (0-1)	0.05	0.00	0.34

Future Research Needs

The assumption used in WEPP (95.7) that hydraulic conductivity is spatially uniform and temporally constant on rangelands is completely inadequate. However, the data does not currently exist to improve upon such an approach. If models of infiltration, such as the Green-Ampt model, are going to continue to be built on the concept of relating a driving force for flow with a resistance to flow, then methods of isolating each factor and accurately measuring soil and vegetation properties associated with each factor need to be drastically improved. Experimental procedures need to be developed which will quantify and explain the spatial and temporal variability in *in-situ* saturated and unsaturated hydraulic conductivity. Plot and watershed studies play an important role in this effort, but must be better designed to collect all necessary information and uniformly applied across time and space so data are consistent and additive. The ARS should consider defining and implementing an experimental procedure involving laboratory studies, rainfall simulation plots, permanent plots and small watershed areas at several locations throughout the U.S. which would provide the data necessary to build, validate and parameterize an infiltration model useful across all ARS hydrology and erosion models.

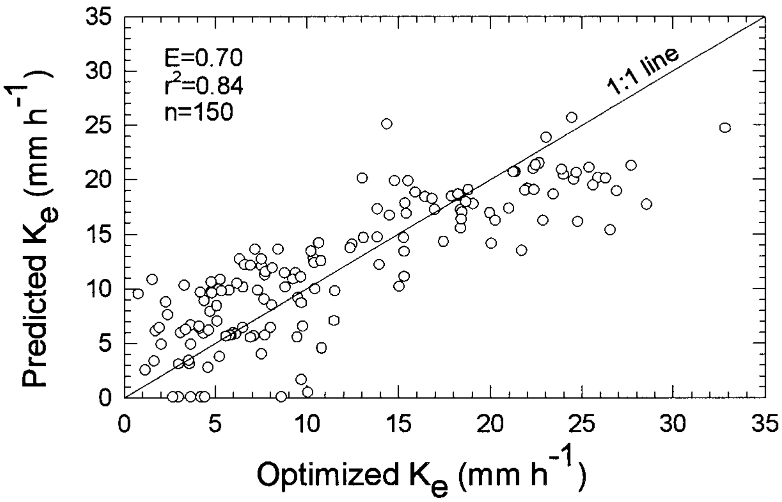

Fig. 1. Comparison of WEPP optimized and predicted effective hydraulic conductivity (K_e, mm h^{-1}) using equations 1 and 2. E is the coefficient of efficiency (Nash and Sutcliff, 1970), r^2 is the coefficient of determination and n is the number of data points.

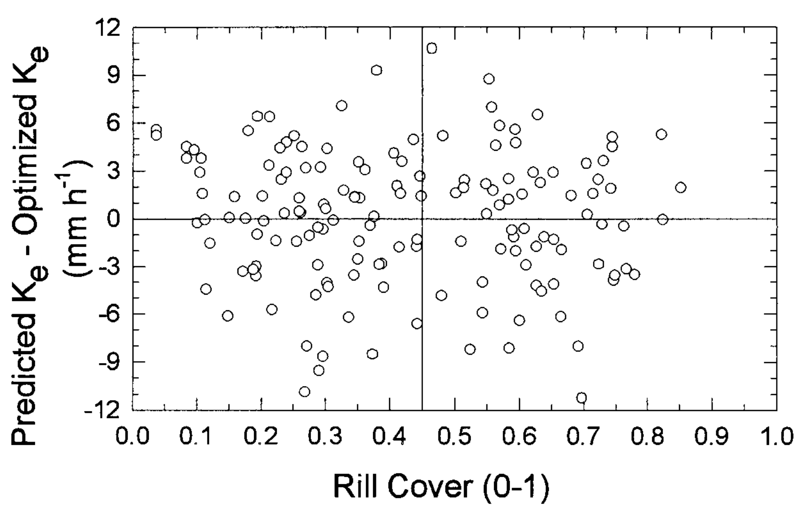

Fig. 2. Difference between WEPP optimized and predicted K_e using equations 1 and 2.

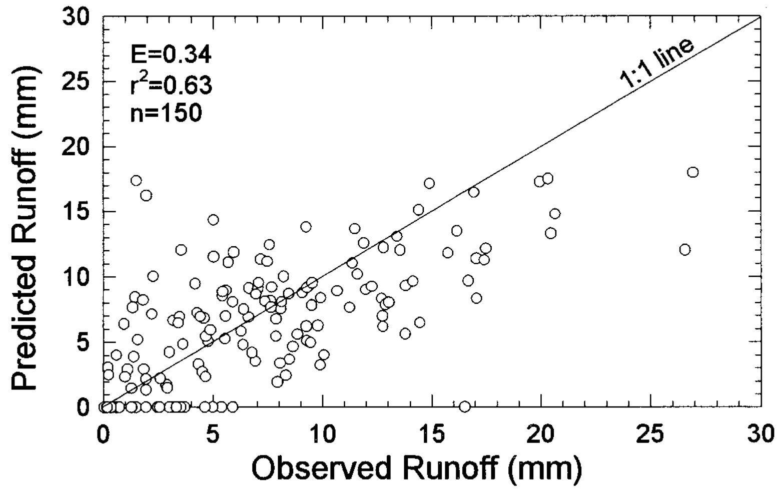

Fig. 3. Comparison of WEPP predicted runoff using K_e values estimated using equations 1 and 2 and observed runoff. The data set of observed runoff is from the same plots that equations 1 and 2 were developed from. E is the coefficient of efficiency (Nash and Sutcliff, 1970), r^2 is the coefficient of determination and n is the number of data points.

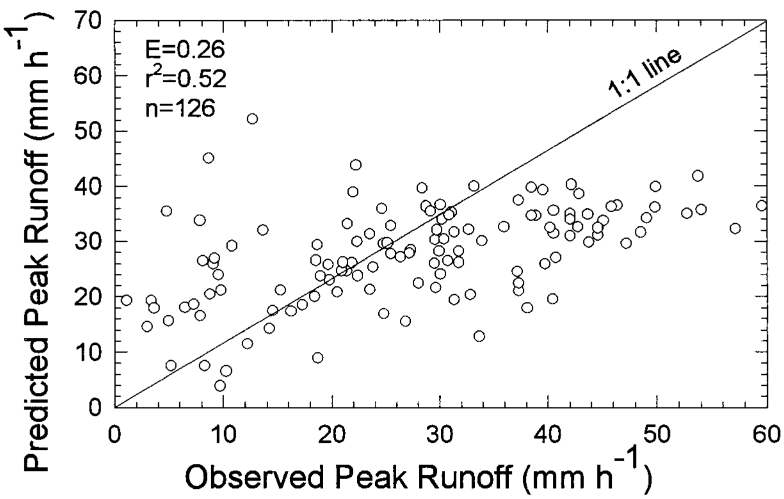

Fig. 4. Comparison of WEPP predicted peak runoff using K_e values estimated using equations 1 and 2 and observed runoff. The data set of observed peak runoff is from the same plots that equations 1 and 2 were developed from. E is the coefficient of efficiency (Nash and Sutcliff, 1970), r^2 is the coefficient of determination and n is the number of data points. The number of data points shown is 126 because 24 plots had zero predicted runoff.

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Infiltration Parameters for the WEPP Green-Ampt Model: Cropland Applications

Nearing, M.A., X. Zhang, B.Y. Liu, and L.M. Risse[1]

Abstract

The Green-Ampt equation has been in existence for more than 80 years, but is rarely used for prediction purposes because of lack of reliable parameter data. Modern, process-based erosion models require the use of process-based infiltration and runoff models. The Water Erosion Prediction Project (WEPP), for example, uses the Green-Ampt infiltration model to predict runoff curves during rainfall events. This paper presents an overview of the parameterization process used by the WEPP technology for the Green-Ampt model. It is by necessity and design quite empirically based. The process of development of the presented relationship was to A) use the WEPP/Green-Ampt model itself in the parameterization process using optimization techniques, B) to rely as much as possible on natural rainfall and runoff data, while using other technologies where necessary, and C) to choose system parameters as predicted by the model as internal model predictors of infiltration parameters. This work represents, perhaps for the first time ever, a comprehensive and well-tested method for parameterizing the Green-Ampt infiltration model.

Introduction

The key parameter for WEPP in terms of infiltration is the Green-Ampt effective conductivity parameter (K_e). This parameter is related to the saturated conductivity of the soil, but it is important to note that it is not the same as or equal in value to the saturated conductivity of the soil. The second soil-related parameter in the Green-Ampt model is the wetting front matric potential term. That term is calculated internal to WEPP as a function of soil type, soil moisture content, and soil bulk density: it is not an input variable. If the user does not know the effective conductivity of the soil, he/she may insert a zero and the model will calculate a value based on the equations presented here for the time-variable case.

The model will run in 2 modes by either: A) using a "baseline" effective conductivity (K_b) which the model automatically adjusts within the continuous simulation calculations as a function of soil management and plant characteristics, or B) using a constant input value of K_e. The second number in line 2 of the soil file contains a flag (0 or 1) which the model uses to distinguish between these two modes. A value of 1 indicates that the model is expecting the user to input a K_b value which is a function of soil only, and which will be internally adjusted to account for management practices. A value of 0 indicates the model is expecting the user to input a value of K_e which will not be internally adjusted and must therefore be representative of both the soil and the management practice being modeled. It is essential that the flag (0 or 1) in line 2 of the soil

[1]USDA-ARS National Soil Erosion Research Laboratory. West Lafayette, IN 47907-1196

file be set consistently with the input value of effective conductivity for the upper soil layer.

Temporally-varying Case: "Baseline" Effective Conductivity

Values for "baseline" effective conductivity (K_b) may be estimated using the following equations:

For soils with ≤ 40% clay content:

$$K_b = -0.265 + 0.0086\, Sand^{1.80} + 11.46 CEC^{-0.75} \tag{1a}$$

For soils with > 40% clay content:

$$K_b = 0.0066 e^{(2.44/Clay)} \tag{1b}$$

where *Sand* and *Clay* are the percent of sand and clay, and CEC (meq/100g) is the cation exchange capacity of the soil. In order for these equations to work properly, the input value for cation exchange capacity should always be greater than 1 meq/100g. These equations were derived based on model optimization runs to measured and curve number (fallow condition) runoff amounts. Forty three soils files were used to develop the relationships. Table 1 shows the results of the optimization and the estimated values of K_b. Figure 1 is a plot of optimized vs. estimated K_b for the 43 soils. Table 2 shows the results of comparisons to measured natural runoff plot data from 11 sites. Model efficiency is a quantification of how well the model predicted runoff on an individual storm basis. At each of the eleven sites the model predicted runoff better on a storm-by-storm basis using the estimated K_b values (Eq. 1a and 1b) than did the curve number approach. For purposes of erosion prediction it is more important to predict the individual storms accurately than to predict the total annual runoff volume, because it is a relatively small number of intense storms which cause most of the erosion.

Physically, the K_b value should approximate the value of K_e for the first storm after tillage on a fallow plot of land. Figure 2 shows a plot of the optimized K_b versus a measurement of K_b obtained using the data from the WEPP erodibility sites under a rainfall simulator. These values are also listed in Table 1. In general, the rainfall simulator measured K_b values tended to be greater than the corresponding optimum K_b values.

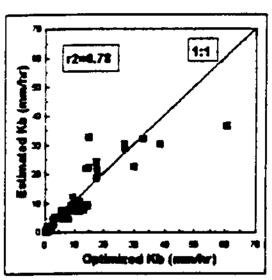

Fig. 1. Estimated values of baseline hydraulic conductivity for time-variable case plotted against those calibrated from curve number predictions.

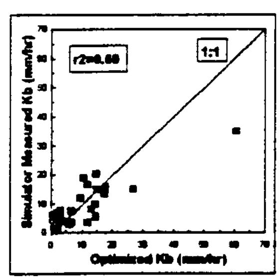

Fig. 2. Baseline hydraulic conductivity values for time-variable case measured under rainfall simulation compared to those calibrated from curve number predictions.

Table 1. Optimized and Estimated Effective Conductivity Values for the Case of Constant K_e and Baseline K_e.

Soil	Sand Content %	Clay Content %	Organic Matter Content %	CEC meq/100g	Simulator Measured K_e mm/hr	Opt. Constant K_{ef} mm/hr	Est. Constant K_{ef} mm/hr	Opt. Baseline K_b mm/hr	Est. Baseline K_b mm/hr
Sharpsburg	5.2	40.1	2.8	29.4	7.3	1.1	1.0	1.8	1.8
Hersh	72.3	10.9	1.1	7.7	15.8	3.9	3.9	17.6	21.3
Keith	48.9	19.3	1.5	18.3	3.5	2.7	2.9	11.5	10.5
Amarillo	85.0	7.3	0.3	5.1	15.0	3.4	4.5	26.6	28.7
Woodward	51.7	13.0	2.2	11.6	12.0	2.5	3.0	9.2	12.0
Heiden	8.6	53.1	2.2	33.3	4.7	0.2	0.2	0.34	0.45
LosBanos	15.5	43.7	2.0	39.1	3.9	0.5	0.7	1.1	1.1
Portneuf	19.5	11.1	1.2	12.6	7.9	1.3	1.6	2.7	3.0
Nansene	20.1	12.8	1.9	16.6	5.3	1.5	1.6	2.8	3.0
Palouse	9.8	20.1	2.6	19.6	2.6	1.2	1.2	2.0	1.5
Zahl	46.3	24.0	2.5	19.5	5.7	2.7	2.8	14.1	9.5
Pierre	16.9	49.5	2.7	35.7	2.4	0.2	0.2	0.71	0.61
Williams	40.8	26.9	2.6	22.7	8.3	2.4	2.5	12.9	7.7
BarnesND	39.3	26.5	3.9	23.2	16.7	2.4	2.5	11.7	7.2
Sverdrup	75.3	7.9	2.0	11.0	20.3	3.6	4.0	14.5	22.2
BarnesMN	48.6	17.0	3.2	19.5	19.1	2.9	2.9	10.4	10.3
Mexico	5.5	25.3	2.5	21.3	6.2	0.2	0.2	0.34	1.1
Grenada	1.8	20.2	1.8	11.8	3.4	0.5	0.4	0.7	1.6
Tifton	86.4	2.8	0.7	2.1	14.9	3.5	4.5	14.8	32.6
Bonifay	91.2	3.3	0.5	1.7	34.8	7.1	6.9	60.2	36.4
Cecil	69.9	11.5	0.7	2.0	13.3	3.6	3.8	17.2	24.4
Hiwassee	63.7	14.7	1.3	4.4	13.6	4.0	3.5	17.2	18.7
Gaston	37.2	37.9	1.7	9.2	3.6	1.2	1.0	6.3	7.7
Opequon	37.7	31.1	2.3	12.9	7.6	1.1	1.1	6.3	7.3
Frederick	25.1	16.6	2.1	8.2	2.9	1.6	1.8	5.9	4.9
Manor	44.0	25.2	2.5	13.2	10.0	2.8	2.7	14.1	9.2
Collamer	6.0	15.0	1.7	9.2	3.6	0.5	0.5	0.73	2.1
Miamian	31.3	25.9	2.4	14.9	4.4	0.7	0.9	3.3	5.5
Lewisburg	38.5	29.3	1.4	12.5	3.7	1.0	1.1	5.5	7.6
Miami	4.2	23.1	1.3	13.3	0.9	1.0	0.9	1.7	1.5
Colonie	90.5	2.1	0.1	10.0		7.2	6.9	38.3	30.4
Pratt	89.0	2.2	0.4	3.1		6.3	6.9	32.8	32.4
Shelby	27.8	29.0	3.0	16.5		1.7	2.0	7.8	4.6
Monona	7.1	23.5	2.0	20.1		1.1	1.1	1.9	1.2
Ontario	44.2	14.9	4.5	11.8		2.7	2.7	8.6	9.4
Stephensville	73.2	7.9	1.6	7.2		3.9	3.9	13.7	21.9
Providence	2.0	19.8	0.8	9.3		0.4	0.4	0.7	1.9
Egan	7.0	32.2	3.7	25.1		1.5	1.1	1.8	1.0
Barnes	39.4	23.2	3.4	18.4		2.4	2.5	10.0	7.4
Thatuna	28.0	23.0	4.3	16.2		1.1	0.9	2.6	4.6
Caribou	38.8	13.7	3.8	13.2		2.8	2.4	8.2	7.6
Tifton	87.0	5.7	0.7	4.1		3.4	4.5	26.6	30.4
Cecil	66.5	19.6	0.9	4.8		4.5	3.7	29.7	22.8

Table 2. WEPP estimated runoff in terms of: A) model efficiency on a storm-by-storm basis and B) in terms of average annual runoff.

A. Comparison of model efficiency

Site	Number of Years	Number of Events	Model Efficiency WEPP Opt. K_b	Model Efficiency CN	Model Efficiency WEPP Est. K_b
Bethany, MO	10	109	0.82	0.72	0.81
Castana, IA	12	90	0.48	0.10	0.12
Geneva, NY	10	97	0.73	0.58	0.62
Guthrie, OK	15	170	0.86	0.77	0.85
Holly Springs, MS	8	208	0.87	0.79	0.69
Madison, SD	10	60	0.77	0.69	0.74
Morris, MN	11	72	0.59	-1.06	-0.21
Pendleton, OR	11	82	0.06	-0.33	-0.69
Presque Isle, ME	9	99	0.45	-0.25	0.32
Tifton, GA	7	64	0.67	0.24	0.59
Watkinsville, GA	6	110	0.84	0.74	0.84

B. Comparison of annual runoff

Site	Number of years	Rainfall	Annual runoff Meas.	Annual runoff CN	Annual runoff WEPP
Bethany, MO	10	754	222	175	205
Castana, IA	12	747	102*	125	148
Geneva, NY	10	828	168*	79	110
Guthrie, OK	15	745	154	78	121
Holly Springs, MS	8	1328	557	216	299
Madison, SD	10	577	56*	69	65
Morris, MN	11	604	40*	33	75
Pendleton, OR	11	595	71	60	27
Presque Isle, ME	9	846	107*	89	47
Tifton, GA	7	1227	289	135	171
Watkinsville, GA	6	1445	431	395	392

*indicates winter runoff not measured

Temporally-varying Case: Fallow Soil Adjustments to Effective Conductivity

In the natural system the hydraulic conductivity of the soil matrix is dynamically responding to changes in the surrounding environment. Therefore, to improve the accuracy of infiltration estimates obtained from the Green-Ampt equation in continuous simulation models, reliable estimates of the hydraulic conductivity during each event are necessary. This requires not only an appropriate input value, but also a method for adjusting the hydraulic conductivity to account for temporal changes in the physical condition of the soil. The method which is used to adjust the effective hydraulic conductivity parameter in the WEPP model was based on the results of a study which used over 220 plot years of natural runoff plot data from 11 different locations. By optimizing the effective Green-Ampt hydraulic conductivity for each event within a simulation, a method of determining the temporal variability in the hydraulic conductivity function was established (Risse, 1994). After a detailed statistical analysis of several different WEPP parameters and functions, the following equation was selected to account for the effects of soil crusting and tillage on the effective hydraulic conductivity:

$$K_e = K_b \left[CF + (1-CF)\, e^{(-C\, Ea\, (1 - rr/4))} \right] \tag{2}$$

where K_e and K_b are the effective conductivity for any given event and the baseline hydraulic conductivity (mm/hr), CF is the crust factor which ranges from 0.20 to 1.0, C is the soil stability factor (m^2/J), Ea is the cumulative kinetic energy of the rainfall since the last tillage operation J/m^2), and rr is the random roughness of the soil surface (cm). This equation has a similar form to the relationships which have been proposed by Van Doren and Allmaras (1978), Eigel and Moore (1983), and Brakensiek and Rawls (1983). By selecting this form for the equation, it was assumed that the value of K_b will represent a freshly tilled or maximum hydraulic conductivity which will decrease exponentially at a rate proportional to the kinetic energy of the rainfall since last tillage as it approaches the fully crusted or final value. While this form is consistent with those in the literature, most of those have been used to calculate the hydraulic conductivity at some time within a given event rather than for a series of successive events. Generally, the energy associated with the rainfall rather than the amount is thought to control the rate at which the surface seal forms. The random roughness term is important as crust rarely forms on surfaces with random roughness greater than 4 cm and the reduction of effective hydraulic conductivity due to the crust will generally be more significant on smoother surfaces (Rawls et al., 1990).

The crust factor, CF, provides a means of estimating the final or fully crusted hydraulic conductivity based on the baseline values. The fully crusted hydraulic conductivity is simply the baseline value multiplied by the crust factor. A relationship developed by Rawls et al. (1990) which states:

$$CF = SC \,/\, (1+\Psi/L) \tag{3}$$

where SC is the correction factor for partial saturation of the subcrust soil, y is the steady state capillary potential at the crust/subcrust interface, and L is the wetted depth. They also derive the following continuous relationships for SC and:

$$SC = 0.736 + 0.0019(\%Sand) \tag{4}$$

$$\Psi = 45.19 - 46.68\ SC \tag{5}$$

The depth to the wetting front is calculated in WEPP as:

$$L = 0.147 - 0.0015(\%Sand)^2 - 0.00003(\%Clay)\rho_b \tag{6}$$

where ρ_b is the bulk density (kg/m^3). If the calculated value of L is less than the crust thickness (0.005 m in WEPP) then it is set equal to the crust thickness. Rawls et al. (1990) used data from 36 covered and uncovered plots to validate the fact that this method could provide reasonable estimates of crusted hydraulic conductivities based on freshly tilled hydraulic conductivities. Table 3 compares the crust factor calculated using these equations to two values of maximum adjustment taken from the natural runoff plot data. At six of the ten sites, the calculated crust factor was within 10% of the maximum adjustment calculated from the data. At Bethany and Castana the reduction in hydraulic conductivity was not as significant as that predicted by the crust factor, while the data from Holly Springs indicated that the crust factor should have been slightly higher. The data indicated that the crust factor calculated by the equations of Rawls et al. (1990) can adequately predict the maximum reduction in conductivity due to crust formation.

Table 3. Comparison of optimized and calculated values for the crust factors and soil stability constants.

Site*	Avg. K_e for	Avg. K_e	CF calc.	CF	CF from	Optimum	Calculated
Bethany	1.72	0.61	0.35	0.77	0.20	0.0001	0.0051
Castana	1.87	1.18	0.63	0.63	0.27	0.0002	0.0001
Geneva	4.35	1.85	0.42	0.27	0.37	0.0020	0.0041
Holly Springs	1.40	0.11	0.08	0.27	0.29	0.0009	0.0036
Madison	3.84	0.70	0.18	0.33	0.20	0.0007	0.0001
Morris	11.57	2.11	0.18	0.23	0.27	0.0034	0.0033
Pendleton	*	0.45	*	0.14	0.28	0.0015	0.0026
Presque Isle	4.13	1.18	0.28	0.16	0.38	0.0033	0.0014
Tifton	13.18	2.16	0.20	0.20	0.20	0.0118	0.0122
Watkinsville	8.13	2.73	0.20	0.55	0.20	0.0312	0.0295

* Pendleton did not have any events with less than 80 mm of rainfall since last tillage.

The soil stability factor, C, represents the rapidity that the effective conductivity declines from K_b to its fully crusted value. The values obtained by fitting E.Q.. (2) to the optimized effective conductivities for the natural runoff plot data ranged from 0.00006 to 0.0312 m^2/J. This generally agreed with the range of values reported in the literature (0.00012-0.0356). For this equation to be widely applicable, the user must have a method for obtaining accurate values of C since few measured values are readily available. Regression analysis between the C values given in Table 3 and soil properties indicated that the primary soil factors influencing the rate of surface seal development were %sand (r=0.68), bulk density (r=0.66), and %silt (r=-0.72). Bosch and Onstad (1988) had similar findings in a study they conducted. Based on these findings, the following equation was developed to estimate the soil stability factor based on soil properties:

$$C = -0.0028 + 0.000113\ (\%Sand) + 0.00125\ (\%Clay/CEC) \tag{7}$$

where CEC is the cation exchange capacity (meq/100g). Bounds of $0.0001<C<0.010$ were imposed on this equation to prevent negative C values on soils with very low sand and clay contents. Using this equation, soils with high amounts of sand or clay and a low CEC would form crust more rapidly. Equation (7) provided estimates of C which were within one order of magnitude of the optimized values for eight of the ten sites (Table 3).

Figure 3 shows the optimized event conductivities plotted against those calculated using the tillage adjustment with an optimized baseline hydraulic conductivity for soils with a high, medium, and low value of C. In these figures, it is evident that the tillage adjustment using the estimated C values is accurately predicting the trend of a reduction in K_e with increasing rainfall kinetic energy since last tillage, however, this adjustment does not account for most of the variability in the K_{opt} values.

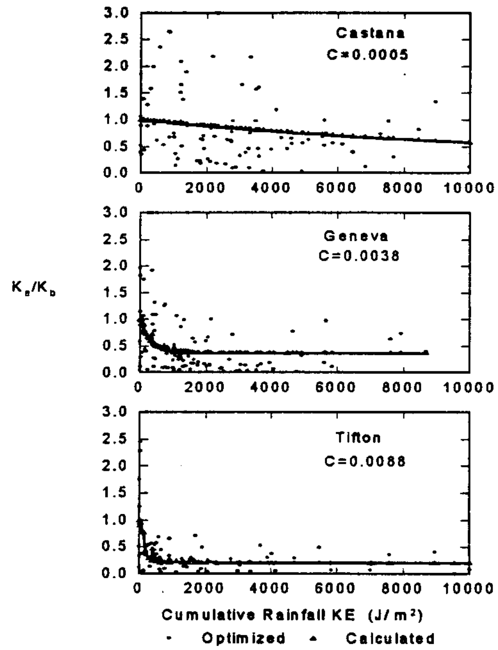

Fig. 3. Comparison of optimized effective conductivities to effective conductivities predicted by the proposed tillage adjustments at three sites.

To compare the effects of using each of these adjustments on predicted runoff amounts, each adjustments was incorporated into WEPP. Two different WEPP versions were tested; 1) a constant K_e version in which no temporal variation was allowed, and 2) a version which included the tillage and crusting adjustment. Both versions were run using calibrated values of hydraulic conductivity. The optimized baseline conductivities and model efficiencies of each of the versions is given in Table 4. The baseline values of hydraulic conductivity were all higher than the effective conductivities obtained for the constant value version. This was expected since the constant values represent the average effective conditions rather than the freshly tilled conditions. Using the tillage adjustment the average effective value, K_e was approximately 42% of K_b. The average model efficiency was higher for the version of the model which used the tillage adjustment and this version performed the better at nine of the eleven sites. The correlation coefficients, r^2, were generally close to the model efficiencies and indicated the same trends. The slope and intercept of the regression line between measured and predicted values can be used as a measure of bias. The results from a perfect model would have a slope of one and a intercept of zero. For both versions of the model and almost every site, the slopes were less than one and the intercepts were greater than zero. This indicates that both of the versions were over-predicting runoff on the smaller events and under-predicting runoff on the larger events The version with the tillage adjustment appeared to be less biased as it had a higher slope.

Table 4. Comparison of optimized baseline conductivities and model results for WEPP a constant value of hydraulic conductivity and the temporally varying values.

Version:	Constant K_e					Kb with tillage and crusting adjustment				
Site	K_e	ME*	Slp.	Int.	r^2	K_b	ME	Slp	Int	r^2
Bethany	1.22	0.81	0.90	0.02	0.81	3.65	0.82	0.91	0.81	0.82
Castana	2.04	0.46	0.82	0.50	0.59	2.38	0.49	0.84	0.05	0.62
Geneva	2.27	0.63	0.83	0.67	0.67	5.14	0.72	0.80	0.32	0.74
Guthrie	6.19	0.85	0.97	-0.99	0.87	16.73	0.85	0.97	-1.04	0.87
Holly Springs	0.31	0.84	0.87	1.39	0.84	0.72	0.87	0.85	1.82	0.87
Madison	1.80	0.74	0.69	1.57	0.75	2.01	0.77	0.71	1.42	0.78
Morris	7.68	0.40	0.69	0.05	0.52	16.41	0.59	0.74	-0.29	0.66
Pendleton	0.51	0.07	0.61	-0.18	0.41	1.76	0.07	0.67	-0.12	0.41
Presque Isle	2.38	0.19	0.55	1.12	0.36	3.82	0.46	0.63	0.68	0.53
Tifton	7.87	0.49	0.79	0.77	0.59	18.14	0.66	0.85	2.19	0.69
Watkinsville	4.41	0.84	0.97	-0.81	0.86	19.15	0.84	1.01	-1.13	0.87
Average		0.56	0.79	0.37	0.66		0.65	0.82	0.43	0.71

* Model efficiency calculated between WEPP predicted runoff and measured values.
Regression statistics calculated between measured and predicted runoff.

Temporally-varying Case: Cropping Adjustments to Effective Conductivity

Temporal Adjustment for Row Crops

Surface cover is known to be effective in reducing soil crusting and increasing effective hydraulic conductivity (K_e). Flow through macropores formed by root and soil fauna under cropped conditions plays an important role in increasing K_e. As compared to the corresponding fallow conditions, the degree of the increase under cropped conditions heavily depends on crop and residue management practices, tillage systems, soil properties, rainfall characteristics as well as their interactions. Wischmeier (1966) observed, that water infiltration was more a characteristic of surface conditions and management than of a specific soil type, and that infiltration increased with larger storms. This indicates that the effects of these variables and their interactions must be considered in order to successfully apply the Green-Ampt equation to cropped conditions.

A total of 328 plot-years of data from natural runoff plots on 8 sites with 1912 measured runoff values were used to develop equations for adjusting K_e for row cropped conditions (Table 5). The management input file were compiled according to recorded data. Plant growth parameters were calibrated to obtain realistic above ground biomass. Soil, slope, and climate input files were prepared based on measured data. Events which accounted for about 60-70% of the total annual runoff were strictly selected from each site based on data quality.

Canopy height has a significant effect on surface runoff (Khan et al., 1987). Based on the measured fall velocities for a raindrop size of 2.5 mm at variant fall heights (Laws, 1941), the following correction factor (C_h) for canopy height effectiveness as cover relative to infiltration was developed

$$C_h = e^{-0.33*h} \qquad r^2 = 0.99 \tag{8}$$

where h is the fall height in m. Average fall height was calculated as one half of the crop height in WEPP. With E.Q.. (8), the effective canopy cover (ccovef) can be computed by

$$ccovef = cancov*C_h \tag{9}$$

in which cancov is the canopy cover. The total effective surface cover (scovef) can be computed by

$$scovef = ccovef + rescov - ccovef*rescov \tag{10}$$

Table 5. Site and crop management descriptions.

Site	Crop management	Number of reps	Years	Number of events
Holly Springs, MS	a. fallow	2	1961-68	208
slope: 0.05 m/m	b. cont. corn, spring TP †	2	"	163
size: 4x22.3 m				
Madison, SD	a. fallow	3	1962-70	59
slope: 0.06 m/m	b. cont. corn, spring TP	3	"	48
size: 4x22.3 m	c. cont. corn, no TP	3	"	50
	d. cont. oats	3	1962-64	15
Morris, MN	a. fallow	3	1962-71	67
slope: 0.06 m/m	b. cont. corn, fall TP	3	"	67
size: 4x22.3 m				
Presque Isle, ME	a. fallow	3	1961-65	65
slope: 0.08 m/m	b. cont. potato	3	"	64
size: 3.7x22.3 m				
Watkinsville, GA	a. fallow	2	1961-67	147
slope: 0.07 m/m	b. cont. corn, spring TP	2	"	97
size: 4x22.3 m	c. cont. cotton, spring TP	2	"	112
Bethany, MO	a. fallow	1	1931-40	109
slope: 0.07 m/m	b. cont. corn, spring TP	1	"	112
size: 4.3x22.3 m				
Geneva, NY	a. fallow	1	1937-46	97
slope: 0.08 m/m	b. summer fallow, winter rye	1	"	77
size: 1.8x22.3 m	c. cont. soybean, spring TP	1	"	45
Guthrie, OK	a. fallow	1	1942-56	170
slope: 0.08 m/m	b. cont. cotton, spring TP	1	"	140
size: 1.8x22.3 m				

† TP, turn plow

Correlation coefficients of selected variables to optimized K_e for each site are tabulated in Table 6. For cover related variables, the correlation coefficients from the pooled data increased in the order of: cancov, ccovef, rescov, and scovef. This sequence implies that 1). The adjustment of canopy cover by E.Q. (9) is useful; 2). Residue cover is more correlated to K_e than canopy cover; 3). The combined effects of the two are greater than either one of them when used alone. The rainfall amount (rain) showed a very strong correlation with K_e. This behavior could be explained by macropore flow phenomena. More importantly, the product of rain and scovef exhibited a better overall correlation coefficient than either rain or scovef, indicating a positive interaction between the two. Thus, this interactive product should be a better predictor for K_e.

Table 6. Correlation Coefficients of selected variables to optimized event hydraulic conductivities.

Site	Canopy cover	Effective canopy cover (ccovef)	Residue cover (rescov)	Total effective surface cover (scovef)	Residue mass on ground	Buried residue mass	Total root mass	Days since last tillage	Rainfall amount (rain)	Rainfall-cover term†
Holly Springs	.11	.11	.26	.27	.24	.17	.31	.20	.31	.41
Madison	.20	.19	.03‡	.17	.07‡	.08‡	.17	-.01‡	.28	.32
Morris	.04‡	.05‡	-.02‡	.05‡	-.01‡	-.04‡	.04‡	-.15‡	.68	.20
Presque Isle	-.04‡	-.04‡	-.16‡	-.08‡	.00‡	.04‡	.01‡	-.05‡	.33	.06
Watkinsville	.18	.19	.20	.31	.18	.28	.31	.05‡	.40	.49
Bethany	.16	.17	-.10‡	.14	-.09‡	.12‡	.06‡	.00‡	.27	.22
Geneva	.43	.42	.27	.49	.28	.37	.49	.08‡	.64	.82
Guthrie	.14	.15	.06‡	.16	.05‡	.30	.27	-.18	.42	.28
pooled ¶	.10	.12	.13	.20	.14	.17	.06	.11	.38	.39

† Rain*scovef.
‡ Not significant at 5% level.
¶ Using the lumped database from all the sites.

Based on the above analyses, the final model structure was proposed as

$$K_e = K_{bare}(\ 1\text{-scovef}\ )\ +\ c*rain*scovef \qquad (11)$$

where K_{bare} is the K_e of bare area in mm/h and can be estimated by Eq. (2), c is a regression coefficient and was estimated for each soil series at each site, and rain is the storm rainfall amount in mm. This equation assumes that K_e for any given area can be conceptualized as the arealy weighted average of K_{bare} and K_e in covered area. The latter, being closely related to the variable of rain*scovef, can be well represented by this variable. This model formulation attempts to reflect the general conditions. For the fallow case, Eq. (11) reduces to $K_e = K_{bare}$. While under the fully covered conditions, the effect of soil crusting is neglected and K_e is adjusted for the effects of surface cover and rainfall amount. The c values were strongly related to basic soil properties such as sand and clay content, and to K_b which is estimated from basic soil properties (Eq. 1). The relationship to K_b can be described by

$$c = 0.0534 + 0.01179*K_b \qquad (12)$$

where K_b is in mm/h. Substituting c with above equation, the final adjustment equation becomes:

$$K_e = K_{bare}(1 - scovef) + (0.0534 + 0.01179*K_b)*rain*scovef \qquad (13)$$

The predicted mean K_e and total WEPP predicted runoff using Eq. (13), along with measured values, are presented in Table 7. The predicted mean K_e agreed well with the optimized mean K_e. The total measured runoff of the selected events and the total predicted runoff matched well with r^2 and slope of regression being 0.94 and 0.99, respectively. The model efficiency, calculated on an event basis, averaged 0.64, which indicates that Eq. (13) works better than just using a constant mean for K_e. This can also be clearly seen in Fig. 4. In addition, the seasonal variation of K_e and runoff were also represented by the equation.

Table 7. Comparison of optimized and predicted effective conductivity, K_e, and measured and predicted total runoff for the selected events.

Site	Management	Total rainfall mm	K_e* Optimized mm/h	K_e* Predicted mm/h	Total runoff Measured mm	Total runoff Predicted mm	Model efficiency (ME)
Holly Springs	fallow	5742	0.53				
	corn	5049	1.34	1.20	1793	2014	.582
Madison	fallow	1553	1.54				
	corn TP	1310	1.83	1.62	322	311	.783
	corn No TP	1359	1.76	1.75	311	275	.747
	oats	410	1.86	1.76	86	88	.775
Morris	fallow	1985	5.85				
	corn	1987	6.11	6.74	319	310	.340
Presque Isle	fallow	1321	1.53				
	potato	1296	1.57	2.75	432	231	.291
Watkinsville	fallow	4277	3.34				
	corn	3566	8.45	9.66	675	793	.823
	cotton	3846	7.36	8.65	834	911	.791
Bethany	fallow	3330	1.42				
	corn	3375	1.73	1.60	1375	1308	.845
Geneva	fallow	2292	2.40				
	winter rye	1912	3.95	2.91	375	534	.511
	soybean	1446	8.70	3.66	51	338	xx
Guthrie	fallow	5313	5.58				
	cotton	4820	8.16	8.87	1239	1204	.793

* Means of all selected events. Calculated on an event basis.

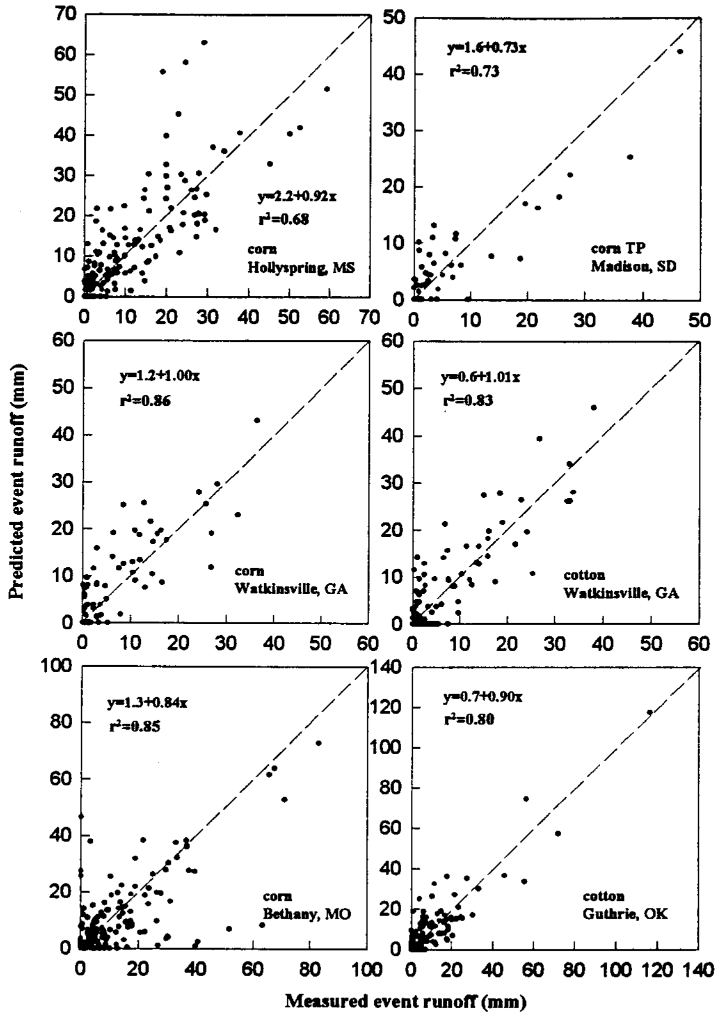

Fig. 4. Measured vs. predicted runoff for each individual storm on selected sites under row crop conditions.

Temporal Adjustment for Perennial Crops

The sites used for row crop adjustment were also used for perennial crops except for the Madison site where perennial crops were not grown (Table 8). Therefore, the same climate, slope, and soil input files were used. The management input files were prepared according to the recorded data, and the plant growth parameters were calibrated. Two common cropping systems, continuous meadow and rotation meadow, were included. A total of 88 plot-years of data with 506 measured runoff values were used for the validation.

Table 8. Background information of the database used for K_e adjustment under perennial crops.

Site	Crop management	Number of reps	Periods used	Years in meadow	Number of events used
Holly Springs, MS	meadow-corn-meadow	2	1962-68	5	101
Morris, MN	meadow-corn-oats	3	1962-71	4	18
Presque Isle, ME	potato-oats-meadow	3	1961-65	1	4
Watkinsville, GA	corn-meadow-meadow	2	1961-67	4	44
Bethany, MO	cont. alfalfa	1	1931-40	10	83
	cont. blue grass	1	"	10	79
Geneva, NY	cont. red clover	1	1937-41	5	19
	cont. blue grass	1	1937-46	10	30
Guthrie, OK	cont. blue grass	1	1942-56	15	96
	wheat-meadow-cotton	1	"	5	32

Since similar correlation relationships between the selected variables and optimized K_e values existed for both row crops and perennial crops, Equation (13) was used to generate the first approximation of effective hydraulic conductivity (K_{appr}) for each event under meadow conditions. The mean optimized K_e and mean generated K_{appr} on the 7 sites were used to develop the following adjustment equation

$$K_e = 1.81 * K_{appr} \tag{14}$$

where K_{appr} is in mm/h and can be replaced by Eq. (13)

$$K_e = 1.81\left(K_{bare}(1 - \text{scovef}) + (0.0534 + 0.01179 * K_b) * \text{rain} * \text{scovef}\right) \tag{15}$$

This final adjustment equation shows that with identical effective surface cover (scovef) the K_e of perennial crops is approximately 1.8 times higher than that from the corresponding row cropped conditions. This is due to the fact that perennial crops, often accompanied by the formation of a thick layer of organic matter or plant residue on soil surface, are more effective in improving soil aggregation, controlling soil crusting, and forming and preserving biopores.

As is shown in Table 9, the optimized K_e and predicted K_e matched well. The coefficient of determination for predicted mean K_e versus the optimized values was 0.90 and slope of regression was 0.96. The total runoff from the selected events was also predicted well. The r^2 and slope of regression were 0.94 and 0.99, respectively. Model efficiency, calculated on an event basis, averaged 0.49 (Table 9), indicating the individual storm runoff was predicted reasonably well. The predicted and measured annual runoff was plotted in Fig. 5. Linear regression fit the data well (r^2=0.76) with little bias (slope=0.88).

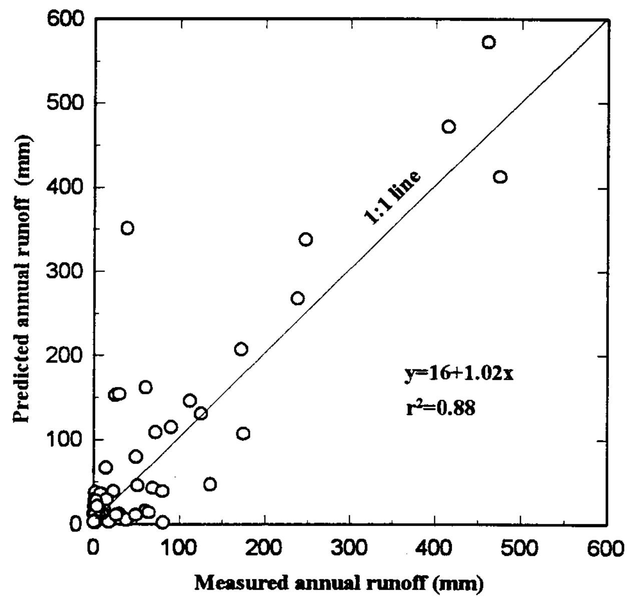

Fig. 5. Measured vs. predicted annual runoff for the data used under meadow conditions.

Table 9. Comparison of optimized and predicted effective conductivity, K_e, and measured and predicted total runoff for the events selected in the years when meadow was grown.

Site	Crop management	Total rainfall mm	K_e*		Total runoff		Model efficiency
			Opt. mm/h	Pred. mm/h	Meas mm	Pred. mm	
Holly Springs	bermuda-corn-bermuda	3497	1.62	2.49	1196	1256	.675
Morris	brome grass-corn-oats	646	12.15	17.55	27	17	.649
Watkinsville	corn-bermuda-bermuda	1682	11.87	13.36	154	269	.573
Bethany	cont. alfalfa	2900	6.40	4.98	310	553	.293
	cont. brome grass	2761	5.47	5.90	308	265	.466
Geneva	cont. red clover	549	6.71	6.54	35	54	xx
	cont. brome grass	1131	10.23	7.97	3	93	xx
Guthrie	cont. bermuda grass	3767	19.55	20.30	189	373	.734
	wheat-clover-cotton	1270	13.81	22.19	112	145	.275

* Means of all selected events. Calculated on an event basis.

Time-Invariant Effective Hydraulic Conductivity Values

For the case of time-invariant effective conductivity, the input value of K_e must represent both the soil type and the management practice. This method is corollary to the curve number approach for predicting runoff, and in fact, the estimation procedures discussed here were derived using curve number optimizations, so the runoff volumes predicted should correspond closely to curve number predictions. One difference between this method and the curve number method is that no soil moisture correction is necessary, since WEPP takes into account moisture differences via internal adjustments to the wetting front matric potential term of the Green-Ampt equation.

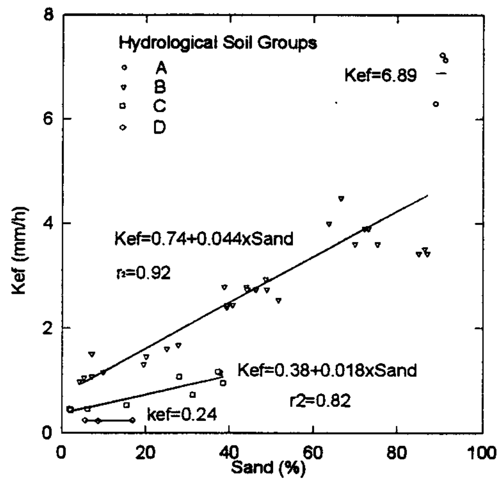

Fig. 6. Optimized effective conductivity values for fallow soil conditions, K_{ef}, plotted versus sand content of the soil. This is for the case of time-invariant conductivity.

The estimation procedure involves two steps. In step one a fallow soil K_e (K_{ef}) is calculated. In step 2 the fallow soil K_{ef} is adjusted based on management practice using a runoff ratio to obtain the input value of K_e.

Step 1: K_{ef} was found to be related to the amount of sand in the upper 20 cm of the soil profile (Fig. 6). Thus, one may use the hydrologic soil group and sand content to estimate K_{ef} (mm/hr):

Hydrologic Soil Group	Formula
A	$K_{ef} = 14.2$
B	$K_{ef} = 1.17 + 0.072(\text{Sand})$
C	$K_{ef} = 0.50 + 0.032(\text{Sand})$
D	$K_{ef} = 0.34$

Step 2: Multiply K_{ef} by the value in the table below to obtain K_e (mm/hr):

	Hydrologic Soil Group		
	A	B,C	D
Fallow	1.00	1.00	1.00
Conv. Tillage - Corn	1.35	1.58	1.73
Conv. Tillage - Soybeans	1.39	1.70	2.00
Conserv. Till. - Corn	1.48	1.79	2.21
Conserv. Till. - Soybeans	1.50	1.91	2.49
Small Grain	1.84	2.14	2.48
Alfalfa	2.86	3.75	6.23
Pasture (Grazed)	3.66	4.34	5.96
Meadow (Grass)	6.33	9.03	15.5

For other cases, such as for crop rotations, ratios of K_e/K_{ef} may be estimated from curve number values using the equation:

$$K_e = 56.82K_{ef}^{0.286}/\,(1 + 0.051e^{0.06CN}) - 2 \qquad (16)$$

Table 10 shows the model results as applied to data from fallow natural runoff plots. The tests indicate that this method gives a slightly better fit to the measured data than does the curve number method, as evidenced by the greater event-by-event model efficiencies. Tables 11 and 12 show the model results as applied to data from several cropped natural runoff plots. In Table 11, Eq. (16) was used to estimate K_e, whereas the ratio values listed above for the 7 management practices were used in Table 12. WEPP produced better model efficiencies for most of the applications than did the curve number procedure.

Table 10. Measured runoff volumes, curve number and WEPP predicted runoff volumes, and model efficiencies for the fallow runoff plot data.

Site	Average runoff per event			Model efficiency	
	Measured mm	CN mm	WEPP mm	CN	WEPP
Bethany, MO	11.47	11.87	14.41	0.10	0.11
Geneva, NY	7.87	6.08	6.45	0.58	0.63
Guthrie, OK	10.91	10.58	12.46	0.77	0.80
Holly Springs, MS	15.17	12.62	13.58	0.79	0.84
Madison, SD	7.96	6.72	9.40	0.69	0.70
Morris, MN	5.55	8.77	10.94	-1.06	-1.20
Pendleton, OR	3.18	1.87	1.24	-0.26	-0.08
Presque Isle, ME	6.91	4.87	4.86	-0.25	0.18
Tifton, GA	19.58	21.70	21.17	0.36	0.43
Watkinsville, GA	13.42	11.98	13.41	0.75	0.83

Table 11. Measured runoff volumes, curve number and WEPP predicted runoff volumes, and model efficiency for the cropped runoff plot data. The estimations of effective conductivity are from the use of Eq. (16).

Site	Management Practice	Average runoff per event			Model efficiency	
		Measured mm	CN mm	WEPP mm	CN	WEPP
Bethany, MO	Alfalfa	3.72	1.25	1.41	0.33	0.49
	Blue grass	3.91	1.30	1.28	0.43	0.42
	Corn	12.20	6.65	7.63	0.66	0.73
Guthrie, OK	Blue grass	1.94	2.04	4.88	0.58	0.32
	Cotton	8.85	9.03	14.21	0.68	0.49
Holly Springs, MS	Corn	11.00	11.91	12.01	0.15	0.38
Madison, SD	Corn	6.70	4.90	6.07	0.55	0.78
	No-till corn	6.22	3.57	4.75	0.50	0.76
Watkinsville, GA	Corn	6.96	9.97	14.15	0.37	0.04
	Cotton	7.48	8.91	12.22	0.49	0.09

Table 12. Measured runoff volumes, curve number and WEPP predicted runoff volumes, and model efficiency for the cropped runoff plot data. The estimations of effective conductivity are from the use of tabulated values in the text.

Site	Management Practice	Average runoff per event			Model efficiency	
		Measured mm	CN mm	WEPP mm	CN	WEPP
Bethany, MO	Alfalfa	3.72	1.25	1.46	0.33	0.50
	Blue grass	3.91	1.30	1.33	0.43	0.43
	Corn	12.2	6.65	7.55	0.66	0.72
Guthrie, OK	Blue grass	1.94	2.04	2.74	0.58	0.80
	Cotton	8.85	9.03	11.5	0.68	0.68
Holly Springs, MS	Corn	11.0	11.91	13.35	0.15	0.29
Madison, SD	Corn	6.70	4.90	7.56	0.55	0.70
	No-till corn	6.22	3.57	5.91	0.5	0.76
Watkinsville, GA	Corn	6.96	9.97	11.4	0.37	0.37
	Cotton	7.48	8.91	10.5	0.49	0.55

Bio-pores Adjustments to Effective Conductivity

Accounting for infiltration differences as a function of wormholes may be made by adjusting the input value of effective conductivity. The suggestions listed here are preliminary guidelines which are based on interpretations of personal communications regarding the effects of biopores on permeability classes from the SCS Soil Survey Laboratory Staff. The first step is to identify the biopore influence class from Table 13 below. Then, the input value of either K_e or K_b as calculated above should be multiplied by the ratio shown in Table 14 below.

Table 13. Classes of biopore influence defined by abundance and size classes.

	Pore Size		
Abundance	Medium	Coarse	Very Coarse
Few	Small	Moderate	Moderately Large
Common	Moderate	Moderately Large	Large
Many	Moderately Large	Large	Very Large

Table 14. Increase in Input K_e or K_b by biopore influence

Input K_e,K_b	Biopore Influence	Ratio for K_e, K_b Increase
Very Low < 0.5	Moderate	12
	Large	15
	Very Large	18
Low 0.5 - 1	Moderate	9
	Large	12
	Very Large	15
Moderately Low 1 - 2	Moderate	6
	Large	9
	Very Large	12
Moderate 2 - 3	Moderate	3
	Large	6
	Very Large	9
Moderately High 3 - 5	Moderate	2
	Large	2.5
	Very Large	3

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Small Watershed Studies: Infiltration Data and Research Needs

G.W. Frasier[1]

Abstract

Infiltration data from small watersheds is usually presented as a difference of the precipitation falling and the runoff from the area. There have been limited data sets collected of point infiltration measurements but essentially no effort to relate the point data to the entire watershed.

Introduction

Water infiltration in a given area can be considered in 2 contexts, the total water infiltrated and the rate of water infiltration. The **total** quantity of water infiltrated into a field or small watershed can be estimated using mass balance techniques (quantity on minus quantity off, ie. total rainfall minus total runoff). The infiltration **rate** on a watershed or field has both a spatial and temporal variability and becomes increasingly difficult to quantify as the size of the area increases. To better understand and model the infiltration process it is usually necessary to measure the rate of infiltration at specific or representative locations. This data is then integrated into an average infiltration rate over the entire area or watershed. The suitability of this approach depends upon how representative the evaluated sites are to the entire area.

Total Infiltration

In practice most watershed data is reported as a quantity of runoff for a given precipitation event. The difference between the two values is the total quantity of water that is infiltrated. Total runoff and precipitation data has been tabulated for many of the ARS instrumented watersheds and is available through the specific watershed research units. Some of the data has been compiled and tabulated by the USDA-ARS Hydrology Lab, Rm. 104, Bldg. 007, BARC-West, Beltsville, MD. As a first approximation the difference between the rainfall and runoff can be considered as total infiltration. This process does not provide any information as to where on the watershed the water has infiltrated. In many sites water may run off the upland sites (ie. relatively low infiltration) and percolate into the dry channel. As a result there is a low total volume of runoff but all the infiltration is occurring in a relatively small , and specific, area of the watershed. This phenomena is a common occurrence in the ephemeral streams of the southwestern portion of the United States and other arid and semiarid regions of the world (Lane 1982, 1983).

There is a potential bias in assuming all the water that does not run off has infiltrated into the soil. In most places there is a measurable and maybe significant portion of the precipitation which is intercepted by the vegetation and evaporated directly back into the atmosphere.

Total infiltration data does not provide any information concerning the location within the

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[1]USDA-ARS, RRRU, 1701 Center Ave, Ft. Collins, Colorado 80526

field/watershed area where infiltration is occurring, the rate of water infiltration, or even when the infiltration is taking place. On these areas it is very difficult to develop any relationships among site parameters ie. soil type, vegetation, slope, surface roughness and the total infiltration volume. As a result these data provide very limited information suitable for the development of process oriented models of infiltration and runoff.

Infiltration Rate

Infiltration rate data has been collected at many locations using techniques such as auger holes or ring infiltrometers. While this data is very useful in assessing the rate that water will penetrate the soil, it does have a bias of only representing a relatively small area, usually less than 1 square meter. Spatial variability of soil properties, ie. texture, organic matter, morphology, both laterally and vertically can significantly affect the magnitude of the measured infiltration rate. Most ring infiltration data was collected for specific studies with little effort in characterizing larger areas such as entire fields or small watersheds.

Various ARS locations were contacted to determine the availability of infiltrometer data sets that could be used in model development or validation. Information requested was: (1) the type of information (data bases) available; (2) how the data was collected; (3) extent of data (time and spatial); and (4) limitations of the data set. Also requested was a general description of the various soil parameters, such as texture, bulk density, etc. The emphasize was on field measured infiltration rates over time, initial and equilibrium. It was anticipated that the most important data would probably come from small plots (with uniform features) or infiltrometer (ie. ring) tests.

Unfortunately, the researchers at many of the locations who collected ring infiltrometer data have either retired or otherwise left and the data is no longer available. In some locations, once the researcher left the original data notes have been lost or destroyed.

Attached are 5 sets of tabulations representing 9 data sets which were compiled as a result of the request. These data sets represent the general type of data which is currently being collected. It is believed there are data sets of a similar nature being collected at other locations, probably as support information for other studies.

One concern with standard ring infiltration data is that the infiltration rate is measured with a head of water ponded on the soil surface, frequently at depths of 30 cm or more. These depths are not representative of conditions when the water is allowed to flow over the land with average depths in the order of a few millimeters. Techniques using falling head ring infiltrometer can provide a closer estimate of infiltration rates with very low water depths. The techniques developed for negative head permeameters does permit estimation of infiltration rates at zero or shallow water depths.

Future Research Needs

There is a need to develop a database of the various infiltration measurements that have been collected in the past. There are already instances of data lost because the researcher has retired or moved to other locations or assignments. There are problems of spatial variability when using point measurements. Techniques are needed to scale the infiltration results of point measurements to larger areas, especially on rangelands where there is a random mix of bare soil (interspaces) and vegetated areas often coupled with surface elevation differences (micro-topography). We frequently find the vegetated areas on higher portions of the soil surface with the bare spaces forming micro-channels down the slope. Preliminary studies have shown the infiltration rate in the vegetated areas may be several orders of magnitude greater in the vegetated areas than in the bare interspace areas. At the present time, mathematical representation of the micro-topography coupled with the spatial variability of the infiltration rate has not been developed.

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Data Set Contributors

J.A. Daniel, USDA-ARS, Durant, Oklahoma
W. E. Emmerich, USDA-ARS, Tucson, Arizona
R.K. Hubbard/D. Bosch, USDA-ARS, Tifton, Georgia
M.A. Weltz, USDA-ARS, Tucson, Arizona

INFILTRATION DATA BASE

LOCATION: Southwest Watershed Res. Ctr

CONTACT PERSON: William Emmerich

USDA-ARS, 2000 East Allen Rd, Tucson, AZ 85719

Tel (520) 670-6380, FAX (520) 670-5550

TYPE	STYLE	SITE USAGE	MEASUREMENT PERIOD		NUMBER OF MEASURE	SOIL PARAMETERS							DATA STORAGE	COMMENTS
			DURATION	FREQUENCY		TYPE	TEXTURE		BULK DENSITY		SOIL MOISTURE			
							DEPTH	INCREMENT	DEPTH	INCREMENT	DEPTH	INCREMENT		
Disc Permeameter 22 cm	Negative Head 10,7,5, 3,1 cm	Rangeland (See note 3)	Summer 94 - Spring 97	Each Season (4)	6 Reps/ Season	See note 1	-------------	-------------	-------------	-------------	-------------	-------------	Computer Spread Sheet-Quatro Pro	See Note 2
Disc Permeameter 22 cm	-5 cm tension +7 cm positive	Rangeland	1991-1993	Periodically	3-4 Saturated & Unsat Each time		-------------	-------------	0-5 cm	0-5 cm	0-5 cm	0-5 cm	Computer Spread Sheet-Quatro Pro	See Note 4

FOOTNOTES

Note 1 Texture and bulk density will be measured at a later date

Note 2 Measurements made in interspaces of a brush dominated site with >50% erosion pavement

Note 3 Treatments----- (1) Control--erosion pavement present; (2) Erosion pavement removed;

(3) Erosion pavement removed for 1 season; (4) Erosion pavement removed for 1 yr.

Spatial Variability-----Each season 6 random measure. on 6 ha.

Note 4 Measurements beside simulator plots at time of simulations

INFILTRATION DATA BASE

LOCATION: Southeast Watershed Res. Lab

CONTACT PERSON: David Bosch E-Mail dbosch@tifton.cpes.peachnet.edu
USDA-ARS, P.O. Box 946, Tifton, GA 31793
Tel (912) 386-3899, FAX (912) 386-7294

TYPE	STYLE	SITE USAGE	MEASUREMENT PERIOD		NUMBER OF MEASURE	SOIL PARAMETERS							DATA STORAGE	COMMENTS
			DURATION	FREQUENCY		TYPE	TEXTURE		BULK DENSITY		SOIL MOISTURE			
							DEPTH	INCREMENT	DEPTH	INCREMENT	DEPTH	INCREMENT		
Auger Holes 3.2 cm	Constant Head Perm.	Cropland Irrigated corn	1995	One time	57	Sandy Loam	0-250 cm	12 cm	0-250 cm	12 cm	0-250 cm	12 cm	Computer Spread Sheet (quatro pro)	

FOOTNOTES

INFILTRATION DATA BASE

LOCATION: Many Places Throughout U.S.

CONTACT PERSON: Mark Weltz E-mail weltz@tucson.ars.ag.gov

USDA-ARS, 2000 East Allen Road, Tuscon, AZ 85719

Tel (520) 670-6380, FAX (520) 670-5550

TYPE	STYLE	SITE USAGE	MEASUREMENT PERIOD		NUMBER OF MEASURE	SOIL PARAMETERS								DATA STORAGE	COMMENTS
			DURATION	FREQUENCY		TYPE	TEXTURE		BULK DENSITY		SOIL MOISTURE			
							DEPTH	INCREMENT	DEPTH	INCREMENT	DEPTH	INCREMENT		
Rainfall Simulator	Mass Balance	Various Thoughtout Western U.S. Rangeland	1987-1993	One time only for most; Some revisted a second year	59 sites	Various	By profile See note 2	By profile	By profile	By profile	By profile	By profile	Foxpro Relational Database	

FOOTNOTES

Note 1 Data was collected as a part of the USDA-IRWET (Interagency Rangeland Water Erosion Team) rangeland rainfall simulation experiments used to develop the Water Erosion Prediction Project (WEPP) model. IRWET is comprised of ARS staff from the Southwest and Northwest Watershed Research Centers, Tucson, AZ and Boise, ID, NRCS staff members in Lincoln, NE and Bosie, ID, and US Forest Service staff in Flagstaff, AZ.

Note 2 Members of the NRCS staff in Lincoln, NE completed a full soil profile, including physical and chemical characteristics, for each site.

INFILTRATION DATA BASE

LOCATION: Misc. Colorado

CONTACT PERSON: Gary Frasier E-Mail gfrasier@lamar.colostate.edu

USDA-ARS, 1701 Center, Ft. Collins, Colo 80526

Tel (970) 498-4232; FAX (970) 482-2909

TYPE	STYLE	SITE USAGE	MEASUREMENT PERIOD		NUMBER OF MEASURE	SOIL PARAMETERS							DATA STORAGE	COMMENTS
						TYPE	TEXTURE		BULK DENSITY		SOIL MOISTURE			
			DURATION	FREQUENCY			DEPTH	INCREMENT	DEPTH	INCREMENT	DEPTH	INCREMENT		
Double Ring	Falling Head	Sheep Creek (See Note 1)	Summer 1996	2 times during summer	3 reps per paddock See Note 2	Clay Loam	To Be Measured	To Be Measured	To Be Measured	To Be Measured	To Be Measured	To Be Measured	Quatro Pro Spread Sheet	Data available after 1996
Disk Permeameter	Tension -3, 6, 12 cm	CPER (See Note 3)	Summer 1994	1 time	5-7 reps See Note 4	Data Available	Data Available	Data Available	Data Available	Data Available	Data Available	Data Available	Quatro Pro Spread Sheet	Data available after 1996

FOOTNOTES

Note 1 Sheep Creek; Riparian; grazing

Note 2 2 grazing treatments (time of year); 3 paddocks per treatment; 3 measurements per paddock

Note 3 Central Plains Experimental Range; Shortgrass rangelands; grazing

Note 4 2 grazing treatments; 2 plots (rainfall simulator) per treatment ; measurements in both vegetative clumps and interspaces

INFILTRATION DATA BASE

LOCATION: Misc Oklahoma

CONTACT PERSON: John Daniel

USDA-ARS, 801 Wilson St., P.O. Box 1430, Durant, OK 74702

Tel (405) 924-5307; FAX (405) 924-5066

TYPE	STYLE	SITE USAGE	MEASUREMENT PERIOD		NUMBER OF MEASURE	SOIL PARAMETERS							DATA STORAGE	COMMENTS
			DURATION	FREQUENCY		TYPE	TEXTURE		BULK DENSITY		SOIL MOISTURE			
							DEPTH	INCREMENT	DEPTH	INCREMENT	DEPTH	INCREMENT		
Double Ring	Falling Head	El Reno/FR-5	1987	One time only	1 Field	Silty Loam	0-15 cm	15 cm	0-15 cm	15 cm	-----	-----	Data Sheet and	Water impounded, hook
30 cm Inner		Rangeland			20 Sites	" "	15-30 cm	15 cm	15-30 cm	15 cm	------	-----	publications	gauge used at 1, 5, 15
90 cm Outer		Grass Covered				" "	30-45 cm	15 cm	30-45 cm	15 cm	------	-----		min and 2 hr interval
Double Ring	Falling Head	Lane, OK	1987 & 1989	One time only	1 Field	Loamy Sand	0-15 cm	15 cm	0-15 cm	15 cm	-----	-----	Computer Spread	Water impounded, hook
30 cm Inner		Grass			8 Sites	Silty Loam	15-30 cm	15 cm	15-30 cm	15 cm	------	-----	Sheet & Field	gauge used at 1, 5, 15
90 cm Outer		Covered				Silty Loam	30-45 cm	15 cm	30-45 cm	15 cm	------	-----	Notebook	min and 2 hr interval
Double Ring	Falling Head	Knox City PMC	1991	One time only	1 Field	Sandy Loam	0-15 cm	15 cm	7.5 cm	15 cm	-----	-----	Spreadsheet &	Water impounded, hook
30 cm Inner		Vegetated			10 Sites						------	-----	Log Books	gauge used at 1, 5, 15
90 cm Outer		Lysmeter Plots									------	-----		min and 2 hr interval

FOOTNOTES

Infiltration Parameter Calibration using Rainfall-Runoff Data

Roger E. Smith[1]

Parameters for any modeled system may be determined from measured input and output data by running an appropriate model of the system and changing parameter values until the model and the measurements agree. This is called parameter *calibration*, or an *inverse* method. It is especially useful when parameters are not physically measurable, or when the parameter represents the effective mean of a spatially variable value.

Calibration may be used to estimate infiltration parameter values when an experimental plot or small catchment provides reliable rainfall and runoff data to which results from a runoff simulation model may be compared. The runoff dynamics should not be ignored, since there are always time delays between creation of rainfall excess and the appearance of that water at the outlet edge of a plot, however small. Good calibration requires good data, not only having accurate rates of rainfall and runoff, but also having coincidence of timing of both records. It is desirable to have a calibration event covering a relatively long period of runoff. Calibration with a variety of storm types is also desirable, although it may not be possible to fit all results equally well with the same parameters.

Infiltrometer or plot data are measurements of runoff through a measuring flume or other outlet device, and the rate of change of storage on the plot surface and within the flume must be taken into account. This requires a model which accurately treats the surface storage dynamics. For this discussion we will assume kinematic wave surface hydraulics, in which the surface storage is controlled by slope and a roughness coefficient. One can show that in calibration, the infiltration parameters interact with the roughness of the surface as well as with each other. Luce and Candy (1994) provide a good example of calibration problems.

Field Data Quality Analysis

There are several important points to consider in judging whether experimental data is likely to provide good calibrated parameters. Large catchments inherently will contain much water in storage during runoff. Using such data to calibrate for infiltration parameters is very difficult. Even using small plot data for infiltration calibration requires understanding that there may be on the order of a minute delay between the actual onset of runoff and the appearance of water in a measuring flume record. A minute may significantly affect parameter values if the time to ponding is only a few minutes from the start of rainfall. This can lead to significant parameter errors.

If runoff is measured by a flume or weir, there may also a significant backwater storage involved. For each discharge rate through the measuring device, there is an associated depth of water which is measured, called a *control depth*, and for each control depth, there is a volume of water in storage behind the device. This storage may be referred to as *backwater storage*. During any time interval when flow is increasing, some of the water that flows into the device,

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[1] USDA-ARS, Water Management Res., AERC Foothills Campus, CSU, Fort Collins, CO 80523

such as a weir, will be used to increase the storage. There is a storage associated with each discharge rate for the device. Accounting for the storage increase during hydrograph rise, and storage loss during recession, is called pondage correction or *derouting*. Derouting must be undertaken if the rate of inflow to the weir is to be estimated. Derouting can be accomplished by solution through time of a linear differential equation which relates inflow rate, storage change and measured outflow, and thus derive the time pattern of inflow. Referring to the inflow to the backwater pond as q_i, the stored volume as V, and the device outflow as q_o, the volume balance is

$$q_i(t)=\frac{dV(h)}{dt}-q_o(h) \tag{1}$$

Depth h determines not only the outflow q_o through the device rating [$q_o(h)$], but also determines the storage V(h). dV/dt can be written A(h)dh/dt, and since surface area A(h) can be determined from the geometry of the flume and the topography behind the flume or weir, the discharge of interest, q_i, can be found be solving the equation

$$q_i(t)=A(h)\frac{dh}{dt}+q_o(h) \tag{2}$$

using that data and the record of measured h(t).

The importance of timing coincidence of rainfall and runoff records cannot be overemphasized. Figure 1 illustrates a clear case of this problem. In this example, the sharp runoff peak physically must be associated with a significant drop in the rate of rainfall excess, yet it comes about one minute before this could have occurred for the rainfall hyetograph shown. Many other such errors in hydrogaph-runoff data would not be so obvious. Fitting infiltration and soil without recognizing the inherent error in timing would result in extremely biased parameters.

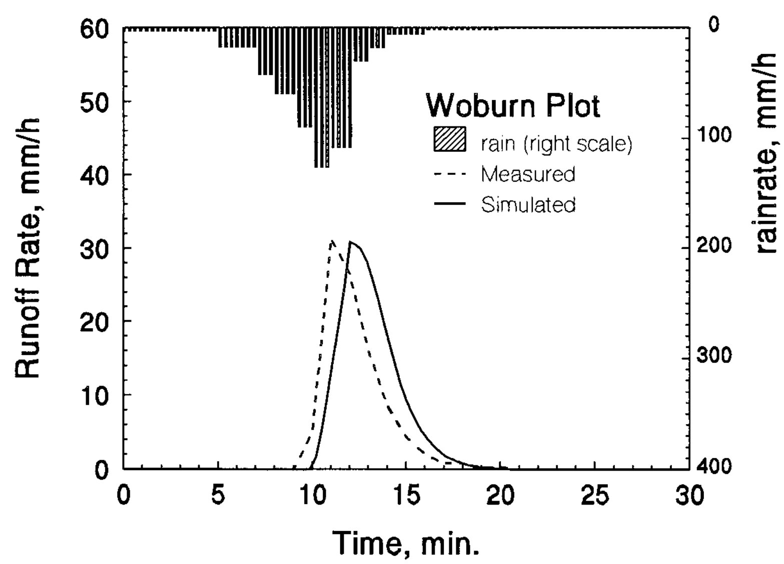

Fig. 1. Example of timing error in data.

Infiltration Parameters

There are two fundamental infiltration parameters, (plus the variable initial soil moisture state), which play a role in most physically-related infiltration equations. Much of the calibration factors discussed here apply to determining parameters of empirical equations as well. The first is saturated hydraulic conductivity, K_s [L/T]. It is a scaling parameter for rainfall rate, r, and runoff will not occur for values of $r < K_s$, or $r^* = r/Ks < 1$.

The other fundamental infiltration parameter is variously called the *capillary drive*, or the *capillary length scale*. Here I will use the symbol G for this parameter. It is really a relative conductivity-weighted integral measure of soil capillary head, h. Relative conductivity, k_r, is defined as $K(h)/K_s$, and G may be calculated from the soil's K(h) relation (if known) as

$$G=\int_{-\infty}^{0} k_r dh \tag{3}$$

Without presenting details of various infiltration equations and their derivation, it is useful here to indicate the role these parameters play in common functions. The above defined parameters appear in the Green-Ampt infiltration equation (expressed as f(I), where I is infiltrated depth [L]):

$$f_c=K_s\frac{G\Delta\theta+I}{I} \tag{4}$$

where $\Delta\theta$ is soil saturation deficit, saturated water content minus initial water content. The same parameters appear in the Smith-Parlange equation:

$$f_c=\frac{K_s}{1-\exp(-I/G\Delta\theta)} \tag{5}$$

or in Philip's equation, in terms of time:

$$f_c=K_s+\sqrt{\frac{G\Delta\theta K_s}{2t}} \tag{6}$$

Scaled relationships are useful in finding infiltration parameters, which allow an analysis of the experiment in terms of the overall position of *f* on the curve. For this purpose, one can define

$$f_* = \frac{f_c}{K_s}$$

$$I_* = \frac{I}{G\Delta\theta} \qquad (7)$$

$$t_* = \frac{tK_s}{G\Delta\theta}$$

which reduces Eq. 4, for example, to

$$f_* = \frac{1+I_*}{I_*} \qquad (8)$$

Figure 2 illustrates the general scaled infiltration relation. At small times all physically-based infiltration relations are asymptotic to the relations

$$f_* = \frac{1}{I_*} \qquad (9)$$

or

$$f_* = \sqrt{\frac{1}{2t_*}} \qquad (10)$$

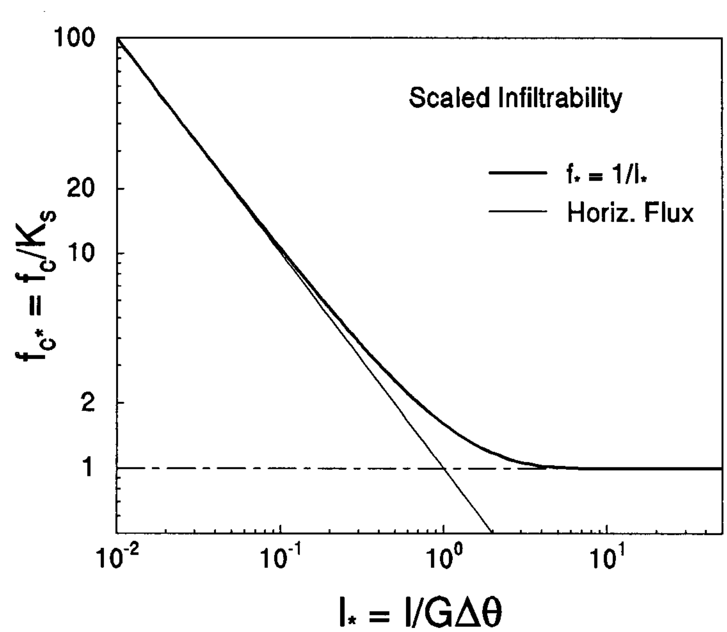

Fig. 2. Illustration of the scaled infiltration relation.

The unscaled version of Eq. 9 is $f_c = Ks(G\Delta\theta)/I$. Thus for short intense storms, the two parameters act as a pair and are indistinguishable. This means that for storms during which this relation holds, the time to start of runoff can as easily be fitted by an adjustment of G as an adjustment of K_s. For a longer storm which carries the infiltration capacity relation into the asymptotic region (the curved area and beyond) it is possible to fit both G and K_s For such an event, K_s will have the most significant effect on runoff amount and rate later in the storm and G may be independently adjusted to match the time of inception of runoff. These parameter characteristics are illustrated in Figs. 3 and 4.

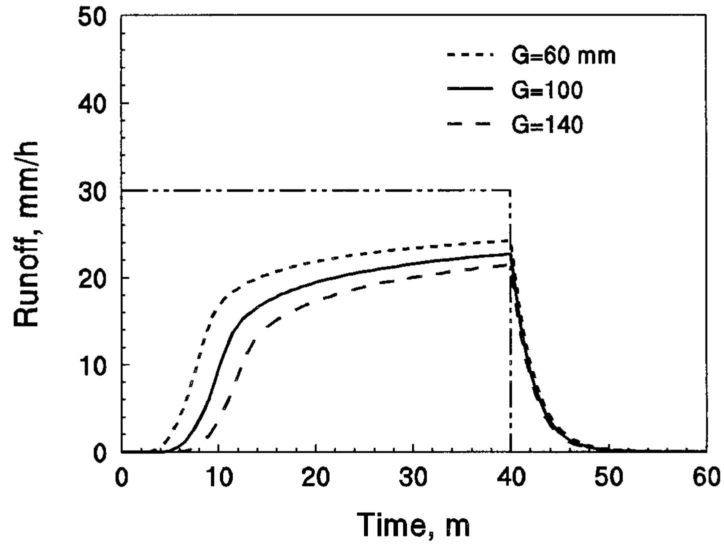

Fig. 3. Effect of G on plot runoff.

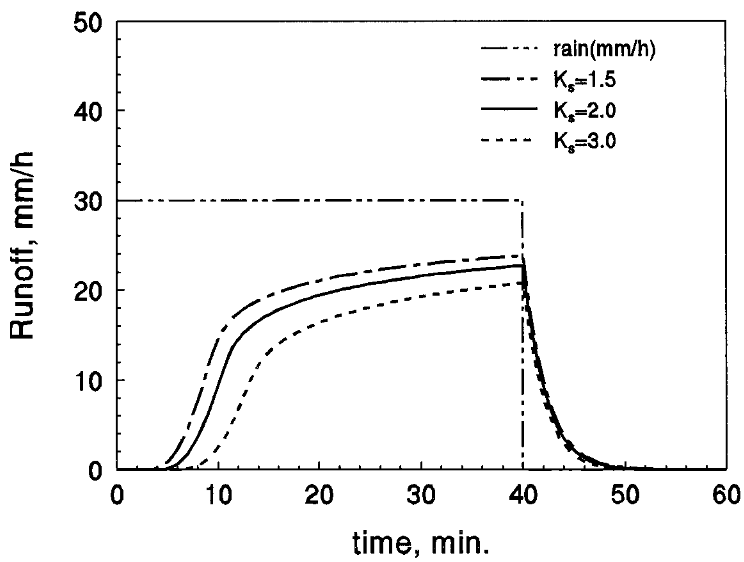

Fig. 4. Effect of K_s on longer hydrograph.

Parameters G and K_s have very similar effects even for these relatively long steady rains. Care must be taken to look independently at both timing and volume of runoff, because there are more than one combination of these two parameters which will match a given amount of runoff. The volume of runoff for an event can be matched by a given level of Ks and a fitted G, or, for example, by a lower Ks (giving a higher level of runoff) and using G to force a later start of runoff. In any case, smaller runoff events are better for fitting infiltration parameters, since larger storms are less sensitive to infiltration. Figure 5 illustrates the parameter interaction indicated by Eq. 9 for a short storm.

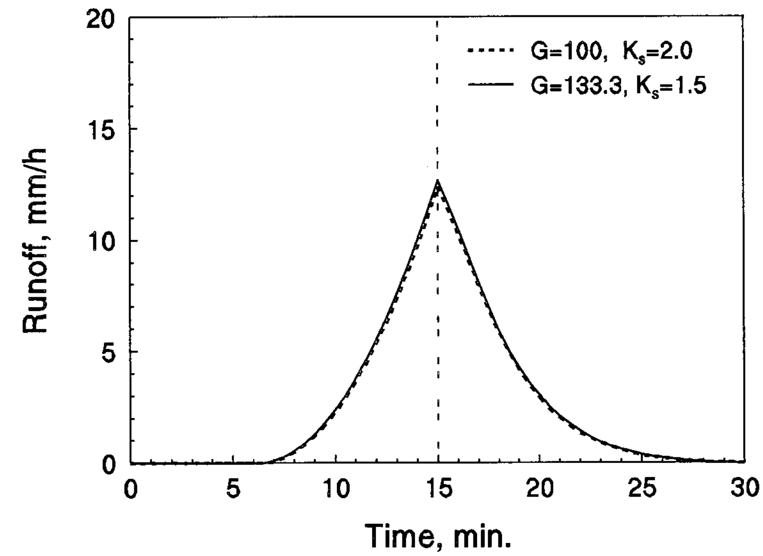

Fig. 5. Interaction of parameters G and K_s in short storms.

Complicating Factors

It is common to expect or assume the value of f after a one or two hour rain to be close to or equal the value of K_s. For this purpose the scaled infiltration curve can reveal whether this is a realistic assumption. Even approximate values of G and K_s can be used to estimate the value of f_* and thus how near K_s one should expect to be. The complicating factors of air counterflow (most critical for fine soils and high application rates) and soil layers can confound the calibration, of course, and often can cause a flattening of the $f(t)$ relation far in advance of the predictions of the above relations. The amount of information in rainfall/runoff data is insufficient to calibrate a relation for more complicated conditions.

Hydraulic Roughness

Figure 6 illustrates the basic effect that changing values of n will have on a simple plot outflow with constant rainfall rate. The dashed line shows rainfall excess, the rate at which runoff is produced in place. The difference between this curve and the runoff rate (measured at the bottom of this hypothetical plot) is the rate at which water is going into storage. Likewise the area between the two curves up to any time T is the volume of storage V(T) on the plot surface (assuming rainfall excess, r(t)-f(t), is uniform):

$$V(T)=\int_0^T [r(t)-f(t)-q(t)]dt \tag{11}$$

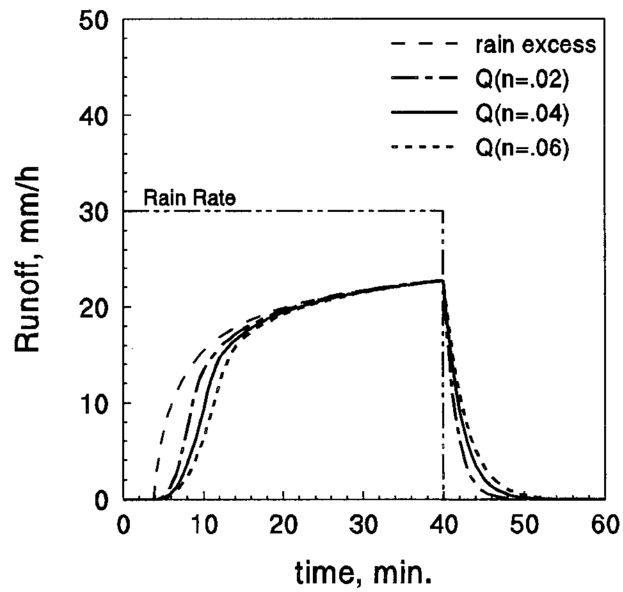

Fig. 6. Effect of roughness on surface storage.

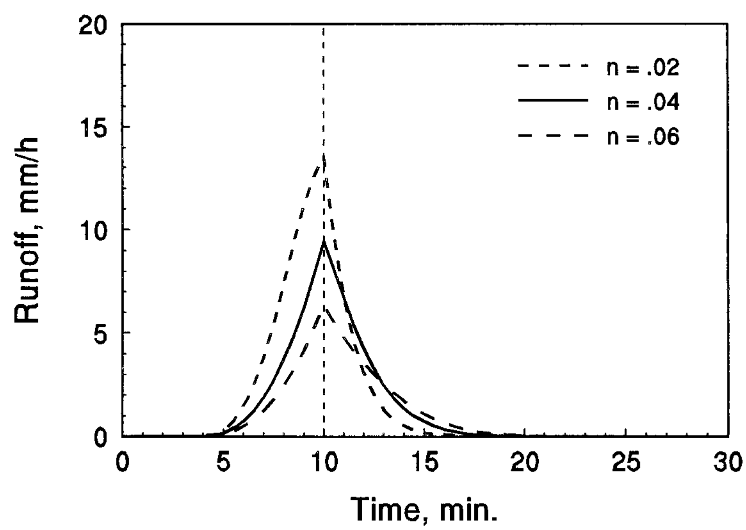

Fig. 7. Effect of roughness for short storm.

In contrast to the situation in Fig. 6, Fig. 7 shows that the hydraulic roughness is important for estimating the peak runoff of short rainfall events. Changing the amount of runoff which must go into storage can have a significant effect on the peak for such short events. It can also be seen from these figures that for short storms there is a subtle interaction between the infiltration curve, which was assumed known in Fig. 6, and the roughness, since each can affect the shape of the rising hydrograph and the peak runoff for a short rain. This is another reason why a longer rainfall is far superior to a short one for purposes of parameter estimation. While changes in (K_sG) or n both have an effect on the shape of the rising portion of the hydrograph, later in the storm the effects of infiltration are significantly different from the effects of roughness. Complexity of a long storm is not a problem when an appropriate runoff routing model is used, since K_s can be adjusted to match the volume, and roughness used to calibrate the peaks produced by any later bursts of intense rain.

Effect of Microtopography: The recession after a rain or a plot experiment can be effected by microtopography, which confines flow to an ever decreasing portion of the area as recession proceeds. This causes a reduction in gross infiltration, and lengthens the recession, as illustrated in Fig. 8. While it is not important in estimating the infiltration parameters, simulating this effect in a runoff model can be used to achieve improved fits of recession, also affecting total runoff volume, especially in runoff from quite rough or undulating surfaces.

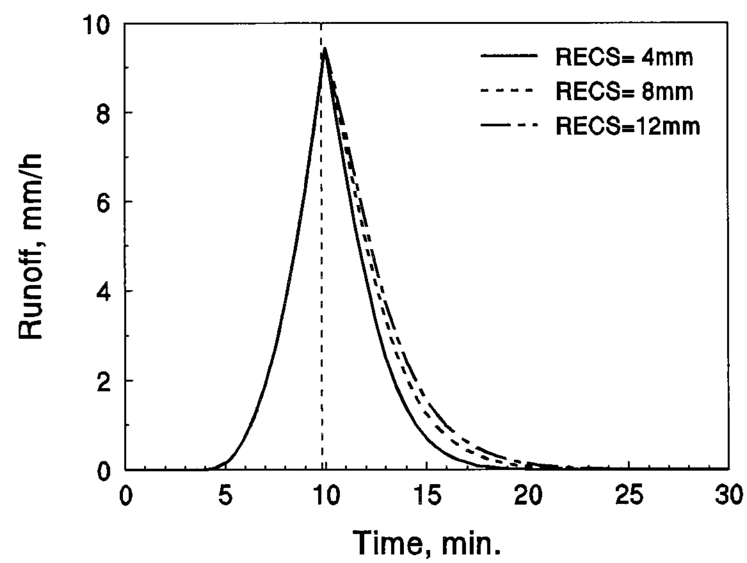

Fig. 8. Effect of microtopographic relief parameter on recession.

Comparative Sensitivity: The results above indicate the interaction of parameters G and K_s, and that runoff and hydrograph shape is considerably more sensitive to the infiltration parameters than to hydraulic roughness. Whatever combinations of G and K_s are used, they should not be such that the physically realistic values for these parameters are severely violated. A general guide should be to calibrate parameters in the following order: K_s, G, n, and then the microtopography (recession), if possible. In any case, several iterations may be required.

Reference

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Specific Goal 2

New Ideas for Characterizing Spatial Variability

Indirect Estimation of Hydraulic Conductivity and its Spatial Distribution

Dennis Timlin[1], Walter Rawls[2], Robert Williams[3], and Laj Ahuja[4]

Abstract

The understanding of the process of water infiltration into soil is important to be able to predict soil erosion, runoff of water and chemicals from soil, water availability to plants, movement of chemicals to groundwater, salt leaching, and groundwater recharge. Saturated hydraulic conductivity, K_{sat} [K(h) at $h \geq 0$)], is probably the most important soil hydraulic parameter for infiltration of water in soils. This parameter is difficult to obtain and is highly variable as it is highly sensitive to soil conditions, such as compaction, macropores, sample size, temperature, and entrapped air. Values for hydraulic conductivity and a measure of its variability are both important for modeling infiltration processes. In this paper we review methods used to estimate hydraulic conductivity and its variability. The methods covered include estimates using soil texture and structure, fractal properties, inverse methods, scaling and remote sensing. We also discuss future research needs in this area.

Introduction

An understanding of the infiltration of water into soil is necessary to model soil erosion, runoff of water and chemicals, water availability to plants, movement of chemicals to groundwater, groundwater recharge, and solute leaching. An important soil parameter that affects infiltration rate is the soil hydraulic conductivity. This includes the saturated hydraulic conductivity, K_{sat} [K(h) at $h \geq 0$)], and unsaturated conductivity. Hydraulic conductivity is difficult to measure in the field as it is highly sensitive to soil conditions, such as compaction, macropores, sample size, temperature, and entrapped air, and thus is highly variable. In a field, the K_{sat} may vary between one to two orders of magnitude. Agricultural managers are also becoming more interested at looking at processes at larger scales such as field and watershed. Therefore, it is important to have the ability to measure of the variability of the hydraulic properties.

Because of the large number of samples required to characterize an extensive area such as a field or watershed, methods to estimate soil hydraulic parameters from simpler data such as soil bulk density or texture, or from a small number of measurements are alternatives to extensive field-based characterization. These methods include the use of soil texture and bulk density as predictors, inverse methods that use optimization procedures to fit parameters in equations that describe water flow in soil, theoretically based equations and remote sensing. The results of using

[1]USDA-ARS, RSML, Bldg 007, Rm 008, 10300 Baltimore Ave, Beltsville, MD 20705

[2]USDA-ARS, HL, Bldg 007, Rm 104, 10300 Baltimore Ave, Beltsville, MD 20705

[3]USDA-ARS, GRL, 7207 W. Cheyenne Street, El Reno, OK 73036

[4]USDA-ARS GPSR, P.O. Box E, Ft. Collins, CO 80522

these methods, however, is not always satisfactory, especially for large scale estimates. Perfection of these methods is still a long way away.

Although it is difficult to estimate soil hydraulic conductivity with an acceptable degree of error using indirect methods, determination of the variability has been more successful. Methods to generate soil hydraulic properties from the mean and variance of readily available soil properties can be used for this purpose. Mulla (1988) showed that, while texture-based estimators of a moisture retention function are not as good as laboratory-based measurements in terms of accuracy, they gave an acceptable representation of the spatial patterns of matric suction. Springer and Cundy (1987) reported that the mean and variance of the field-based parameters, however, were not preserved by the texture-based estimates. They further reported that trends and correlation lengths for the spatial distributions of percent sand and clay were translated to texture-based parameter estimates of saturated conductivity.

The purpose of this paper is to review methods currently used to estimate soil hydraulic conductivity and its variability and discuss future research needs.

Determining Saturated Hydraulic Conductivity, K_{sat}

Scale effects, i.e., differences between measurement scales of K_{sat} and the soil data used as predictors are much more important for K_{sat} than for water retention. Because of this, it is much more meaningful to try to determine a reasonably accurate distribution of K_{sat} (e.g., mean and standard deviation) in a field, rather than highly accurate point values.

Effective porosity as a predictor of K_{sat}

Recent studies have shown that K_{sat} is strongly related to an effective porosity (ϕ_e) defined as the total porosity minus the volumetric soil water content at 33 kPa tension (Ahuja et al., 1984, 1989). The effective porosity was related to K_{sat} by a generalized Kozeny-Carman equation:

$$K_{sat} = B\,\phi_e^n \tag{1}$$

where B and n are constants. The relationship fit data well from diverse locations in Hawaii, Arizona, Oklahoma, and several states of the Southeast U.S. The correlation of this relationship for all the soils was as good as for any one soil individually, which indicates that Eq. (1) is applicable across soil types.

Recent work of Ahuja et al. (1993) has shown that an average K_{sat} of a soil profile is related to drainage of the surface soil in two days after wetting, i.e., to change in soil water content of the surface soil in two days, through Eq. (1) with different values of the parameters B and n. Thus, measurements of soil bulk density to determine θ_s and water content at several locations in a field two days after a soaking rain can provide an estimate of spatial distribution of average profile K_{sat} in the field. Ahuja et al. (1993) give details and analysis of this technique, and several examples of K_{sat} estimations.

Estimating K_{sat} based on pore-size distribution

Childs and Collis-George (1950) introduced a statistical method for calculating permeabilities of porous media. Marshall (1958) derived an equation for calculating permeability based upon the statistical assumptions of pore interaction from the pore size distribution in an isotropic porous medium. Millington and Quirk (1960, 1961) further developed the method showing explicitly the assumption of the interaction of soil pore areas and radii for a monotonic sequence of pore size classes. Assuming the Poiseuille equation applies, the Marshall equation provides an expression for calculating the hydraulic conductivity. Mualem and Dagan (1978) showed the general commonality between all statistical methods for calculating hydraulic conductivity and recommended the empirical determination of the parameters from a variety of soils. Here, we present an empirical form of the Marshall equation to obtain an approximate estimate of K_{sat} (Rawls et al., 1993):

$$K_{sat} = \left(\frac{g\rho}{8v}\right) \left(\frac{\phi^x}{k^2}\right) \sum_{i=1}^{m} (2i - 1)\, R_i^2 \qquad (2)$$

where

g = gravitational acceleration, cm s^{-1}
v = viscosity, g cm^{-1} s^{-1}
ρ = density, g cm^{-3}
R_i = average pore radius in the i^{th} porosity class, cm
ϕ = total porosity
x = pore interaction exponent
k = total of pore size classes
m = number of effective pore size classes contributing to saturated flow

Using the equivalent pore radius for the Sierpinski carpet at the i^{th} recursive level from Tyler and Wheatcraft (1990), reduces Eq. (2) to (Brakensiek et al., 1993):

$$K_{sat} = 4.41 * 10^7 \left(\frac{\phi^x}{k^2}\right) R_i^2 \qquad (3)$$

where the units of K_{sat} are cm h^{-1} and the values for v and ρ are taken at 20° C.

In Eq. (3) the exponent x may be set equal to 4/3 as proposed by Millington and Quirk (1961). The maximum pore radius, R_i, was calculated following the methodology of Tyler and Wheatcraft (1990)

$$R_i = \frac{0.148}{h_b} \qquad (4)$$

where

h_b = geometric mean soil bubbling pressure (cm).

Green and Corey (1971) proposed the following equation for estimating the *k*-value,

$$k = m \left[\theta_S / \left(\theta_S - \theta_L\right) \right] \tag{5}$$

where

θ_S = total porosity
θ_L = porosity at a lower water content, e.g., Ahuja et al., (1984) used -33kPa.
m = a constant, set equal to 12 as proposed by Marshall (1958)

The following estimation equation for *k* was also developed for relating the *k* to the fractal dimension, D,

$$k = 1.86\, D^{5.34} \quad ; \quad r^2 = 0.83 \tag{6}$$

where

D = fractal dimension of soil porosity as derived by Tyler and Wheatcraft (1990). The porosity fractal dimension can be estimated by $D = 2 - \lambda$
λ = Brooks and Corey pore size distribution index

The parameters in Eq. (3) can be estimated from readily available soil properties - for example, matrix porosity by methods given by Rawls et al. (1983); the -33 kPa water content, the Brooks and Corey bubbling pressure, and pore size index by methods presented by Rawls and Brakensiek (1985), and the fractal dimension by methods presented by Brakensiek and Rawls (1992). This empirically modified Marshall equation has not been validated but preliminary unpublished results indicated the estimates of the soil-matrix K_{sat} provided by this equation were comparable to those given by Eq. (1) with parameters obtained from measured ϕ and K_{sat} values.

Equation (2) can also be used for calculating the saturated hydraulic conductivity of macropores. The value of ϕ applied is the areal porosity of the macropores, ϕ_m, which have a radius greater than 0.2 mm. The exponent *x* is assumed to have the same value as the matrix (x = 1.333), and R_l is the radius of the largest macropore. The calculation of *k* by Eq. (5) or (6) is not appropriate for macropores because all the porosity at surface ponding contributes to K_{sat}. Rawls et al. (1993) assumed that m = 12 and calibrated *n*-values to soil properties through multiple linear regression producing the following equation

$$k = 87.37 + 74\, R_l - 52\, D \tag{7}$$

where

R_l = maximum macropore radius, cm
D = fractal dimension of the soil particles

Equation (7) fit the data quite well with an R^2 of 0.88.

Methods to Determine Unsaturated Hydraulic Conductivity Relationships, K(h) or K(θ)

Convergence-scaling hypothesis

Just as for the θ(h) relationships, a linear relationship has been observed to exist between log K and log (-h) below the air-entry value for a variety of soils (Brooks and Corey, 1964). An interesting recent finding was that the slope and intercept of this log-log linear relationship are themselves linearly related, and that this latter linear relationship is approximately unique across several soil types evaluated (Ahuja and Williams, 1991). This relationship is a result of convergence of log K vs. log (-h) lines for all soils to within a narrow band around a point (Ahuja and Williams, 1991). Utilizing this apparently unique relationship between slope and intercept of log K vs log (-h), one can easily define the K(h) curve below the air-entry value if the air-entry value is known. An air-entry value can be obtained using regression equations derived by Rawls et al. (1982). As is done for θ(h), only a single pair of measured values (K,θ) are needed to determine all the parameters. Finally, the knowledge of K_{sat} provides a closure of the complete K(h) relationship.

Saturated conductivity and the water characteristic curve

Two approaches have been used to calculate the relative hydraulic conductivity ($K_r(h) = K(h)/K_{sat}$) from the soil water characteristic curve. The first approach is a macroscopic method based on the generalized Kozeny equation (Brooks and Corey, 1964; Laliberte et al., 1968). The second approach is a statistical method originated by Childs and Collis-George (1950), refined by Marshall (1958), Millington and Quirk (1961), and Mualem (1976), among others, and evaluated by numerous investigators. Mualem (1976) provides details of both these approaches and their refinements.

Campbell (1974) derived the following expression for $K_r(h)$, based on his (Campbell) model of soil water characteristic function:

$$K_r(h) = K_{sat} \left(h_e/h\right)^{2+\frac{3}{b}} \tag{8}$$

where *b* is a constant. A similar expression will result from employing a Brooks and Corey (1964) model of the soil water characteristic, with the residual soil water content, θ_r, assumed to be a non-zero constant. Employing this formulation for θ(h) and using the Mualem (1976) variation of the statistical approach, van Genuchten (1980) derived the following equation for $K_r(\theta)$:

$$K_r(\Theta) = \Theta^{½}\left[\, 1 - \left(1 - \Theta^{1/m} \right)^m \right]^2 \tag{9}$$

where m = 1 - 1/n, and $\Theta = (\theta - \theta_r) / (\theta_s - \theta_r)$ and *n* is defined for van Genuchten's closed form equation for moisture release:

$$\Theta = \frac{1}{\left[1 + (\alpha h)^n\right]^m} \tag{10}$$

Van Genuchten et al. (1991) has developed a procedure to determine the parameters α, and n of Eq. (9) and (10) simultaneously with a least-squares curve-fitting algorithm. Matching the $K_r(h)$ or $K_r(\theta)$ calculated above at one measured value of K(h), usually a near-saturated conductivity of the soil matrix, one can obtain the absolute K(h) or K(θ) function.

The success of this method varies greatly, however. Durner (1994) felt that failures of models to estimate K from moisture release curves may be due more to poor representation of the moisture release curve in certain regions of wetness rather than failure of the model. He proposed superimposing unimodal retention curves of the van Genuchten type to better fit the retention curve. Because this method is very sensitive to errors in measuring the θ(h) curve, Tseng and Jury (1993) recommended taking dense measurements of matric potential and water content in time and space when collecting moisture release data. Green and Corey (1971) compared parameter values for Eq. (3) when calculating unsaturated hydraulic conductivity and found that, when a measured conductivity was used as a matching factor, the fit with experimental data was better.

There has been little success in using scale factors determined for θ(h) curves to determine K(θ).

Regression Based Equations for K_{sat} and K(h)

Generally these relationships have been much less successful than equations developed for θ(h) curves. Rawls et al. (1983; 1985) have published extensively on these relationships and there are many others available in the literature. Bonsu and Laryea (1989) reported that regressions based on sand and clay contents were not successful because the distributions were often bimodal. They also reported that parameters from effective porosity did not do as well either. They found clay and sand volumes to be better predictors, probably because they included some measure of soil structure. The coefficients from the regressions, however, did not reproduce the same range of conductivities as measured. Overall, researchers have had more success with locally derived equations than with equations developed from national data sets.

Inverse Problem Methodology and Numerical Simulation Methods

Progress is being made in methods to simultaneously identify multiple unknown parameters of some suitable functional forms of K(h) and θ(h) by repeated numerical solutions of the flow equations for water in soil. The inverse method is a parameter estimation technique that uses data from transient flow events and numerical or analytical solutions of the governing equations for water flow in soil subject to imposed boundary conditions. This method is typically carried out by running a simulation model with initial estimates of the unknown hydraulic properties as input. The simulation model output is compared to measured output and an appropriate non-linear estimation method is used to iteratively adjust the parameters until the errors between measured and predicted data are minimized. The function containing measured and predicted data that is minimized is known as the objective function. The hydraulic properties may include the properties themselves such as K_{sat} or unsaturated conductivity or parameters for the equations that are used to calculate conductivity as a function of matric potential or water content in the latter case. It is assumed that the soil hydraulic properties may be described by a relatively simple deterministic model that contains a relatively small number of known parameters (Russo et al., 1991).

A variety of methods have been used to obtain the data to be used in the objective function. The early work with inverse parameter estimation used outflow data from undisturbed soil cores sitting on a porous ceramic plate (Parker et al., 1985). A positive pressure was applied to the surface of the sample and water flow out the bottom was measured (Kool et al., 1987). Sanlini et al. (1995) used water content values obtained by drying cores by evaporation. They fit an exponential relationship for K(h). The data used in the objective function also include field or laboratory infiltration data (Russo et al., 1991), field drainage data (Eching et al., 1994a), or sequences of moisture in soil profiles over time (Ross 1993). Warrick (1993) used infiltration measurements and fit a scale factor, the ratio of K_{sat} to an inverse characteristic length. Zachmann et al. (1981, 1982) used the cumulative discharge at the bottom of a soil column during gravity drainage to identify parameters of K(h) and θ(h). Dane and Hruska (1983) and Wall and Miller (1983) used measured soil water content profiles during drainage for this purpose. Tensiometric data could be used as well.

This method is very promising but suffers from a number of serious drawbacks that has limited its acceptance. The nature of the objective function which is non-linear in the model parameters and the governing equations may lead to serious problems. One problem concerns non-uniqueness of the solution. This means that for any given set of output data there may be more than one set of input data that gives the same result. Non-uniqueness may be caused by trying to fit too many parameters, or excessive error in the data the model is attempting to match (Kool and Parker, 1988; Mous, 1993). Another problem, related to non-uniqueness, is identifiability of the parameters. Given an equation that describes the relationship between hydraulic conductivity and matric potential, for example, with three parameters, there may be an infinite number of combinations of two of the parameters that generate the same response. For example, changes in output caused by changes in K_{sat} can sometimes be compensated for by changes in θ_s- θ_r (Mous, 1993; Durner, 1994). For another example see Russo et al. (1991). The solution will not be stable if small errors in the response (i.e., transient flow data) result in large changes in values of the parameters. The more error there is in the measurements, the fewer the number of terms that can be fit (Warrick, 1993). A limitation of the evaporation and other methods arises from the difficulty of obtaining an accurate gradient near saturation where conductivities are high and gradients are low (Wendroth et al., 1993). Stability can be maximized by including as much independent information about the parameters being fit as possible (Kool and Parker, 1988; Eching and Hopmans, 1993). For example, when trying to identify the unsaturated hydraulic conductivity relationship, knowledge of K_{sat} or an independent measure of a parameter in a particular hydraulic model can improve stability. Crescimanno and Iovino (1995) reported that θ(h) and K(θ) functions from multistep outflow data and supplemented with three equilibrium θ(h) values were more reliable than the θ(h) and K(θ) functions from one-step outflow experiments supplemented with independently determined θ (h) values. Eching et al. (1994b) also reported that use of a measured water retention curve improved the estimates.

In a study on identifiability, uniqueness and stability of inverse problem methodology Russo et al. (1991) concluded that identifiability depended on the structure of the model used to describe the hydraulic property. Uniqueness and stability depend on the hydraulic model, on the quality of prior information on the hydraulic model parameters, and on measurement error. Durner (1994) reported poor performance of an inverse method when using a Mualem-van Genuchten K(θ) relationship when n<1.25. The method was also sensitive to the difference θ_s-θ_r. Error in measurement may be compensated for by quality of prior information on the hydraulic model parameters. Even if these considerations are fulfilled the method may still yield poor results if the measured data do not represent a wide range of conditions that will result in significant changes in the response function as the parameters change (Kool et al., 1985). In the end the problem may still require quite a bit of 'juggling' before a satisfactory solution is found. Overall, it may be better to use a relatively simple hydraulic model (less parameters) and use a small range of flow data (near the wet range, for example) where the simple model can describe the functional relationship well. Then use a number of such relationships to describe the full range of measured data.

Spatial Variability and Scaling Factors for Hydraulic Conductivity

Scaling factors are used as a to describe the spatial variability of a parameter. Warrick (1993) fit the scale factor, the ratio of the saturated conductivity to the characteristic length to infiltration data to which Phillip's infiltration equation (Phillip, 1957) was fit. Daamen et al. (1991) used soil texture to calibrate a scaling factor which they then used to generate a soil moisture release curve for an area using a reference curve. Because of the conditions at their study site, soil color was a good predictor of the scale factor because it indicated the presence of exposed B horizon material.

Inverse problem methodology has also been used to find scale factors for hydraulic parameters for field-scale and variability studies to describe their spatial variability. Inverse methods work well if only a scale factor needs to be fit because the number of parameters are reduced. Eching et al. (1994a) measured hydraulic properties on soil cores in the lab, calculated scaling factors and then measured soil water content profiles during drainage at a number of locations in a field. They scaled the drainage data and obtained a reference drainage curve to which they fit hydraulic properties using an inverse method. The scaling factors determined in the lab were then used to determine a distribution of hydraulic conductivities in the field.

Ahuja and Williams (1991) proposed the use of the Gregson, Hector, McGowan scaling method to predict the wetting front capillary potential (ψ_w), a parameter in the Green-Ampt type infiltration equation defined as:

$$\psi_w = \int_{\infty}^{0} \frac{K(\psi)}{K_s} \tag{11}$$

This function could be determined from field measured infiltration-time data. Equation (1) can also be used to estimate the spatial distributions (cumulative frequency distributions) of K_{sat} (Ahuja et al., 1984, 1989) by using ϕ_e to develop scaling factors based on the concept of similar media (Miller and Miller 1956; Warrick et al., 1977).

Spatial Variability

It has been reported that mean hydraulic conductivity increases with the measurement scale up to a critical distance (Rovey, 1994). It is not certain how much of this effect can be attributed to measurement scale or to technique. The effect also varies with the nature of the soil medium. Overall, small-scale values of K_{sat} may or may not average to regional values as numbers are increased. The errors in estimation of K may be large, Vereecken et al. (1992) estimated moisture supply capacity and found that estimation errors in the hydraulic properties were much greater than those due to map unit variability. Terrain attributes also become important at larger scales. Moore et al. (1993) reported significant correlations between soil properties such as organic matter and soil texture and terrain attributes which suggests terrain attributes should be useful in indirect methods. Of course, this may limit the generality of the equations.

The use of scaling along with inverse problem methodology appears to be a promising method to quantify spatial variability of soil hydraulic properties. Hopmans (1987) compared several methods to scale soil hydraulic properties and reported that the largest reductions in errors occur when different scaling factors are used for the K(h) and θ(h) relationships. Also, scaling reduced the sums of squares more for the θ(h) curve than for K(h). Moolman (1988) did report some linear correlations for scaling factors for different hydraulic models, for example the two parameters of the Phillip's equation (Phillip, 1957). This is similar to the type of relationships reported by Williams and Ahuja (1992) for the slope and intercepts of log(θ)-log(-h) relationships.

There does seem to be some potential for averaging point measurements and using similar media scaling theory to obtain regional estimates of hydraulic conductivity. It has been shown that the reference curve derived from similar media scaling agreed with curves derived by averaging soil hydraulic functions and with those obtained by optimization of the mean outflow for a set of 20 soil samples (van Dam et al., 1994).

Predicting Hydraulic Conductivity from Remotely Sensed Data

Can remotely sensed data on water content be reconstructed by modeling using average soil hydraulic properties and scaling? Lascano and van Bavel (1980) felt that it was possible but one could not use a random distributions of parameters, they should be distributed as landform characteristics. Since the variability of water content was greatest after rainfall, they felt it would be better to use water contents shortly after rainfall. Regional latent and sensible heat fluxes have been estimated using remotely sensed measurements in a surface energy balance equation (Feddes et al., 1993). The authors merged remotely sensed data with sparse measured ground data to estimate parameters to use to calculate evaporation fluxes from remotely sensed data. Feddes et al. (1993) proposed that these evaporation fluxes along with remotely sensed water contents, if available, could be used in the objective function of an inverse method using a soil-water-vegetation model. In numerical experiments they demonstrated that a reference curve for large-scale hydraulic conductivity could be obtained by inverse modeling of regional evaporation fluxes and remotely sensed water content. Local scale K(h) curves could then conceivably be obtained by using similar media scaling factors developed independently. The problem, however, still presents a large number of numerical difficulties because of the relatively large number of parameters needed to be estimated and the potential for measurement error.

Estimation of Sorptivity

Less work has been done with the goal of estimating sorptivity from simpler properties than with hydraulic conductivity. Sorptivity has been related to landscape properties (Gillieson et al., 1994; Van Es et al., 1991). Van Es et al. (1991) found sorptivity to be correlated with color development, clay and sand content. This suggests that it may be useful to consider landscape position in regression equations along with vegetation and soil type for regional-scale estimations. Stronger spatial dependence of sorptivity under dry than under wet conditions has also been shown (van Es et al., 1991 and Moolman, 1988). Moolman (1988) reported a correlation between the two parameters in Phillip's infiltration equation, one of which is sorptivity. This suggests that it may be possible to develop parameters for Phillip's equation similar to that done by Williams and Ahuja (1992) for $\theta(h)$ and K(h). Sorptivity has also been shown to be related to capillary length which is also used as a scale factor (Warrick and Broadbridge, 1992).

Other Methods

There are a number of other promising methods that come from other fields of engineering and process optimization and have just begun to appear in the soils literature. These include several so called 'artificial intelligence' methods. Neural networks have been used to develop predictors for moisture release curves using soil texture and bulk density (Pachepsky et al., 1996). The advantage of a neural network is that it can map non-linear relationships better than linear least squares methods. It may have promise for developing functional relationships for hydraulic conductivity as well. A genetic algorithm is another optimization tool that has been used to parameterize crop models (Sequeria and Olson, 1995) and is promising as a tool for inverse methodology. An advantage is that it works well in a solution space that has multiple minimums.

There has also been some work on modeling approaches that do not depend on equations derived from the Naiver-Stokes equations to describe the movement of fluids in porous material. Di Pietro and Melayah (1994) describe a method to model water infiltration by interacting lattice gas-cellular automata.

Future Work

The real challenge in our efforts to come up with new methods to describe soil variability at different scales is to come up with measurement methods that are appropriate at different scales. Currently we are using measurement methods carried out at a small scale and then using various methods to upscale them to determine hydraulic properties at the field scale. Russo et al. (1991) suggest that the length scale of the flow measurements should be on the same order of magnitude of the correlation scale of the hydraulic properties being identified. This suggestion is complicated by the fact that the correlation scale often varies with the size of the sample. While one can talk about finding a value of saturated conductivity for a representative elementary volume (REV) of soil, the determination of what is representative may vary for different scales. Some specific suggestions for future research follow.

- Indirect methods require some field measurements of the hydraulic properties and measurements of soil water status or flux during the development stage. There is a large potential for error which can be due to the limited number of measurements in time and space, separation distance between readings of water content and matric potential, and use of volume averaged readings taken from point measurements (Tseng and Jury, 1993). A technique that may be useful to minimize the effects of these errors is to use a method proposed by Bosch (1991) to predict the error that results from using point estimates of matric potential to determine mean matric potential in a heterogeneous soil profile.

- Models of K vs h or K vs θ are often not valid throughout the entire range of the independent variable. Methods to describe soil hydraulic properties by other means such as sum of simple functions (Ross and Smettem, 1993) or spline functions may be promising ways to incorporate the best attributes of several models into one function.

- Terrain attributes have been shown to be correlated with certain soil properties (Moore et al., 1993; van Es et al., 1991). This suggests that terrain attributes should be incorporated into relationships for surface soil attributes, soil hydraulic properties and soil texture to obtain regional-scale prediction equations.

- Inverse methods still hold much promise. As the performance of computers continues to improve, some of the computational limitations to inverse methods will diminish. Furthermore, there are now a number of commercially available soil moisture measuring devices including TDR and capacitance based devices that can be used to collect water content data in spatially and temporally dense patterns. This can be a valuable source of data for inverse methods to determine hydraulic properties at the field scale. It can also be combined with scaling methods.

- Soil structure, chemistry, and macro voids greatly influence soil hydraulic conductivity but are more difficult to measure than soil texture or bulk density. Future research should be directed towards methods to quantify these properties and incorporate them into indirect methods.

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Estimating the Soil Water Retention Curve with Soil Bulk Density and -33 kPa Value

Robert D. Williams[1], Laj R. Ahuja[2], and Dennis Timlin[3]

Abstract

Two methods to estimate the soil water retention curve are summarized: similar-media scaling and a one-parameter model. Both require the saturated soil water content, or its estimate from the soil bulk density, plus one measured point on the curve (such as the water content at a matric potential of -33 kPa). Similar-media scaling also requires knowledge of one complete curve to serve as a reference. The one-parameter model is based on the log-log form of the soil water retention curve below the air-entry value of ψ, and requires a generalized slope-intercept relationship. Although the methods estimate the soil water retention curve fairly well, visual comparison of the calculated vs measured soil water content showed less scatter in the relationship to the 1:1 line, concomitant with smaller calculated error terms, for the one-parameter model.

Introduction

The soil water retention curve, that is the relationship between the soil water content and the soil matric potential, is one of the two basic hydraulic properties of the soil. This relationship is often used to estimate the other soil hydraulic property, the hydraulic conductivity as a function of soil water content or soil matric potential. Both these hydraulic function relationships are needed for modeling infiltration based on the current detailed theory (*i.e.,* Richards equation). The simplified infiltration equations, like the Green-Ampt approach, require the knowledge of saturated hydraulic conductivity and a parameter called the wetting-front suction or capillary drive. Where measured values are not available, the soil water retention curve can be used to estimate both of the above infiltration parameters. This approach is, thus, useful in obtaining the spatial distribution of infiltration parameters more easily.

Unfortunately, however, the laboratory and field methods used to determine the soil water retention curve itself are time consuming and labor intensive, especially when a large number of samples are required to account for spatial variability or to characterize a large area. Because of the expense in time and labor, attempts have been made to estimate the soil water retention curve from soil properties which are more easily measured (e.g. soil texture or bulk density) and/or limited data.

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[1]USDA-ARS, Grazinglands Research Laboratory, 2707 W. Cheyenne Street, El Reno, OK 73036.

[2]USDA-ARS, Great Plains System Research Unit, P.O. Box E, Fort Collins, CO 80522.

[3]USDA-ARS, Systems Research Laboratory, Bldg. 007, Rm 008, 10300 Baltimore Ave, Beltsville, MD 20705.

Various techniques have been employed to estimate the soil water retention curve. Multiple regression analysis has been used to relate soil water content at fixed matric potentials to soil texture, bulk density and organic matter (Rawls *et al.*, 1982; 1983). A similar analysis has related parameters of a certain functional representation of the soil water retention curve to the above soil data (Saxton *et al.* 1986; Wösten and van Genuchten, 1988). Furthermore, physico-empirical models of the soil water retention curve have been derived from a more complete particle-size distribution (Arya and Paris, 1981; Haverkamp and Parlange, 1986). Similar-media scaling has been used, employing the bulk density or saturated soil water content, plus one measured point on the soil water retention curve and a known reference curve for a given soil type (Ahuja *et al.* 1985; Williams and Ahuja, 1992). A recent approach is based on the log-log model of the soil water retention curve using one known value and a generalized slope-intercept relationship (Gregson *et al.*, 1987; Williams *et al.*, 1992). Although all these methods have been used successfully (Ahuja *et al.*, 1985; Williams *et al.*, 1992), here the discussion is limited to two techniques: similar-media scaling and the one-parameter model.

Materials and methods

The soils used here are the same as those used previously (Ahuja and Williams, 1991; Williams and Ahuja, 1992). All the data on the soil water retention curve, $\psi(\theta)$, were measured on undisturbed cores. Soil type, source of the data, number of samples, lowest value of ψ in the data set, and the residual soil water content (θ_r) estimated for each data set are given elsewhere (Ahuja and Williams, 1991). The θ_r for each data set was estimated from one characteristic curve, close to the mean curve of the data set by fitting a Books and Corey (1964) log-log relationship. It was assumed to be constant within a data set. For six soils θ_r was zero, but for two others, Teller sandy loam and Pima clay loam, it was 0.02 m^3m^3. In calculations below the soil water content at the -33 kPa matric potential was used as the measured $\psi(\theta)$ value required by both models.

Similar-media scaling

Details of this technique are give elsewhere (Ahuja *et al.*, 1985). It relies on the extended similar-media scaling concept (Warrick *et al.*, 1977; Simmons *et al.*, 1979). The matric potential for a fixed degree of saturation ($S = \theta/\theta_s$, where θ_s is the saturated value) at site *i*, $\psi_i(S)$, is related to a mean matric potential value, $\psi_m(S)$, by

$$\alpha_i \psi_i(S) = \psi_m(S) \tag{1}$$

where α_i is a scaling factor for site *i* that applies at all different values of *S*.

In our work we redefined *S* as $(\theta-\theta_r)/(\theta_s-\theta_r)$, with θ_r remaining constant within a soil type. We assumed that below the air-entry matric potential, $\psi_m(S)$ can be expressed by a power-form equation (Brooks and Corey, 1964)

$$\psi_m(S)=AS^{-M} \tag{2}$$

with A and M constants. The constant A is the air-entry potential, ψ_{em}. Substituting AS^{-M} for $\psi_m(S)$ in Equation (1) and rearranging yields

$$S(\psi_i)=(\alpha_i\ \psi_i\ /A)^{-1/M}. \tag{3}$$

Equation (3) can describe the variation in S at different sites at a fixed value of ψ

$$S_i\ (\psi)=(\alpha_i\ \psi/A)^{-1/M}. \tag{4}$$

This equation has two unknowns, and a set of α_i values that must be determined. Suppose that for one given value of ψ (e.g. -33 kPa) the values of S_i at all different sites are known, and the mean of these S_i values is $\hat{S}$. At the site where S_i is closest to $\hat{S}$, the complete $S(\psi)$ function is also given, and its slope is taken as an adequate approximation of the slope $-1/M$ of the scaled mean $S(\psi_m)$ in the form of Equation (4). Then, solving Equation (4) for α_i gives

$$\alpha_i=(A/\psi^*)S_i^{-M} \tag{5}$$

where ψ^* is the known fixed value, taken here as -33 kPa, for all sites, and $S_i = S(\psi^*)$. Now we impose the condition that the mean of the α_i values for N sites is equal to 1.0, or that the sum of all α_i values is equal to N

$$\sum_{i=1}^{N}\alpha_i=N=(A/\psi^*)\sum_{i=1}^{N}S_i^{-M} \tag{6}$$

which yields

$$A=N\psi^*/\sum_{i=1}^{N}S_i^{-M}. \tag{7}$$

With A determined, all different α_i values can be found from Equation (5). Knowing each α_i value, the $S_i(\psi)$ function for each site is obtained by using Equation (4). The ψ value where $S_i(\psi)$ equals 1.0 is the air-entry value, ψ_{ei}, for location i. With known θ_{Si}, or one estimated from the soil's bulk density, the $S_i(\psi)$ can be converted to $\theta_i(\psi)$. In our work we used the known θ_{Si} values. The fixed value of ψ^* used was - 33 kPa. The reference curve for obtaining M was a curve in the data set whose θ at -33 kPa and θ_S were closest to the mean values.

The one-parameter model

The one-parameter model is based on the work of Gregson *et al.* (1987) and, as in our earlier work, will be referred to as GHM (Ahuja and Williams, 1991; Williams and Ahuja, 1992; 1993). This model is based on the log-log linear form of the soil water retention curve (Brooks and Corey, 1964) in the matric potential range below the air-entry value. We modified the GHM model to include the residual water content, θ_r, where $\psi(\theta)$ is expressed as:

$$\ln|\psi| = a + b\ln(\theta - \theta_r) \tag{8}$$

where a and b are the intercept and slope of the log-log linear relationship, respectively. Gregson *et al.* (1987) found that Equation (8), with $\theta_r = 0$, provided a good fit to several sets of data for British and Australian soils. They also found a strong, linear relationship between the intercept (a) and slope (b) of Equation (8) ($r \geq 0.99$), which could be expressed as:

$$a = p + qb \tag{9}$$

Substituting Equation (9) into Equation (8) yields the one-parameter model, provided that an approximated value of θ_r for the soil type under consideration is known or can be estimated from information in the literature:

$$\ln|\psi| = p + b[\ln(\theta - \theta_r) + q] \tag{10}$$

Equation (10) can then be used to estimate the entire $\psi(\theta)$ relationship below the air-entry value of ψ, simply from one measured value on the $\psi(\theta)$ curve. The known (ψ,θ) value is used to determine the only unknown parameter, b, in Equation (10). We used the values of soil water content at -33 kPa and derived b as:

$$b = \frac{\ln(\psi_{-33\ kPa}) - p}{\ln(\theta_{-33\ kPa} - \theta_r) + q} \tag{11}$$

The known or bulk-density estimated θ_S value caps off the $\psi(\theta)$ curve and enables determination of the ψ_e value. Again we used the known θ_S values.

In our previous work (Ahuja and Williams, 1991) we showed that the $\psi(\theta)$ data of all the soils were described quite well by Equation (8). In addition the constants a and b of this regressed equation were highly correlated linearly, in accordance with Equation (9). Values of p and q for each soil were determined using a linear regression through the (a,b) data pairs. Although we found the a vs b relationship for all soils were fairly close to each other, it was still better to divide the soils into three textural groups having somewhat different a versus b relationships. The soils were grouped and average p and q values were determined for each group (Table 1).

Data analysis

For each method the calculated water content was plotted against the measured water content. In addition, a mean error and root-mean-square error at each potential were determined. The mean error at each potential was calculated by summing the difference between the calculated and measured soil water content and dividing by the number of observations. The root-mean-square error at each matric potential was calculated by taking the square root of the sum of the squares of the differences between the calculated and measured soil water content divided by the number of the observations minus one. The mean error is an error in the estimation of mean $\theta(\psi)$ at any fixed ψ, while the root-mean-square error is an average error of estimation for individual data points at any given potential.

Table 1. Average p and q values for soil texture groups.

Group	Soils	Textural ranges	p	q
			ln(kPa)	$\ln(m^3/m^3)$
1	Oxisols, Kirkland, Renfrow, Pima	Loam-silty clay loam-clay loam	1.415	0.839
2	Norfolk, Teller, Bernow	Sandy loam-sandy clay loam	0.343	1.072
3	Lakeland, Bernow	Sand	0.541	1.469

Results and discussion

Similar-media scaling

Typical results for the similar-media scaling technique are presented in Fig. 1. It should be noted that the calculated versus measured water content for Renfrow, Teller and Kirkland are for the -10, -100 and -1500 kPa matric potentials, while those for Lakeland are for the -1, -5 and -10 kPa matric potentials. In most cases the calculated water content is fairly close to the 1:1 line. Most of the error shown for Lakeland sand is at the -1 kPa matric potential. The mean error ranged from -0.03 to 0.03 m^3m^{-3}, while the root-mean-square error ranged from 0.01 to 0.05 m^3m^{-3}.

Errors for similar-media scaling are generally smaller than those for the Rawls *et al.* (1982) texture model. In previous studies we have shown that the similar-media scaling techniques gave much better results than the texture-based regression methods (Ahuja *et al.*, 1985; Williams *et al.*, 1992). However, we have also shown the errors for the regression based models are somewhat smaller when one known $\psi(\theta)$ value is used. Even with this improvement the similar-media scaling technique gives better results than the broad-based regression models (Ahuja *et al.*, 1985).

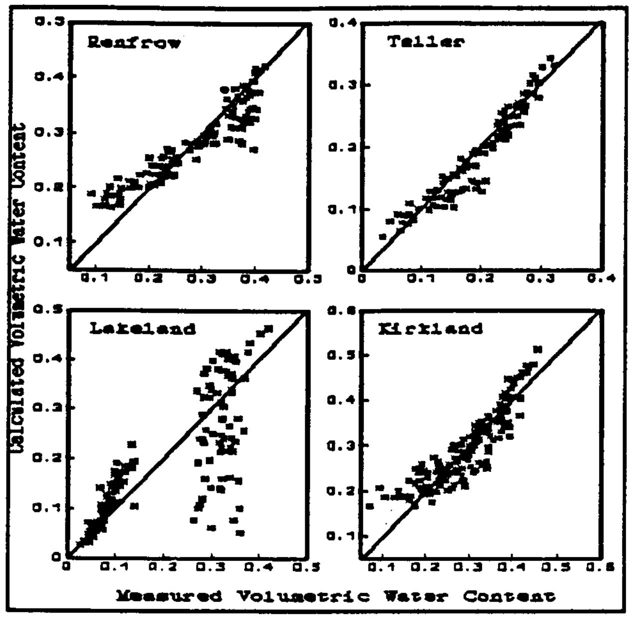

Fig. 1. Calculated versus measured soil water content resulting from the similar-media scaling technique.

One-parameter model

Results for the GHM model are shown in Fig. 2. Here again we used the same matric potentials as shown in Fig. 1. Note that the scatter is somewhat reduced in the Lakeland sand results, here the mean error and root-mean-square error were 0.008 and 0.033 m^3m^{-3}, respectively, at -10 kPa as compared to 0.02 and 0.025 m^3m^{-3} for similar-media scaling. Overall the mean errors ranged from 0.002 to 0.015 m^3m^{-3}, while the root-mean-square errors ranged from 0.011 to 0.064 m^3m^{-3}. These errors were quite similar to those for similar-media scaling.

For a range of soil textures we have shown that the GHM model provides better estimates of θ than the similar-media scaling technique (Williams and Ahuja, 1992). In the worse case examined (Pima clay) GHM was at least equal to similar-media scaling. If we had used individual p and q values for each of the soils presented above, rather than the group values as presented in Table 1, the errors would have been smaller and the estimates based on the GHM method better. However, if the GHM method is to have any utility at all, we consider that the use of group p and q values is necessary.

In the results presented above the soil water content at the -33 kPa matric potential was used as the known $\psi(\theta)$ value to calculate the slope, b. However, in many cases, as for example in the SOILS-5 database or the USDA-SCS soil surveys, only the available water content (AWC) or its range is provided. The applicability of GHM would be greatly extended if AWC could be used in place of a known $\psi(\theta)$ value to calculate b. In working with the data a curvilinear relationship between AWC and b was observed (Williams and Ahuja, 1993). This relationship between AWC and b was fitted with a 2nd degree polynomial regression. The polynomial regressions for each soil were statistically significant ($p \leq 0.05$) but accounted for only 21 to 69%

of the variability in the data. Using group p and q values and b estimated from the AWC the soil-water content was calculated. Using this method there was an increase in the scatter in relation to the 1:1 line which was reflected in larger root-mean-square errors ranging from 0.03 to 0.09 m^3m^{-3}. Mean errors ranged from 0.011 to 0.064 m^3m^{-3} and were similar to results reported with the GHM model when $\theta_{-33\,kPa}$ was used as the measured $\psi(\theta)$ value. When individual p and q values were used with individual AWC-based equations for b for each soil, errors in estimating soil-water content were only slightly larger than those if θ at the -33 kPa matric potential had been used to calculate b (Williams and Ahuja, 1993).

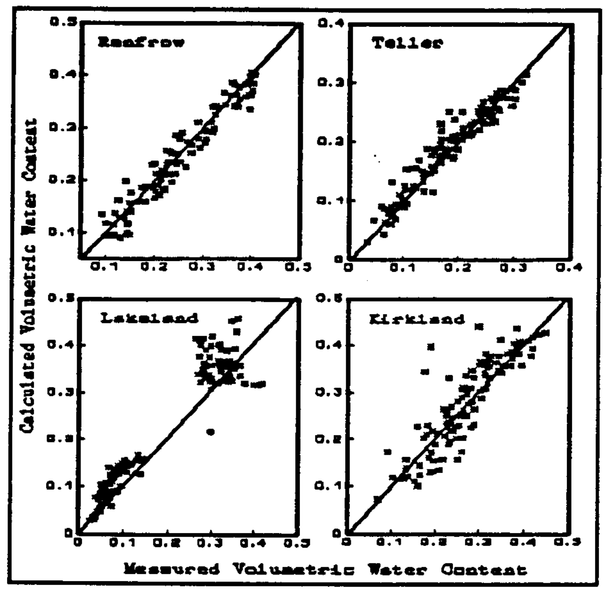

Fig. 2. Calculated versus measured soil water content resulting from the use of the GHM model.

In other work a linear relationship between p and the natural log of the air-entry pressure [$\ln(\psi_e)$], as well as between q and the natural log of the effective porosity [$\ln(\theta_e)$] (Fig. 3) (Williams and Ahuja, 1993) has been demonstrated.

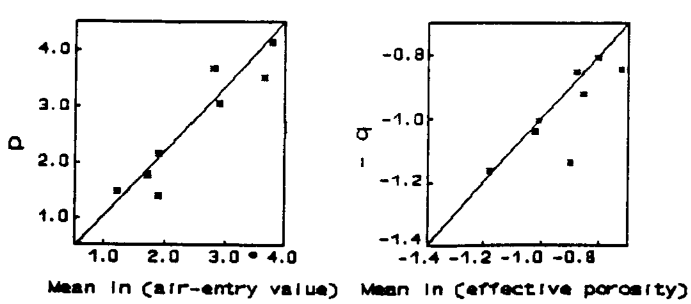

Fig. 3. Relationship between p and mean $\ln(\theta_e)$, and between q and mean $\ln(\psi_e)$ for eight U.S. soils.

If this relationship holds then we may be able to use the texture class mean values of ψ_e and θ_c provided by Rawls *et al.* for p and q values. Figure 4 shows preliminary results for Renfrow, Teller, Lakeland and Kirkland soils. When these results are compared to Fig. 2 we can see a reduction in the scatter in relation to the 1:1 line.

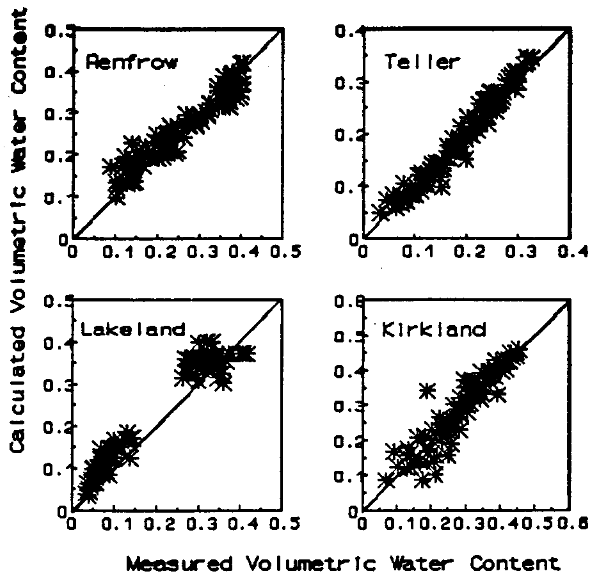

Fig. 4. Calculated versus measured soil water content using GHM with p and q values based on mean values of $\ln(\theta_e)$ and $\ln(\psi_c)$, respectively.

This reduction in the scatter in the data is reflected in smaller error terms. For example using the original p and q values the GHM mean errors ranged from 0.001 to 0.03 m^3m^{-3}, while the root-mean-square errors are generally 0.04 and 0.05 m^3m^{-3}. When the p and q values based on $\ln(\psi_e)$ and $\ln(\theta_c)$ are used the mean errors ranged between 0.001 and 0.01 m^3m^{-3}, while the root-mean-square errors are generally around 0.03 m^3m^{-3}. The results presented in Fig. 4 were based on GHM using $\theta_{-33\,kPa}$ as the known $\psi(\theta)$ value. We have also estimated $\theta_{-33\,kPa}$ using a regression model (Rawls *et al.*, 1982) based on texture and organic matter content to estimate θ_{-33kPa}. Using this estimated value the results were similar to those presented in Fig. 4. However, in this latter case the error terms are somewhat larger: mean errors ranged between 0.001 and 0.094 m^3m^{-3}, while the root-mean-square errors ranged between 0.04 and 0.155 m^3m^{-3}.

Scaling with the one-parameter model

The one-parameter model can also be used for scaling the spatially or temporally variable $\psi(\theta)$ data. If we express the one-parameter model as

$$\ln[-\psi_i\ (\theta)]=p+b_i\ [\ln(\theta)+q] \tag{12}$$

Here p and q are constants independent of the spatial location, and b_i is the only parameter that is dependent on the location. Thus, b_i is a scaling factor for each location i. Rearranging (12), we obtain:

$$\ln(\theta) = [\ln[-\psi_i\ (\theta)] - p]/b_i - q$$
$$\ln(\theta) = scaled\ \ln[-\psi_i\ (\theta)] \tag{13}$$

Thus, at a fixed θ value, the right-hand term of Equation (13) should be the same for all spatial locations. This term is call the *scaled* $\ln(-\psi_i)$ (Ahuja and Williams, 1991). If the correlation between the intercept (a) and slope (b) in Equation (9) is perfect, plots of $\ln(\theta)$ versus the scaled $\ln(-\psi_i)$ for all different locations should coalesce into a single 1:1 relationship. Typical results for this scaling technique using group p and q values with individual slopes (b_i) are given for Renfrow (best case) and Cecil (worst case) soils.

When this scaling technique was compared to similar-media scaling the relative efficiency of GHM scaling $\psi(\theta)$ for different locations within individual soils ranged from 39 to 83% (Ahuja and Williams, 1991). These efficiencies were generally greater than the efficiencies calculated by the extended similar-media scaling using one measured value for each $\psi(\theta)$ curve. If the constants p and q in Equation (9) are found to be approximately the same in different soils, The GHM method would be a universal method of scaling and estimation across all soils types. This would be a distinct advantage over the similar-media scaling approach, which is generally restricted in its application to within a soil type.

Fig. 5. Results of scaling $\theta(\psi)$ for Renfrow and Cecil soils using GHM.

Conclusions

Similar media scaling and the one-parameter model (GHM), for estimating the soil water retention curve from saturated soil water content, or its estimate from the soil bulk density, plus one other measured point on the curve were compared. Overall visual comparison of the calculated vs measured soil water content for both methods showed less scatter in the relationship to the 1:1 line, concomitant with smaller calculated error terms, for the one-parameter model.

Both methods tend to better than the broad based regression models based on soil texture bulk density and organic matter content. Using previously derived *p* and *q* values for the one-parameter model estimated values were generally within 0.02 to 0.04 m^3m^{-3} of the measured soil water content for soils obtained from a large database. We also provided a regression relationship between *b* and the available water content (AWC) to estimate *b*, since in many cases AWC is available in the USDA soil survey reports where $\theta_{-33\ kPa}$ is not. Using *b* thus estimated in the one-parameter model gives only slightly larger errors in calculating the water content at different potentials than when using $\theta_{-33\ kPa}$. The newest development with the one-parameter model is the use of the $\ln(\psi_e)$ and $\ln(\theta_c)$ for *p* and *q* values, respectively. This enables us to use Rawls *et al.* (1982) mean $\ln(\psi_e)$ and $\ln(\theta_c)$ values for texture classes, providing 11 *p* and *q* values for broad textural ranges. We also showed that the one-parameter model can be used for scaling and that this method is more efficient than similar-media scaling. These last two points need further evaluation.

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Use Of Remote Sensing for Determining Spatial Infiltration Characteristics

Thomas J. Jackson[1] and Walter J. Rawls[1]

Abstract

Most of the current infiltration technology involves the use of point measurements of specific characteristics. These point measurements, together with physically-based models, give a fairly detailed description at the local scale. However, they do not give the necessary spatial information which is essential if the spatial and temporal dynamics of the process are to be quantified. Remote sensing offers a unique approach to measuring and integrating these parameters in a spatial manner. A literature review of the remote sensing techniques for estimating soil moisture, surface energy, water balance parameters and soil hydraulic properties which can be used for describing spatial infiltration properties was conducted. Future research directions for describing spatial infiltration parameters using remote sensing were proposed.

Introduction

Most of the current infiltration technology involves the use of point measurements of specific characteristics. Although these measurements are accurate, the natural variability of soil, water and vegetation parameters makes it extremely difficult to collect enough samples to accurately represent an area. These point measurements, together with physically-based models, give a fairly detailed description at the local scale. However, they do not give the necessary spatial information which is essential if the spatial and temporal dynamics of the process are to be quantified. Remote sensing offers a unique approach to measuring and integrating these parameters in a spatial manner.

The new generation of Earth observing satellites currently projected for the late 1990's have some potential for measuring and monitoring water, soil and vegetation parameters. These new sensor systems will provide concurrent remotely sensed data over a much broader spectral range than have ever been available. This will open up new opportunities to explore relationships between spectral information and surface processes related to hydrologic systems such as infiltration. Remote sensing can be used to monitor several types of variables related to infiltration: soil water, soil properties and vegetation properties. Much work remains to be done in linking these measurements to infiltration.

Literature Review

[1]USDA-ARS, Hydrology Lab, Bldg. 007, Rm. 104, BARC-West, Beltsville, MD 20705-2350

Microwave Remote Sensing of Soil Moisture

Passive microwave remote sensing of soil moisture has been studied with increasing interest for more than 20 years. Previous results have been summarized in Jackson and Schmugge (1989) and Ulaby et al. (1986). The basis for microwave remote sensing of soil water is the strong dependence of the soil's dielectric properties on its water content due to the large contrast between the dielectric constant of water and that of dry soil. The consensus of the research also established that the principal target parameter affecting the measured microwave emissivity is the volumetric soil water in the surface 5 cm of the soil (Jackson and Schmugge, 1989).

Recent research has focused on the effects of variables that have an effect on the interpretation of the emissivity data in an attempt to develop a useable and adaptable retrieval algorithm. Variables considered include soil properties (Schmugge, 1980 and Dobson et al., 1985), surface roughness of the soil (Choudhury et al., 1979 and Ulaby et al., 1986) and vegetation, Ulaby et al., 1986 and Jackson and Schmugge, 1991). Results of these studies have shown that the effects can be corrected when the necessary ancillary data are available and when the basic structure for a retrieval algorithm exists (Jackson, 1993).

Active microwave remote sensing (radar) has also been used for estimating surface soil water. Extensive information on the theory can be found in Ulaby et al. (1986). Radar technology has been adapted so that it can provide satellite based observations with the same nominal resolution of visible and near infrared sensors. There are several radar satellites currently operational (ERS-1 and 2, IERS, and RADARSET). A drawback to radar data for soil water studies is that the observations are also very sensitive to several other variables (roughness and vegetation). Therefore, only limited success has been achieved in this area. Solutions to these problems have been proposed (Oh et al, 1992); however, the data necessary is not likely to be available.

A continuing question in soil water remote sensing is how to retrieve the soil water profile. Reviews of previous research (Jackson, 1986 and Kostov and Jackson, 1993) suggest there are four basic approaches to the problem: 1) regression techniques based on developing and using a regression equation to calculate profile soil water from surface layer measurements; 2) knowledge-based techniques which use a-priori information on the hydrological properties of soils and statistical data on the depth profile to predict the profile moisture; 3) inversion techniques that use representative soil and temperature profiles and an inversion algorithm to calculate the profile soil water; and 4) combinations of remotely sensed data and water balance models; these use remotely sensed surface soil water for calibration or updating of the models, or as input data to the models to calculate the temporal behavior of the profile moisture.

Progress in this area has been made by utilizing two microwave measurements at different wavelengths (Reutov and Shutko, 1986). Using this approach, experimental results show that the moisture in the surface meter of the soil can be retrieved with an accuracy of 5% by volume (Mkrtchjan et al., 1988). New approaches to this problem have been proposed (Kostov and Jackson, 1993 and Entekhabi et al., 1994) that cannot be pursued with currently available data bases.

Multispectral Estimation of Surface Energy and Water Balance Parameters

In order to quantify the hydrologic cycle spatially, reliable estimates of the surface energy balance and evapotranspiration are required. Ground and aircraft-based remote sensing data in the optical wavelengths (Hatfield et al., 1984a; Reginato et al., 1985; Jackson et al., 1987; Kustas et al., 1990) as well as satellite-based data (Carlson et al., 1981; Price, 1982; Moran et al., 1989) have been used in concert with surface energy balance models for estimating ET from field to regional scales.

Efforts in combining information from visible and thermal wavelengths in models to account for variation in vegetation cover and soil water have also been explored (Nemani et al., 1993). In addition, synergistic approaches using microwave remote sensing which is unaffected by clouds with optical wavebands have been conducted.

Ground, aircraft, and satellite-based remote sensing data in the optical wavelengths have been successfully used to estimate instantaneous and daily surface energy fluxes over both cultivated surfaces (Price, 1982; Hatfield et al., 1984; Reginato et al., 1985; Jackson et al., 1987; Moran et al., 1989; Kustas et al., 1990) and partially vegetated surfaces in natural ecosystems (Kustas et al., 1989). Remotely sensed data have also been used to update physically-based models which continuously simulate transfers of moisture and energy in the soil-vegetation-atmosphere continuum (Carlson et al., 1981 and Ottle et al., 1989).

Laser altimeters in aerial surveys have provided rapid and accurate assessment of landscape surface features. Airborne laser altimeters have been used to measure topography (Krabill et al., 1984), vegetation properties (Ritchie et al., 1992, 1993), and erosion/stream features (Ritchie and Jackson, 1989; Ritchie et al. 1993). Most recently, the laser data has been used in a model for computing local and basin scale aerodynamic roughness (Menenti and Ritchie, 1994). This will be a very useful and cost-effective method for estimating the variation in surface roughness over large areas which can be used in atmospheric models simulating the surface energy balance over complicated landscapes.

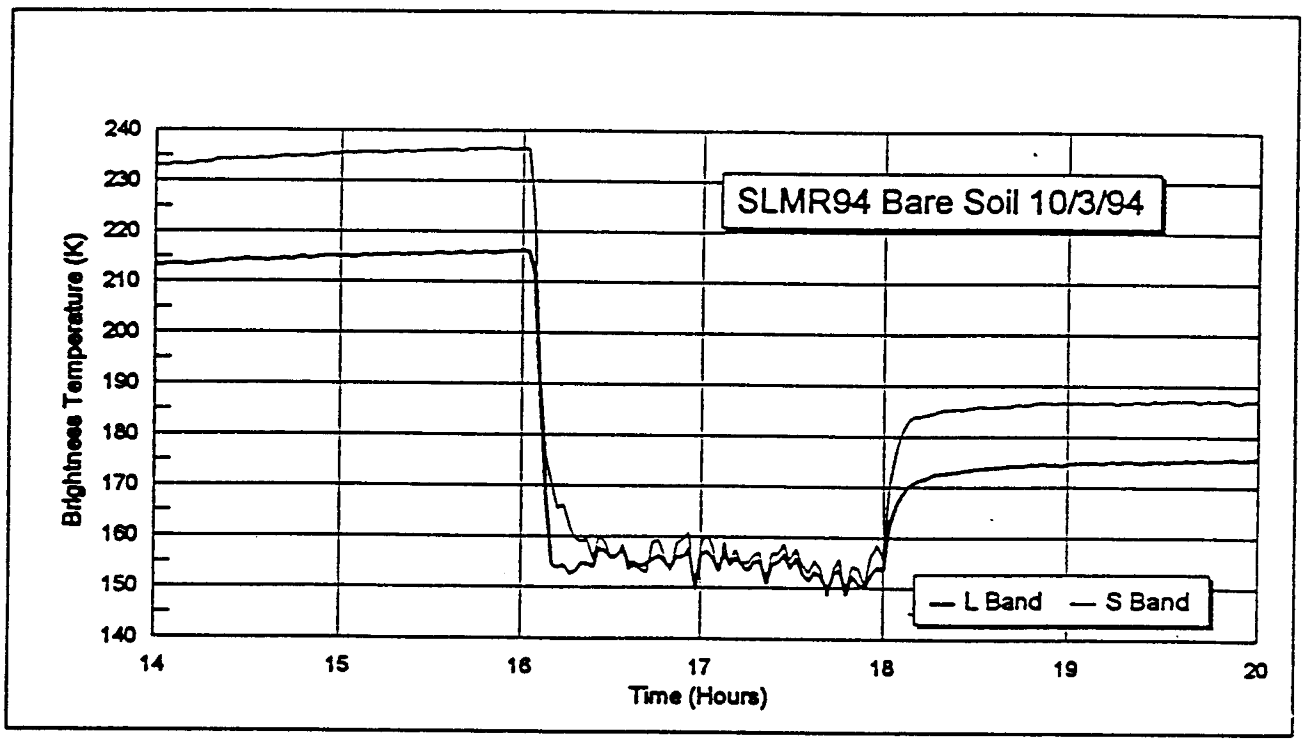
Estimation of Soil Hydraulic Properties

Areal estimates of both soil water at different depths and areal evaporation fluxes potentially allows the estimation of effective hydraulic soil properties over large areas.

Numerous methods are available for the direct measurement of hydraulic properties of soils. Most of these methods are extremely time consuming and expensive, especially when characterizing a large area. Alternatively, a large number of indirect methods have been derived to estimate hydraulic properties from more easily measured soil properties such as soil texture and other data routinely available from soil surveys (Rawls et al., 1991 and van Genuchten et al., 1992). Ahuja et al. (1993), using numerical modeling, developed a method which produced average profile saturated hydraulic conductivity based on two day drainage.

An example of what can be measured with a ground based passive microwave radiometer system is presented in Figure 1. These observations were made over a bare soil before, during, and after, an irrigation. The instruments were on a boom truck over the irrigation system and data was recorded every two minutes. The L-band is a longer wave length than the S-band and responds to a deeper soil depth. These very preliminary results show interesting features especially after irrigation was stopped at 18 hours.

Some work has been done relating changes in remotely sensed variables to infiltration. Camillo and Schmugge (1984) examined the relationship between change in brightness temperature and the antecedent rainfall using simulated data. In this approach, it is assumed that soil properties are uniform. The changes in brightness temperature, therefore, are related to the differences in antecedent rainfall. This can also be reversed by specifying the rainfall or assuming the same rainfall. Under these conditions the changes in brightness temperature would be related to differences in drainage properties of the area. Camillo et al. (1986) used experimental data (obtained using a truck based system) to explore this idea. Van de Griend et al. (1985) performed a sensitivity analysis of diurnal surface temperature to various soil properties. Recently Mattikalli et al. (1995) explored the use of brightness temperature change in assessing spatially distributed hydraulic conductivity. Hollenbeck et al. (1996) also investigated what information on soil hydraulic properties can be extracted from passive microwave data.



Rain starts (16 hr) Rain stops (18 hr.)

Fig. 1. Microwave measurements during irrigation.

Future Research

Parameters that can be detected remotely are a reflection of the continually changing water and energy balance of the entire soil profile as it interacts with the atmosphere. While meteorological conditions tend to be uniform over large areas under some conditions, the opposite is true of soil properties in the case of hydraulic and thermal characteristics.

Relationships between soil hydraulic / infiltration characteristics and remotely sensed data should be established using theoretical and experimental analysis. The effects of scale and resolution on these relationships should be evaluated using a combination of ground based, aircraft and satellite remote sensing. The specific research needs are:

- Investigate the incorporation of multitemporal and spatially distributed passive and active microwave data with indirect methods for determining soil hydraulic properties over large areas.

- Integrate remotely sensed soil water estimates from microwave sensors, vegetation cover from visible and near-IR, surface temperature from thermal-IR, and vegetation height, density and roughness from laser-altimeter data into energy balance and hydrologic models for computing spatial distributed infiltration properties.

- Explore profile moisture estimation methods utilizing multifrequency and multitemporal remote sensing observations.

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Measurement Methods to Identify and Quantify Spatial Variability of Infiltration on Rangelands

Ginger Paige and Jeff Stone[1]

Abstract

This paper summarizes the current state of the art in measuring the spatial variability of infiltration on rangeland watersheds. Infiltration is known to vary extensively across spatial and temporal scales due to heterogeneities in the soil properties as well as the vegetation and cover characteristics. Studies conducted to measure the spatial variability of infiltration on rangelands have found that the ability to measure the spatial variability of infiltration is a function of both the method and the scale of measurement. Current measurement methods are primarily conducted at point or small plot scale and measure either saturated (ponded) infiltration or unsaturated (rainfall) infiltration. The benefits and limitations of these methods as well as areas for future research are discussed.

Introduction

Hydrologic processes which occur on rangelands are highly variable in space and time due to the nature of the climatic input, topography, soils, and vegetation. Infiltration, an important component of the rainfall-runoff process, is significantly affected by both the temporal variability of rainfall and snow melt and the spatial variability of soil and vegetation properties. The hydrologic response of an area is influenced significantly by the characteristics and areal extent of the cover and soil variability. Rangeland vegetation is composed of communities or complexes of species and can include trees, shrubs, grasses and forbes, each which influence the soil surface and sub-surface characteristics in a different manner. A single infiltration rate or a lumped average is often used to define the infiltration capacity of a watershed without considering the location of areas of high and low infiltration capacity (Morin and Kosovsky, 1995). Lumping of distributed parameters can lead to distortions in the results of distributed process based models (Lane et al., 1995). Measurement of the variability of vegetation and soil properties is relatively easy, quantifying the effects of that variability on the infiltration process and subsequent impacts on runoff generation is much more difficult. This is due in part to difficulty in measuring the infiltration process.

Spatial variability is first attributed to the inherent heterogeneity of the rangeland infiltration characteristics, and second to the method of measurement itself (Jury, 1985; Aboulabbes, 1984; Merzougi, 1982). The scale of infiltration measurement has ranged from watershed studies using natural rainfall, to large and small plot studies using a variety of rainfall simulators, to point studies using infiltrometers (Branson et al., 1972). Many of the point infiltration methods are now being used to characterize the spatial variability of infiltration across an area. Infiltrometers and rainfall simulators are the two predominant methods which have been

[1]USDA-ARS, SWRC, 2000 E. Allen Road, Tucson, Arizona 85719

used to measure infiltration and its spatial variability on rangelands, though other methods have been used. Both methods have limitations in their ability to simulate infiltration as it occurs under natural rainfall conditions.

Since the 1980s, a number of studies have used point measurements with geostatistics in an attempt to quantify the spatial variability of hydrologic processes (Bosch and Goodrich, 1996). Point measurements can be limited in their ability to characterize the spatial variability of infiltration in relationship to hydrologic characteristics such as topography, elevation, soil, and other watershed characteristics. Other important factors which need to be considered are: 1) the portion of the measurement area or watershed contributing to infiltration and runoff (partial area contribution); 2) the method and scale of measurement; and 3) the sampling design (random, grid, transect, irregular spacing).

Field Observations

Infiltrometers

The majority of the studies conducted to measure the spatial variability of infiltration across a watershed have used point measurements such as ring infiltrometers or disk permeameters. These types of measurements have several advantages: the infiltration rate is measured directly, the measurement time is relatively quick, and the cost of the experiment is low so that many measurements can be made. A summary of the studies reviewed which were conducted using infiltrometers is presented in Table 1.

One of the first studies to measure the spatial variability of soil hydraulic properties in the field was conducted by Nielsen at al. (1973). Steady state infiltration measurements were made at twenty 6.5 m square plots. The infiltration rate varied from 0.5 mm/hr to 50 mm/hr, with a CV of 91%. Steady infiltration rate fit a log-normal distribution; the infiltration rate was highly correlated with the percent saturation, but not correlated with water content. Sharma et al. (1980) used a double ring infiltrometer (inner ring diameter of 46 cm) to measure the spatial variability of infiltration and sorptivity at the R-5 watershed near Chickasha, Oklahoma. Measurements were made at 26 sites in the watershed in a regular grid pattern with a spacing of about 60 m. Steady state infiltration rates were always reached within 60 min. No obvious pattern in the distribution of the infiltration parameters was found with respect to soil type or position in the watershed. The frequency distribution, however, was found to be log-normal. Subsequent studies using infiltrometers have also found that the results were best described by a log-normal distribution (Sharma et al., 1983; Loague and Gander, 1990; Achouri and Gifford, 1984; Merzougi and Gifford, 1987; Grah et al., 1983).

Variability studies of infiltration have used both classical statistics and spatial statistics to describe the variability and resulting distributions of the measured values (Bosch and Goodrich, 1996). The coefficient of variation (CV) has commonly been used to describe the variability of infiltration capacity (Warrick and Nelson, 1980) which characteristically has a large CV (Tables 1 and 2). The CV, however, is only an indicator of the extent of and not the distribution of the variability over an area. In order to describe the spatial distribution of the variability, researchers began to use geostatistical methods. Geostatistical methods and kriging had been successfully used to determine the spatial variability of infiltration and sampling requirements on an agricultural field in Davis, California (Vieira et al., 1981). When applied to rangeland watersheds, geostatistical methods have often found correlation lengths ranging from several meters (Aboulabbes et al., 1985; Grah et al., 1983) to no variance structure at all (Merzougi and Gifford, 1987; Achouri and Gifford, 1984). The scale of the measurement used in proportion to the sample spacing and the size of the area being measured has been found to be very important in determining the spatial variability.

Geostatistical methods were used to determine the optimum sampling procedure at R-5 watershed, based on 50 initial steady-state infiltration measurements made along a transect at 5-m intervals (Loague and Gander, 1990). A total of 157 measurements were taken across the watershed using a 25-m grid spacing based on the range suggested from the semivariogram of the initial transect. A final transect of 40 steady-state measurements was made at 2-m intervals. They found that the range of spatial persistence for infiltration on the R-5 catchment was very small and that the 25-m grid was not sufficient to map the infiltration variability. The scale of spatial correlation between measurements was found to be less than 20-m.

Achouri and Gifford (1984) and Merzougi and Gifford (1987) used a 2-m interval sampling grid on grazed and ungrazed sites in Utah. Each study used a double ring infiltrometer to measure 70 and 104 locations at each site, respectively. The results from both studies suggest that the infiltration rates are randomly distributed for the sample interval of 2 m. In each case kriging could not be used to interpolate between measurements as no variance structure was found to exist.

Grah et al. (1983) investigated the distribution of infiltration relative to slope position and overland flow paths on a small watershed on the Wasatch plateau in central Utah. Infiltration rates were measured at 5 minute intervals using double ring infiltrometers. The infiltration rate was highly correlated with both vegetation cover and soil bulk density for all sampling times. A significant relationship was found between 55 minute infiltration rates and overland flow distance. The range of spatial correlation increased with an increase in infiltration time from 3.4 m at the 1 minute interval to 17.4 m at the 55 minute interval along the flow path. This suggests that the spatial correlation of infiltration rate varies with time during the infiltration process, becoming more homogeneous over time as the affect of the suction term in early infiltration decreases with the increasing soil moisture.

Aboulabbes (1984) compared the semi-variograms from two different transects on the same watershed in Morocco. Steady state infiltration measurements were made with double ring infiltrometers at 1 m intervals along both transects. The two transects had significantly different space dependence structure, indicating that neither one could be used to represent the spatial variability of infiltration across the watershed. A Gaussian model, used for one of the transects, showed a spatial correlation distance of 18 m. The other transect could only be fit with a linear model using 25 m of the transect. In general, all the semi-variograms indicated a large nugget effect and a spatial correlation structure over a very short distance.

Other Infiltrometer Methods

Disk permeameters and tension infiltrometers (White et al., 1992; Elrick and Reynolds,

1992) are variations of the cylinder infiltrometer method. The infiltration rate into the soil surface can be measured under ponded (disk permeameter) or unponded (tension infiltrometer) conditions. Whitaker (1993) measured infiltration at 10 m intervals along a 300 m transect on the Walnut Gulch Experimental Watershed in southeast Arizona using both a disk permeameter and a tension infiltrometer. Measurements, with both the disk permeameter and the tension infiltrometer, were made at each site approximately 20 cm apart within a 5-day period. The disk permeameter was used with a positive hydraulic head of 0.5 cm and the tension infiltrometer was set at a hydraulic head of -5 cm. An average infiltration rate of 266 mm/hr with a standard deviation of 231 and a CV of 87 % was found for the 30 sites using the disk permeameter. The average infiltration rate using the tension infiltrometer and a -5 cm head was 53.8 mm/hr with a standard deviation of 22 and a CV of 42 %. The CV for infiltration was much lower with the tension infiltrometer than the disk permeameter, though the average initial moisture contents were similar. The average infiltration rate measured with the tension infiltrometer is comparable with infiltration rates determined using WEPP (Water Erosion Prediction Project, USDA, 1995) rainfall simulator plots on the same watershed. The infiltration rates from the rainfall simulator varied from 49 to 57 mm/hr for the dry, wet, and very wet runs, with an average of 53.2 mm/hr.

Rainfall Simulators

Infiltration measurements on rangelands using rainfall simulators usually measure the infiltration rate indirectly. The steady state infiltration rate is often calculated as the difference between rainfall application rate and the equilibrium runoff rate. The initial infiltration rate is assumed to equal the application rate until runoff commences. The rainfall simulator plots have varied in size from 1 m^2 to over a hectare (Meyer, 1994). Small simulators are often used as ponded infiltrometers, taking measurements at several locations across an area to determine the spatial variability of infiltration. Studies using large simulators, such as the rotating boom simulator used in the WEPP studies, often measure the variability of the cover characteristics within plots and relate it to the calculated infiltration rates. A summary of the studies reviewed which were conducted using small rainfall simulators and disk permeameters is presented in Table 2.

Small Simulators

Aboulabbes (1984), and Merzougi and Gifford (1987) compared infiltration measurements from ponded infiltrometer rings with those from a modular rainfall simulator plots under both wet and dry conditions. Both methods exhibited large variability in infiltration rates across the watershed (Table 2). Infiltration rates were found to be exponentially distributed in most cases. As expected, a significant difference was also found between the two methods. The ponded ring infiltration rates were much higher than the modular simulator except at very low application rates. The infiltration rates from the double-ring infiltrometers were significantly affected by initial moisture conditions. The results of the autocorrelation and semi-variogram analyses conducted were similar to the results found by Achouri and Gifford (1984). Merzougi and Gifford (1987) found that the infiltration measurements were not spatially correlated, i.e. there was a complete lack of variance structure and the measurements were all independent. Only 18-36% of the variance could be explained by cover characteristics. A significant difference was found however, between grazed and ungrazed sites and between rainfall simulator and double ring infiltrometer measurements. The infiltration rates measured with the double ring infiltrometer were 2 to 3 times higher than the rates determined by the rainfall simulator. These results are similar to the findings of Aboulabbes et al. (1985).

Springer and Gifford (1980) found the distribution of measured infiltration rates for a site in south western Idaho could be described by either a normal or a log-normal distribution. Data reported by Gifford and Busby (1974) were used to describe the spatial distributions of infiltration. A sprinkler infiltrometer was used to measure infiltration rates in twenty four 0.23 m^2 plots over a five year period on thirteen dates. The sprinkler infiltrometer was run for a 25 minute period at an intensity of 76 mm/hr. The average infiltration rate varied from 56 mm/hr to 28 mm/hr, while the CV varied from 0.68 to 0.34 over the 5 year period. The results were similar to those predicted by Rao et al. (1979).

The Green and Ampt (1911) infiltration equation, or some modification of it (Mein and Larson, 1973; Chu, 1978), is often used to determine infiltration parameters from rainfall runoff field studies in spatially varied rangelands (USDA, 1995; Kidwell et al., 1996). Devaurs and Gifford (1986) used the Green and Ampt infiltration equation with parameters determined from field data and soil textural properties to characterize infiltration on spatially varying rangelands. Using a least squares method to fit the field data, they found limitations in the ability of the Green and Ampt equation to describe the observed variable infiltration patterns on rangelands. When using Green and Ampt parameters predicted from soil texture data, the method was most appropriate for disturbed sites with infiltration rates less than 30 mm/hr.

Simulated rainfall was also used to compare infiltration rates and erosion at 28 study sites in 5 different watersheds in the Great Basin area of Nevada (Blackburn, 1975). Infiltration rates were positively correlated with slope and negatively correlated with soil moisture. Percentage of large diameter (>2 in.) rock cover was poorly correlated with infiltration; whereas, the percent small diameter rock cover was positively correlated with infiltration. Percent bare ground was strongly correlated with infiltration rates. Poesen et al. (1990) found soil surface rock cover increases the infiltration rate into the soil, and the effect of the rock cover on the infiltration rate is proportional to the percent cover.

Large simulators

The rainfall application rate is an important factor to consider when using rainfall simulators to determine the spatial variability of infiltration at both the point and plot scales (Aboulabbes et al., 1985; Hawkins, 1982). A comparison between point and plot scale measurements using rainfall simulators found that the point measurements were unable to describe the infiltration at the plot scale at low rainfall intensities (Cundy, 1982). The ability of the point measurements to describe the infiltration processes at the plot scale improved at higher rainfall intensities. Dunne et al. (1991) found that infiltration rate varied with flow depth, and that rainfall intensity had a strong effect on the apparent infiltration rate on short hill slopes. Rainfall intensity influenced flow depth along the slope and therefore had a secondary effect on the spatial pattern of infiltration. The apparent infiltration was also found to be affected by the microtopography, as well as the hill slope length and gradient. At high rainfall intensities the onset of runoff is more likely to be determined by the rainfall intensity.

Lane et al. (1987) used a rotating boom rainfall simulator to measure infiltration and evaluate the effects of cover characteristics on infiltration. They found final infiltration rates decreased as the vegetative canopy cover and rock and gravel cover decreased. Tromble et al. (1974) found that the soil- vegetation complex and antecedent moisture had a significant effect on infiltration rates. Tisdall (1951) found antecedent soil moisture had a significant effect on infiltration rate. Bolton et al. (1990) found vegetation had a slight, but significant, effect on infiltration rates. Busby and Gifford (1981) and Simanton et al. (1991) found that removing canopy cover had little direct effect on infiltration and runoff processes.

Discussion

The spatial variability of rangeland soils and soil hydraulic properties is well recognized, however, the methods to measure and characterize the spatial variability are limited. Current methods used to measure infiltration are limited in their ability to measure the process in the field under natural conditions and to quantify the spatial variability. Studies that have used point measurements across a watershed have often found large variations in final infiltration rates and large CVs. These measurements are not realistic in measuring the infiltration process during a rainfall event, or in quantifying the interactions between soil, cover, topography, and rainfall intensity. Larger scale measurements made with rainfall simulators, often measure variability within a plot (vegetation and cover, slope, micro-relief, etc.) but then relate this variability to a lumped infiltration rate for the entire plot which was determined indirectly.

Classical statistical methods measure changes over distance and determine the number of samples necessary to characterize an area based on the frequency distribution of the observations, but provide no information about the variability of the observations with respect to the position or coordinates of the area (i.e. spatially) (Vieira et al., 1981). Rogowski (1972) proposed a variability criteria to indicate the size and type of an area that is sufficiently uniform to be represented by a single soil property or characteristic such as infiltration. Geostatistical techniques, autocorrelograms and semi-variograms have recently been used to determine the range of correlation of infiltration values in space. As discussed earlier, the spatial correlation of infiltration on rangelands has been found to be very small, often less than 2 m, using current geostatistical methods (Loague and Gander, 1990). Grah et al. (1983) found spatial correlation distance increases with longer infiltration periods and when evaluated along flow paths. The ability of the autocorrelogram to compute the spatial variability of infiltration is dependent on the length of the transect measured (Peck, 1983). Vieira et al. (1981) suggested measuring infiltration rates at the finest grid possible with enough samples to detect the spatial structure before determining the appropriate variogram model. They also emphasized that the semi-variogram (not the autocorrelogram) should be used to determine sampling distance because it represents the average for all directions.

The measurement scale has been found to have a direct impact on the resulting variability of the infiltration measurements. Sisson and Wierenga (1981) and Baily (1995) found that the infiltration rate increased and the variance decreased with an increase in measurement area. Jury (1985) conducted a critical review of the studies of the spatial variability of soil physical parameters in solute migration. Five studies were evaluated where scaling theory was used to interpret the variability of measured parameter values at different sample sites. The scaling factors inferred from the measurements of soil parameters depended critically on the method of measurement. A significant correlation was found between the correlation length of a parameter and the sample size spacing used to develop the variogram, indicating that the correlation length parameters depend on the sample grid spacing used to obtain the variogram or correlogram. This implies that neither scaling factors nor correlation length parameters are measurable field properties using current methodologies (Jury, 1985), and that sampling and measurement methods to determine the spatial variability of soil parameters controlling transport needs further study. Russo and Jury (1987) analyzed the effects of grid size on the ability to estimate correlation scale. Reasonable correlation estimates were found when the sampling distance was smaller than the scale of the underlying process being measured.

Vegetation has been found to be one of the primary factors influencing infiltration on rangelands (e.g. Lane et al., 1987; Blackburn et al., 1992). Gifford and Busby (1974), however, found measuring the cover characteristics did not improve the potential to predict the hydrologic response of a big sage brush site which had been highly modified. Dunne et al. (1991) found empirical studies in the literature to be confusing as to how vegetation affects infiltration processes on rangelands. Of the many factors controlling infiltration on rangelands, the role of desert and range vegetation and desert or erosion pavement are not well understood or quantified (Lane et al., 1987). Percent vegetation cover was found to be consistently positively correlated with final infiltration rates (Aboulabbes, 1984). Stepwise multiple regression analysis, however, was not successful in predicting the infiltration rates from other measured watershed and soil properties including vegetation.

There is a need to measure both the spatial variability of infiltration and the spatial characteristics of the structural properties and cover characteristics which influence infiltration at the same time (Bosch et al., 1993). Multiple geostatistical analysis (e.g. Co-kriging) using slope, vegetation and cover characteristics, and soil structural properties should be tested. The development of a new method or variations of existing measurement methods and sampling designs should be used to incorporate landscape topography and overland flow processes.

Suggested Topics for Future Research Include:

- Integration of methods: e.g. tension infiltrometers and large rainfall simulators used on the same plot.

- Incorporate topography and dominant flow paths as well as sample spacing into sampling design for infiltrometers.

- Determine a relationship between the scale of measurement and the measurement method in order to minimize the affect of the method (or size of the measurement) on the resulting spatial variability.

- Measure both the spatial variability of infiltration and the spatial characteristics of the structural properties and cover characteristics which influence infiltration at the same time.

- Multi-variable geostatistical methods should be considered as a framework for measuring the spatial variability of infiltration on rangelands.

Table 1. Infiltration rates and associated CVs for selected rangeland sites using infiltrometers.

Study	Method	Infiltration mm/hr	CV (%)	Number	Location	Comments
Nielsen et al. 1973	single ring	6.1	91	20	Fresno, CA	log-normal
Sharma et al. 1980	double ring	47	60	26	R-5 Chickasha, OK	log-normal
Loague & Gander 1990	1-m ring				R-5 Chickasha, OK	log-normal no spatial correlation
	grid	56.8	73	157		
	transect 1	76	48	50		
	transect 2	23.4	43	50		
Achouri 1982	double ring	(range)			Utah	no variance structure log-normal
	ungrazed	116 - 216	54-73	70		
	grazed	45 - 76	36-49	70		
Grah et al. 1983	double ring	412	72	120	Utah	correlation distance ≤17 m along flow path correlation with vegetation
Aboulabbes 1984	double ring				Morocco	correlation distance ≤ 18m
	20 -cm dry	334	73	53		
	20-cm wet	169	92	53		
	30 cm dry	304	67	53		
	30 cm wet	148	129	53		
Merzougi & Gifford 1987	double ring				Eureka, Utah	cover explained 36% of the variance
	ungrazed	124.4	44	104		
	grazed	49.2	37	104		

Table 2. Infiltration rates and associated CVs for selected rangeland sites using small simulators and disk permeameters.

Study	Rainfall intensity/ initial condition	Infiltration mm/hr	CV (%)	Number	Location	Comments
Small Simulators:						
Aboulabbes 1984	75 mm/hr dry 75 mm/hr wet 100 mm/hr dry 100 mm/hr wet	25 32 34 32	76 68 88 74	53 53 53 53	Morocco	30 cm diameter
Springer & Gifford 1980 (Gifford & Busby 1974)	75mm/hr Native Sagebrush plowed/ seeded plowed/seeded/grazed plowed/seeded/grazed	 55 36 - 56* 28 -34*	 34 43 - 68* 40 - 64*	 22 20 - 24* 19 - 24*	Southern Idaho	5 years normal & log-normal
Devaurs & Gifford 1986	very wet (64 mm/hr) unfenced fenced tilled	 15 20 9	 30 21 14	 22 45 27	Reynolds Creek Boise, Idaho	small (0.37 m^2) plots with large rainfall simulator plots
Merzougi & Gifford 1987	grazed ungrazed	17 31	110 75	104 104	Eureka, Utah	randomly distributed pure nugget effect
Disk Permeameter:						
Whitaker 1993	Ponded h = 5 cm Unponded h = -5 cm	 266 54	 87 42	 30 30	Walnut Gulch Tombstone, AZ	300 m transect

* range of values for the five year period.

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Statistical and Geostatistical Characterization of Spatial Variability

David Bosch[1] and David Goodrich[2]

Introduction and Background

Classical statistical approaches assume field variability is purely random and observations of soil hydraulic properties are statistically independent regardless of their spatial position (Vauclin et al., 1983). Statistical characterizations include parameter estimates of a mean, variance and distribution. These estimates can be used as effective parameters or Monte Carlo simulations can be made using the distribution of the data. However, variations in soil properties tend to be vertically and horizontally correlated over space. Because of this, geostatistical characterizations are often used to incorporate the variability into infiltration and chemical transport modeling. Geostatistical methods facilitate the examination of spatial and temporal correlations in the data (Nielson et al., 1973). Tools such as kriging allow estimation of spatially correlated data using measurements taken in close proximity to the point at which the estimate is being made.

Statistical characterization of spatial structure of hydraulic characteristics is important for several forms of analyses (Unlu et al., 1990): such as 1) estimating point or spatially averaged values of soil hydraulic properties using kriging techniques, 2) designing sampling networks and improving their efficiency, and 3) stochastic modeling in order to understand the overall response of heterogeneous flow systems (Kitanidis and Lane, 1985).

Field Observations

Despite its utility, the application of geostatistics to modeling of unsaturated zone processes is a relatively new field, limited by a lack of high quality field data. Nielsen et al. (1973) performed one of the first extensive field experiments quantifying the variability of hydraulic properties of a Panoche clay loam over a field area. Significant variability was found in the particle composition, soil-water pressure, bulk density, and saturated and unsaturated hydraulic conductivity in the soil profiles. Nielsen et al. (1973) reported an average and a standard deviation (sd) of the saturated hydraulic conductivity (Ks) in the vertical direction of 1.5 cm hr^{-1} and 0.05 cm hr^{-1} respectively. Byers and Stephens (1983) obtained core samples in horizontal and vertical transects in order to study the statistical and stochastic properties of particle-size parameters and Ks. Laboratory measured Ks was found to be log-normally distributed, with a mean and a sd of 61.2 cm hr^{-1} and 36 cm hr^{-1}. Hopmans et al. (1988) examined data collected in horizontal and vertical transects over a 650 ha watershed. The mean ln Ks value reported was 1.7 with a sd of 0.6 (Ks in cm day^{-1}). Brace (1980, 1984) and Clauser (1992) found permeabilities tend to grow with the characteristic scale of measurement for both sedimentary and crystalline rocks.

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[1] USDA-ARS, SEWRL, PO Box 946, Tifton, Georgia 31793

[2] USDA-ARS, SWRC, 2000 E. Allen Road, Tucson, Arizona 85719

Carvallo et al. (1976) measured unsaturated hydraulic conductivity ($K(\psi)$) versus depth in five infiltration plots within a 0.01 ha area in a Maddock sandy loam with an average particle composition of 80 % sand, 11 % silt and 9 % clay. Tensiometers were installed at seven depth intervals down to a maximum of 1.52 m. Soil-water characteristic data determined from soil samples were used in conjunction with tensiometer measurements to compute $K(\psi)$. Significant variability in $K(\psi)$ was found both between plots and over vertical profiles. Standard deviations of $K(\psi)$ ranged from 0.3 cm hr^{-1} at saturation to 0.003 cm hr^{-1} at the residual water content.

In addition to these values, correlation scales for various soil characteristics have been reported in the literature. Some of the reported values for ln Ks are presented in Table 1. These data include values for vertical Ks along horizontal transects only. As the table shows, considerable variability exists between the observed correlation scales and also with respect to the material.

Russo and Jury (1988) found the sample network configuration had a distinct effect on estimation of covariance function parameters. They compared the effect of two different sampling networks for estimating the covariance function on the predicted head variance. It was found that using a modified sampling network with irregular spacing was superior to use of a systematic sampling network (regular grid) for estimation of parameters of the covariance function. This was particularly true for fine textured soils when a three-dimensional treatment of hydraulic conductivity variations was employed and when relatively small correlation scales were present.

Saturated and residual water contents have been measured extensively and representative statistical characteristics for these parameters are available in literature. Some values are presented in Table 2. An excellent review of similar data is presented in Jury (1985).

Relationship to Scale

Examination of the data in Table 1 reveals a close relationship between the overall scale at which the measurements were taken and the correlation scale of the ln Ks data (Fig. 1). A very high correlation coefficient (0.93) exists between these two variables. For a more general case, data assembled by Jury (1985) also indicate a significant positive correlation between the correlation length of several measured soil physical properties and the sample grid spacing used to develop the variogram or correlogram. The implication of this is that correlation length is not a property of the measured soil, but rather a property of the measurement methods and scale. When working with soil pH data, Gajem et al. (1981) found an apparent correlation length of 1.5 m when a 0.2 m measurement grid spacing was used, a correlation length of 21.6 m when a 2.0 m spacing was used, and a correlation length of 130 m when a spacing of 20 m was used.

Table 1. Observed standard deviations and correlation scales for hydraulic conductivity measured along horizontal transects.

Source	Material	Parameter	Standard Deviation	Correlation Scale (m)	Overall Scale (m)
Delhomme (1979)	limestone aquifer	ln Ks	2.3	6,300	30,000
Binsariti (1980)	basin fill aquifer	ln Ks	1.0	800	20,000
Russo and Bressler (1981)	Hamra Red Mediterranean soil	ln Ks	0.4	1	100
Luxmoore et al. (1981)	weathered shale subsoil	ln Ks	0.8	1	14
Sisson and Wierenga (1981)	silty clay loam soil (alluvial)	ln Ks	0.6	0.1	6
Viera et al. (1981)	Yolo soil (alluvial fan)	ln Ks	0.9	15	100
Devary and Doctor (1982)	alluvial aquifer (flood gravels)	ln Ks	0.8	820	5,000
Byers and Stephens (1983)	fluvial sand	ln Ks	0.5	1	15
Hoeksema and Kitanidis (1985)	sandstone aquifer	ln Ks	0.6	45,000	50,000
Unlu et al. (1990)	Panoche soil, 75 cm depth	ln Ks	0.6	8	80
Unlu et al. (1990)	Panoche soil, 105 cm depth	ln Ks	1.0	9	80
Anderson and Cassel (1986)	Portsmouth sandy loam, A horizon	ln Ks	722	1.5	500
Anderson and Cassel (1986)	Portsmouth sandy loam, A horizon	ln Ks	5.5×10^4	2.5	500
Anderson and Cassel (1986)	Portsmouth sandy loam, A horizon	ln Ks	2.2×10^3	na	500
Hopmans et al. (1988)	sand, BC horizon	ln Ks	0.8	0.75	12.5

Table 2. Representative saturated and residual moisture content and standard deviations for various soil textures.

Source	Parameter	Soil Texture	Soil Horizon or Orientation	Mean	Standard Deviation
Hopmans and Stricker (1989)	θsat	sand	A	0.406	0.032
Hopmans and Stricker (1989)	θsat	sand	BC	0.391	0.045
Hopmans and Stricker (1989)	θsat	sand	D	0.437	0.055
Wierenga et al. (1989)	θsat	gravelly sandy loam	vertical	0.321	0.032
Wierenga et al. (1989)	θr	gravelly sandy loam	vertical	0.086	0.020
Burden and Selim (1989)	θsat	silt loam	horizontal	0.54	0.045
Burden and Selim (1989)	θr	silt loam	horizontal	0.14	0.095
Nielsen et al. (1973)	θsat	clay loam	vertical	0.454	0.045
Cameron (1978)	θsat		vertical	0.470	0.045
Anderson and Cassel (1986)	θsat	sandy loam	A	0.457	0.071
Anderson and Cassel (1986)	θsat	sandy loam	Btg	0.391	0.100
Anderson and Cassel (1986)	θsat	sandy loam	Bg	0.310	0.071
Anderson and Cassel (1986)	θr	sandy loam	A	0.096	0.032
Anderson and Cassel (1986)	θr	sandy loam	Btg	0.075	0.032
Anderson and Cassel (1986)	θr	sandy loam	Bg	0.044	0.032
Carvallo et al. (1976)	θsat	sandy loam	vertical	0.393	0.014
Russo and Bressler (1981)	θsat		0-0.9 m	0.367	0.045
Russo and Bressler (1981)	θr		0-0.9 m	0.078	0.045
Greminger et al. (1985)	θsat	Yolo loam	0.3 m	0.459	0.010
Greminger et al. (1985)	θsat	Yolo loam	0.6 m	0.486	0.014

Russo and Jury (1987) studied the uncertainty of the estimation of correlation length scales for stationary fields using 100 computer generated realizations of a two-dimensional isotropic second-order stationary field with a given correlation length. They found both the accuracy of correlation scale estimates and the fitted variogram increase with the number of sample points and as the correlation scale of the underlying process increased. To obtain reasonable estimates of the correlation scale, the distance between sampling points must be smaller than half the range of the underlying process. However, data from a one-dimensional transect may result in underestimation of the correlation scale of the underlying process by a factor of two or more.

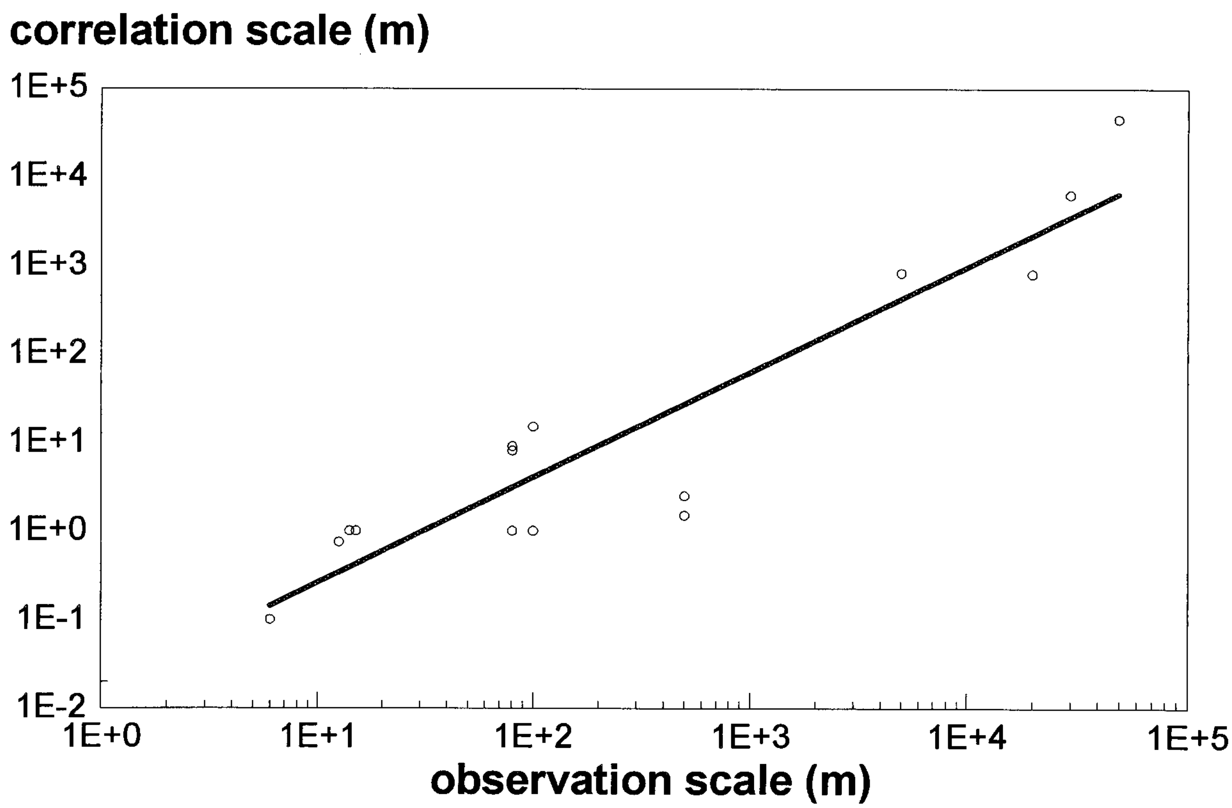

Fig. 1. Relationship between observation scale of the measurements and the calculated correlation scale for the Ks data of Table 1.

Neuman (1994) also observed an increase in estimated correlation scale with increasing measurement scale in computed aquifer permeabilities and dispersivities. He reviewed the data summarized by Gelhar (1993) and plotted the correlations scales of hydraulic conductivity and transmissivity data versus the characteristic lengths of the corresponding fields sites. The data indicated a consistent increase in reported correlation scale with field length over a range of correlation scales from 10 cm to 45 km. In general it appeared the correlation scale was typically 10 % of the field scale. For figure 1, the correlation scale was approximately 8 % of the observation scale.

To explain the phenomena observed by Neuman (1994), he noted that if hydrogeologic media is viewed as a nested sequence of distinct hydrogeologic units with a discrete hierarchy of scales then "one obtains a semivariogram function that increases with the mean tracer travel distance(s) in a stepwise rather than gradual fashion" as the semivariograms of the discrete units are superimposed. This would correspond to crossing from one soil type or geologic material to another. Further, each step in such an echelon represents a natural correlation scale at which the log permeability is statistically homogeneous or nearly so; while other scales are locally either inactive or suppressed (Neuman, 1994). The change in dominant natural scales from one setting to another can result in an infinite variety of possible semivariograms. To resolve this variation Neuman (1994) stated that data from a large number of settings would be required so that enough scales are sampled to ascertain the "underlying commonality of these semivariograms in the form of a generalized power law."

Interpretation and semivariogram analysis assumes a natural scale over which the property in question is statistically homogeneous. As Neuman (1994) points out, increasing correlation scales with increasing field scale implies statistical homogeneity of log permeabilities is at best a local phenomenon occurring intermittently over narrow bands of the scale spectrum. Hence one must question the utility of routinely associating geologic medium properties with REV's as has been the custom for several decades.

Using theoretical simulations, Starks and Fang (1982) showed that a nonlinear drift in the mean value of the parameter, if not filtered out before analysis, would produce an apparent increase in correlation length as the sample density was decreased. In the theory of regionalized variables from which geostatistics is derived, each measured parameter is considered to be a single realization taken from a single probability distribution. In order to apply this theory it is necessary to assume the property is spatially stationary so that each location is described by the same probability distribution, and spatial covariances depend only on the separation between measurements and not on the absolute location (Journel and Huijbregts, 1978). Thus, a drift, or a change in the mean of the data would be a violation of the underlying assumptions. Data sets collected along long transects crossing over several different soil types may contain this nonlinear drift component.

Summary

As this review has shown, considerable variability exists in natural porous media. This variability precludes precise characterization of the hydraulic parameters. Experience has shown soil parameters vary widely and at best we can determine the statistical and spatial characteristics of this variability. It appears that geostatistics has considerable applicability to infiltration science. However, the relationship between correlation scale and measurement scale is disturbing. The correlation scale is not solely a function of the parameter being measured, but also a function of the measurement scale. While interesting from an observational standpoint, it is not clear whether the relationship is useful in the predictive sense. On the average, it appears the correlation scale is approximately 10 % of the observation scale. However, from site to site this may vary considerably. Application of this relationship in scaling-up would not account for the site specific differences. Thus, care must be taken in study design, data analysis and data interpretation.

One possible avenue to approach the problem centers around soil classification. Studies which sample across two different soil types break the underlying assumptions behind geostatistical analysis. Characteristics of different soil types are expected to have different means and distributions. Crossing over from one soil type into another, as is frequently done in these studies, violates the assumption of spatial stationarity. Our efforts may be better directed toward characterizing either single soil types or examining relationships between soil types lying in similar landscape positions.

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Characterization of Soil Spatial Variability with Fractals

Daniel Giménez and Walter J. Rawls[1]

Introduction

Fractals were introduced to the scientific community by B. B. Mandelbrot (Mandelbrot, 1983; Feder 1988). Burrough (1983) pioneered application of fractals to study spatial variations of soil properties. Later studies extended the application of fractals to the description of aggregate size distribution (Perfect and Kay, 1991), pore roughness (Kampichler and Hauser, 1993; Giménez, 1995; Pachepsky et al., 1995), soil surface roughness (Huang and Bradford, 1992), and to the prediction of transport coefficients such as hydraulic conductivity (Rieu and Sposito, 1991; Rawls et al., 1992; Giménez et al., 1994).

Applications of fractals to hydrology have been limited to the analysis of stream network and river basins (Mc Leenan et al., 1991). The spatial dependence of infiltration rates and saturated hydraulic conductivity has been demonstrated using geostatistical methods (Sisson and Wierenga, 1981; Vieira et al., 1981; Lauren et al., 1988). Other studies, however, has shown a random variation of infiltration rate and hydraulic conductivity (Smetten, 1987; Mohanty et al., 1994). This apparent contradiction could be caused by variations that are dominated by short range heterogeneities not captured by a large sampling interval (Burrough, 1983; Armstrong, 1986; Culling 1986).

As noted by Moltz and Boman (1993), applications of fractal concepts to hydrology have been the characterization of processes or soil properties with fractal-like behavior (Chan et al., 1994, and Moltz and Boman 1995). There has not been much effort, however, to link both approaches. It is this possibility, together with the simplicity of the model, that makes fractals an attractive alternative to traditional studies of spatial variability of soil properties.

The objectives of this section are to: i) review application of fractal concepts to the characterization of spatial variability of soil properties, and ii) indicate potential areas of research in the application of fractals to the field of spatial variability of hydraulic properties and infiltration.

Modeling Spatial Structure with Fractals

Spatial dependence of soil properties has been modeled as a fractional Brownian motion (fBm) or as a discrete fractional Gaussian noise (dfGn). These models are generalizations of Brownian motion (Bm) and white Gaussian noise, respectively.

A fractional Brownian motion, $fBm \equiv B_H(x)$ is a function of space, x, that has normally distributed (Gaussian) increments, $B_H(x+h)-B_H(x)$, zero mean, and expected variance, $\langle B_H(x+h)-B_H(x)\rangle^2$, of the form:

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[1]USDA-ARS, Hydrology Laboratory, Bldg. 007, Rm. 104, BARC-West, Belstville, MD, 20705

$$\langle B_H(x+h)-B_H(x)\rangle^2 \propto h^{2H} \tag{1}$$

where <> denotes expected values, H is the Hurst exponent, a real number between 0<H<1, and h is the lag. A plot of Eq. (1) for several lags is called a semivariogram, a widely used geostatistical technique. A linear log-log semivariogram of a given property indicates fractal distribution of that property. Exact (mathematical) fractals result in unbounded semivariograms. Statistical (natural) fractals, on the other hand, have upper and lower bounds to the distribution that can be observed in semivariograms of soil properties (Burrough, 1983; Huang and Bradford, 1992). A fractal model can only be applied within the scale range enclosed by the upper and lower bounds. The H coefficient can be related to the autocorrelation of the function . Setting $B_H(0)=0$, the autocorrelation between past,-$B_H(-x)$, and future, $B_H(x)$, increments is (Mandelbrot, 1983):

$$\frac{\langle -B_H(-x)B_H(x)\rangle}{\langle B_H(x)\rangle^2} = 2^{2H-1} - 1 \tag{2}$$

A H=0.5 results in uncorrelated increments (Bm); whereas for H≠0.5, data in a sequence is correlated. For H<0.5, correlations are negative (antipersistence), while a H>0.5 indicates positive correlations (persistence) (Feder, 1988). Persistence/antipersistence is related to the range of effect that an event will have on the following events. Persistence, implies that an average trend in the past is likely to continue in the future (long-range effect). Antipersistence, on the other hand, results in average trends continually reversed (short-range effect). Prediction of an unknown value is, therefore, more difficult for series that show antipersistence (Burrough, 1983).

A discrete fractional Gaussian noise (dfGn) is defined as the sequence of increments of a fBm (Korvin, 1992):

$$X_x^H = B_H(x)-B_H(x-1) \tag{3}$$

where dfGn$\equiv X_x^H$ with the same exponent H, and therefore the same correlation properties as fBm. Discrete fGn has been used to model long geophysical records using the rescaled range method, R/S, (Feder, 1988; Mandelbrot and Wallis, 1995):

$$R(h)/S(h) \propto h^H \tag{4}$$

where R(h) is the range between the minimum and the maximum values of observations as a function of a lag h, and S(h) is the standard deviation over the same h. In all cases a fractal dimension, D, is obtained as D=2-H.

It should be noted that in presence of a spatial sequence of measurements, one can model it by assuming either a dfGn or a fBm process. Unfortunately, the choice of either model will result in different values of H (Moltz and Boman, 1995). A discrimination between a fBm and a dfGn can be achieved by performing a power spectral analysis (Hough, 1989; Korvin, 1992):

$$S_z(f) = 4\int_0^{\infty} C_z(h)\cos(2\pi f h)\,dh \tag{5}$$

where f is frequency, and $C_z(h)$ is the autocovariance function of process z: $C_z(h) = <z(x+h)z(x)> - <z(x)>^2$. A log-log plot of power spectrum vs frequency resulting in a straight line indicates a fractal distribution; with the slope of the line, $-\beta$, related to H. For dfGn, β=2H-1 while for fBm, β=2H+1 (Tubman and Crane, 1995).

Alternative Models

Recently, Painter and Paterson (1994) proposed a new approach to model spatial variability in sedimentary formations, that is a generalization of a fBm. The new model called fractional Lévy motion (fLm) does not require a Gaussian distribution of increments, instead it uses Lévy-stable distributions that are more appropriate to model processes made up of several independent random processes and that, therefore, show a more abrupt variability.

Iturbe et al. (1995) studied the spatial structure of soil water content in an area of 848 km^2 using remotely sensed data. They found that soil water content was a stochastic self-similar process. For those processes, the expected value of a moment of order q at a scale factor λ, $<Y_\lambda^q>$, scales as (Gupta and Waymire, 1990):

$$<Y_\lambda^q> \propto \lambda^{q\kappa} <Y_1^q> \tag{6}$$

where the scaling exponent κ is a linear function of q for fractal processes such as a fBm. Processes that are multifractal have a scaling exponent that is a nonlinear function of q.

Typically, all of the above methods average spatial information, for instance a semivariance is calculated using all points in a data set separated by a given lag, regardless of the position of the points. Similarly, a Fourier transform contains information on frequency but not on space. Wavelet analysis, on the other hand, preserves spatial information by using building blocks containing information on local frequency (Chui, 1992). This property makes wavelets a promising tool to study multiscale processes (Li and Loehle, 1995; Muzy et al., 1994).

Applications of Fractals to Spatial Distribution of Soil Properties

Typically, fractal studies on the spatial distribution of soil properties has been done through the analysis of semivariograms (Eq. (1)) originally designed for either classification and mapping or for soil/water management studies. Separation intervals for the former are on the order of several meters (Burrough 1983; Bartoli et al., 1995), whereas for the latter separation intervals are about 0.5 to 2 m (Vieira, 1981; Mohanty et al., 1991; Mohanty et al., 1994). Burrough (1983) obtained fractal dimensions of spatial distribution of pH, clay and silt content, and thickness of the A horizon from semivariograms. In general, fractal dimensions were larger than 1.5 (H<0.5) indicating short-range correlations (antipersistance). He concluded that soil variation is more irregular than landform, river discharge, geological sediments or climatic data variations (Burrough, 1981; 1983). Culling (1986) reported fractal dimensions between 1.7 and

1.9 for pH variations in soil. Spatial distribution of shear strength resulted in fractal dimensions between 1.7 and 1.9 (Armstrong, 1986; Folorunso et al., 1994) increasing to values close to 2 on paddy fields (Pan and Lu, 1994). Bartoli et al. (1995) obtained fractal dimensions of clay, water dispersed clay, and water stable aggregates from a transect crossing three pedological units. They found that a fractal dimension varied among pedological units.

Moltz and Boman (1993, 1995) analyzed vertical distribution of log-transformed K_{sat} from three sites with the R/S method with lags between about 1.5 and 15 m. They found distributions were fractal with H values ranging between 0.67 and 0.82. For the same data set, however, persistance (H>0.5) or antipersistance (H<0.5) can be obtained depending on whether the assumed model is a dfGn or a fBm, respectively (Moltz and Boman, 1995).

Kemblowski and Chang (1993) studied distributions of log-transformed K_{sat} in a horizontal and in a vertical transect spanning 25 and 8 m and separation intervals of 0.5 and 0.2 m, respectively. By fitting semivariograms, they found D=1.48 and D=1.83 for the horizontal and vertical transect, respectively. Those values of D fitted well the spectral density function. The slope of the line was calculated as β=2H+1, indicating vertical and horizontal distributions of K_{sat} of a fBm type.

Mohanty et al. (1991) measured K_{sat} at 66 locations (4.6 m apart) and two depths (0.15 and 0.3m) using field (Guelph permeameter) and laboratory techniques. The soils were derived from glacial till and subjected to no-till. A log-log plot of their original semivariogram of log-transformed field measured K_{sat} (0.3 m depth) resulted in a D=1.87 (Fig. 1). The semivariogram shows a certain periodicity that could indicate random processes acting at different scales (multifractal distribution of K_{sat}).

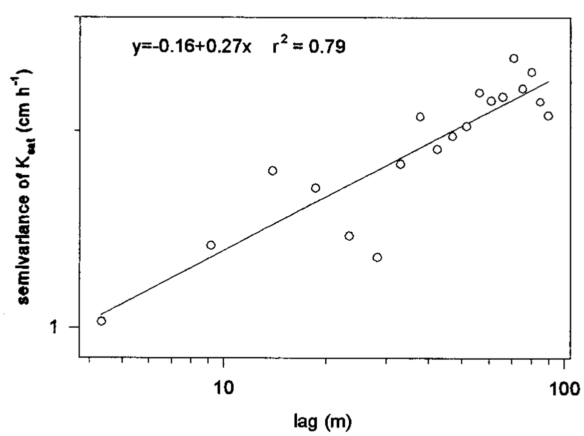

Fig. 1. Semivariogram of ln K_{sat} measured in the field at 0.3 m depth (data by Mohanty et al., 1991). The slope of the fitted line gives 2H=0.27 and D=1.87.

Lauren et al. (1988) studied the effect of sample size and spatial distribution on K_{sat} values on 37 sites along a 370-m transect (sites 10 m apart) on a argillic horizon. Samples were rectangular columns with sizes (m): A (1.60x0.75x0.20), B (1.20x0.75x0.20), C

(0.50x0.50x0.20), and circular columns with sizes (m): D (0.20x0.20) and E(0.07x0.06). Visible planar and cylindrical voids were traced on acetate sheets. Only columns B and C showed spatial correlation. Table 1 shows fractal dimensions that we calculated by fitting Eq. (1) to log-log semivariograms of K_{sat} (columns B and C), and void area (tubular and cylindrical voids).

Table 1. Fractal dimensions, D, calculated from semivariograms of K_{sat} and void area presented by Lauren et al. (1988). Measurements of K_{sat} were made on 0.2-m-heigh columns with an area of B: 1.20x0.75 m, and C: 0.50x0.50 m, respectively. Cylindrical and planar refer to pore shape.

	K_{sat}		void area	
	B	C	cylindrical	planar
D	1.81	1.92	1.79	1.82
r^2	0.72	0.74	0.86	0.86

The values of D indicate short-range correlation, something to be expected in well-developed soil structure such as the one present in an argillic horizon. The lack of spatial correlation for the smaller samples (columns D and E) is probably caused by an excessively large (10 m) lag with respect to the size of the samples. An alternative analysis is to use Eq. (6) to study the scaling of the standard deviation of measurements obtained on increasingly larger samples, σ_i, with i=1..5. Fig. 2 shows a log-log plot σ_i / σ_1 as a function of soil volume (Tables 1 and 2 in Lauren et al., 1988). From the slope of the line, a D=1.84 is obtained. This value is in agreement with D values found from semivariogram of K_{sat} and void area (Table 1).

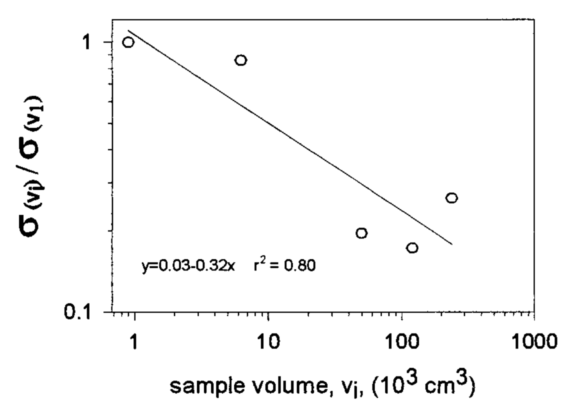

Fig. 2. Scaling of standard deviation, σ_i, of K_{sat} measurements as a function of soil volume v_i (data by Lauren et al., 1988). The slope of the fitted line is 2H=0.32 and D=1.84.

Spatial distribution of infiltration rates was studied by Vieira et al. (1981) by performing 1280 measurements at 1 m spacing, distributed in eight columns with 160 measurements/column. Measurements were on a alluvial fan of the Yolo soil series. A log-log plot of the semivariogram of the 1280 measured infiltration rates is presented in Fig. 3. The scatter in the plot could be due to directional variations usually found in an alluvial fan. Despite variability in the measurements a straight line fits the log-log semivariogram up to a lag of about 30-m, resulting in a fractal dimension of D=1.72.

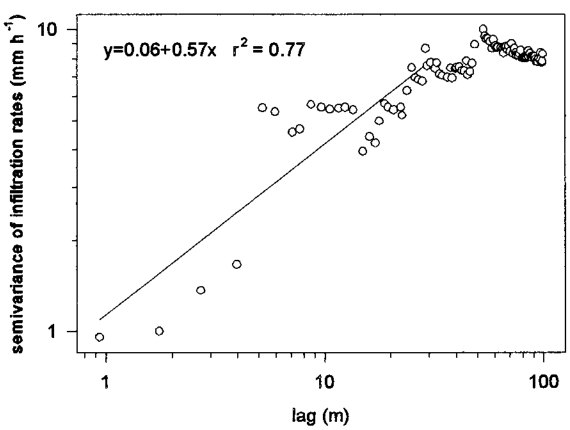

Fig. 3. Semivariogram of infiltration rates (data by Vieira et al., 1981). The slope of the fitted line gives 2H=0.57 and D=1.72.

Sisson and Wierenga (1981) studied infiltration rates in an area of 6.35x6.35 m. The soil was a Typic Torrifluvent consisting of a surface layer (about 0.60 m thick) of fine silty clay loam. Measurements were made with three ring diameters (0.051, 0.254, and 1.27 m), with a larger ring enclosing five of the immediately smaller diameter rings. Five 1.27-m rings were in each of the five columns and rows, resulting in a total of 125 measurements with the 0.051-m ring per column/row. A power spectrum of measurements with the 0.051-m ring carried up to a lag of 40 (32% of the data) plotted in a log-log scale is shown in Fig. 4. Each point is an average of five realizations. The fitted straight line has a slope of β=0.47 indicating a dfGn distribution of infiltration rates with a D=1.27.

In conclusion, spatial dependence of soil properties is well documented and can be represented using fractal models. Almost invariably fractal dimensions of spatial distribution of soil properties are > 1.5 indicating short-range effect. Separation intervals in the order of a few centimeters resulted in D<1.5 and a distribution of infiltration rates of the dfGn type (Sisson and Wierenga (1981). Smaller separation intervals are likely to reveal structural influences on soil processes, whereas larger separation intervals are likely to capture variation caused by landscape/soil formation features.

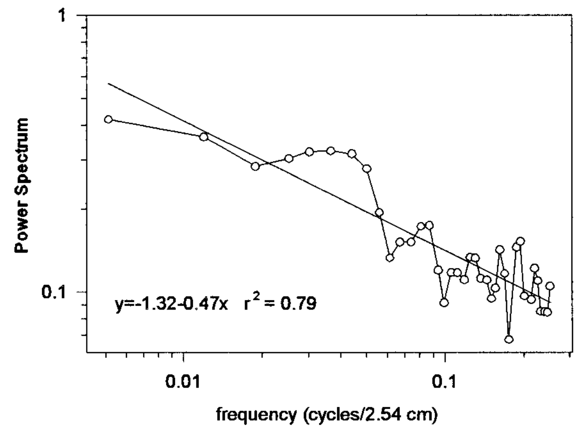

Fig. 4. Power spectrum of infiltration rate (data by Sisson and Wierenga, 1981). Slope of the fitted line β=0.47=2H-1 gives D=1.27.

Potential Areas of Research

- Soil variability is the result of nested processes acting at different scales (Burrough, 1983). Such systems are more likely to be resolved by a multifractal approach (Folorunso et al., 1994; Bartoli et al., 1995). Multifractal techniques should be tested in modeling spatial dependence of properties/processes.

- An important area of research is in finding indirect methods to characterize soil spatial variability. A possibility is to study relationships between fractal dimensions defining distribution of different soil properties and those defining spatial distribution of soil processes. For instance, macropores are related to K_{sat}, infiltration rate, and by-pass flow. The question remains as to how the distribution of those properties relate to each other. An example of such relations is given in Table 1. Hatano et al. (1992) related solute transport to the fractal characteristics of stained patterns of dyes. This type of approach needs to be generalized to field conditions and expanded to account for spatial variability.

- Measurements of a soil property on different soil volumes is an interesting alternative to regularly spaced measurements, especially for defining short-range variations (Fig. 2). This technique could be complemented with other methods that use spatially distributed measurements.

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Specific Goal 3

Characterizing Temporal Variability

Temporal Variability: Tillage, Reconsolidation, and Compaction Effects on Infiltration Parameters

Walter J. Rawls[1]

Abstract

The temporal effects of tillage, reconsolidation and compaction on infiltration parameters has normally been accounted for by changes in bulk density of the soil which can be translated in changes in the soil porosity. A literature review was conducted summarizing the environmental and management effects on bulk density Also, included are the effects of changing bulk density on soil hydraulic properties such as the water retention curve and saturated hydraulic conductivity and current methods to model temporal changes in bulk density. In conclusion a list of future research needs to be able to adequately describe the temporal variability of tillage, reconsolidation, and compaction effects on infiltration parameters.

Introduction

Factors which influence the infiltration process have been grouped into the following categories: soil, soil surface, management, and natural factors (Brakensiek and Rawls, 1989). Agricultural management (grazing/tillage systems) affect all the categories; however, the primary factors affected are vegetation, surface cover, and soil properties such as bulk density, structure, organic matter, and organisms. The temporal effects of tillage, reconsolidation and compaction on infiltration parameters has normally been accounted for by changes in bulk density of the soil which can be translated in changes in the soil porosity (Rawls, et al 1991). This has a major effect on hydraulic soil properties such as water retention characteristics and hydraulic conductivity. Soil bulk density is affected by many processes such as root growth, crop growth, runoff, rainfall, wind, water erosion, freeze-thaw, wetting and drying, weathering and disturbances made by man and animal such as tillage and grazing. (Cassel, 1980; Porter and McMahon, 1987; Jones, 1983).

Literature Review

Management Effects on Bulk Density

Most infiltration studies have examined the influence of various management practices on infiltration during a standard storm at a specific time of the season (Burwell and Larson, 1969; Dexter et al, 1983; Onstad, 1984; Roth et al, 1988,). Several studies have considered the changes in the soil surface with time such as soil roughness and depression storage (Blake and Gilman, 1970; Freebairn et al 1989, Murphy et al. 1993). Unger and Kasper (1994) reviewed the effect of compaction on root growth.

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[1]USDA-ARS, Hydrology Laboratory, Bldg. 007, BARC-West, Beltsville, Maryland 20705-2350.

The effect of tillage on soil bulk density has been demonstrated by various investigators (Cassel, 1980; Choi et al, 1988; Zobeck and Popham, 1990; Onstad et al, 1984; Onofoik, 1988; and Bauer and Black, 1981) and it is the consensus that tillage reduces bulk density. The magnitude of the decrease varies according to tillage characteristics, soil properties, and soil water content. Rawls et al. (1983) compiled data on the effect of tillage systems on porosity and related it to soil texture. They concluded that the porosity increase caused by tillage increased from clay to sand with a maximum increase of about 25%. They also reported that the more pulverized soils exhibited a smaller decrease in porosity and plow-disk-harrow, rotary, plow/pack tillage increased porosity less than a plow system.

Radke and Berry (1990) used infiltration measurements to detect changes in soil properties caused by tillage, cropping systems and grazing patterns. Significant changes in infiltration did not occur until four years after converting from a conventional to a low input cropping system. They reported a decrease in infiltration with increases in stocking rates for the summer and over winter indicated no pattern. Onstad et al, 1984 showed for tilled soils that bulk density increased exponentially with the first 10 cm of rainfall after which there was little change. Tillage can also cause tremendous spatial variability because it creates areas which have been compacted by wheel tracks and areas which have been fluffed. Pikul et al (1990) reported on the effect of chisel and paraplow tillage and over winter had on macroporosity. The macroporosity was about the same for both treatments for the total profile; however, each had a different distribution with depth. Also, the macropore structure was stable over the winter. In recent years there have been extensive amounts of research conducted on describing the variability of infiltration between in-row, between-row and traffic-row. With the advent of the tension infiltrometer there is an improved ability to study the pore structure of these conditions.

Brown et al. (1992) found that steel tracked crawler caused lower bulk densities than wheeled tractor and rubber tracked crawler caused bulk densities between steel and wheeled tractors. They also reported as others that the trafficked interrows had higher bulk densities than the non trafficked rows.

Environmental Effects on Bulk Density

The effect of frozen ground on infiltration has primarily been related to the water content at time of freezing (Kane, 1980; Lee, 1983). Rawls and Brakensiek (1989), using Lee's data, developed a relationship to predict a frozen ground correction for saturated hydraulic conductivity based on the ratio of water content at freezing to 1/3 bar water content. Pikul et al (1996) showed that soil ripping when the soil was frozen significantly increased the infiltration. The effect of freezing and thawing cycles on bulk density has infrequently been reported.

Soil crusting is a natural compaction of a thin layer of the soil at the surface. Onstad (1984) showed that the infiltration rate of a bare soil stabilized after 10 cm of prior rainfall indicating the development of a stable crust. Rawls et al (1995) showed that a crust could decrease the infiltration rate by 50% or more. There has been extensive process modeling of water flow through a surface crust (Sumner and Stewart, (ed) 1992).

Modeling Temporal Changes in Bulk Density

One approach taken to predict soil bulk density is to relate soil bulk density to static soil properties. Jones (1983) discussed the effect of soil texture on the critical bulk density for root growth. Heinonen (1977) used clay, fine silt to predict bulk density for soils with little organic matter. For soils with more organic matter weight losses at 700 and 500 degrees C have been used (Curtis and Post, 1964); Jeffrey, 1970). Rawls (1983) used sand, clay and organic matter; while Alexander (1980) used organic carbon, 15 bar water content, the ratio of 15 bar water to clay, silt and sand contents, parent material, calcium carbonate equivalent and mean soil horizon depth. Mausbach (1984) analyzed over 2000 bulk densities for midwestern soils and found that parent material, texture, and pedogenic processes were important in stratifying bulk density. The NRCS (personnel communication R. B. Grossman) has a procedure for estimating bulk density to fill in for missing data in their soil property tables. The procedure is based on morphological description, knowledge of the parent material, organic carbon, COLE and extractable iron.

For compacted areas Blackwell and Soani (1981) and Gupta and Larson (1979) present models on how to describe the effect of compaction. Gupta and Allmaras (1987) reviewed the status of physically based compaction models and concluded that soil compaction modeling had developed to where it could be useful. However, the lack of sufficient data bases on the limiting variables covering a wide range of soil types, implements, and crops is the reason they have not been sufficiently verified.

Allmaras et al (1967) were one of the first to attempt to model the temporal bulk density effects. Since then Williams, et al. (1989), Flanagan and Nearing (1995), and Porter and McMahon (1987) have developed models to describe the temporal variability of bulk density of tilled soils. The WEPP model (Flanagan and Nearing, 1995) uses an approach similar to that developed by Williams et al. (1987) where the tilled bulk density is predicted based on a predicted consolidated bulk density and a tillage disturbance factor. They have regression functions for consolidating the tilled bulk density for rainfall, weathering, coarse fragment, and time effects. Most of the models which have been developed are based on limited soils data base. These models have incorporated the effects of rainfall, weathering and soil water content on bulk density; however, no attempt has been made at incorporating the spatial tillage effects such as in-row, between row and traffic row, effects of plant roots, effect of depth into the models.

Effects of Bulk Density on Soil Water Properties

The effects of temporal changes of bulk density on the water retention and hydraulic conductivity characteristics of the soil have primarily been accounted for in the estimation of the characteristics based on soil physical properties (Rawls, et al., 1991). Figures 1 and 2 were developed using equations developed by Rawls, et al. (1991) and illustrate how the water retention curve is affected by a change in porosity. Also, almost every soil based procedure for estimating the soil matrix saturated hydraulic conductivity includes the soil porosity and Fig. 3, developed using equations developed by Rawls, et al. (1983), illustrates the effect of porosity changes on the saturated hydraulic conductivity. Tillage has an extremely severe effect on macroporosity and most prediction methods require a pore size distribution (Rawls, et al, 1996) which has not been characterized temporally.

In addition to the temporal effects of bulk density on hydraulic conductivity, the temporal effects of vegetation and surface cover have normally been incorporated empirically into the hydraulic conductivity by developing regression equations relating cover characteristics to conductivity (Rawls and Brakensiek, 1989; and Flanagan and Nearing, 1995).

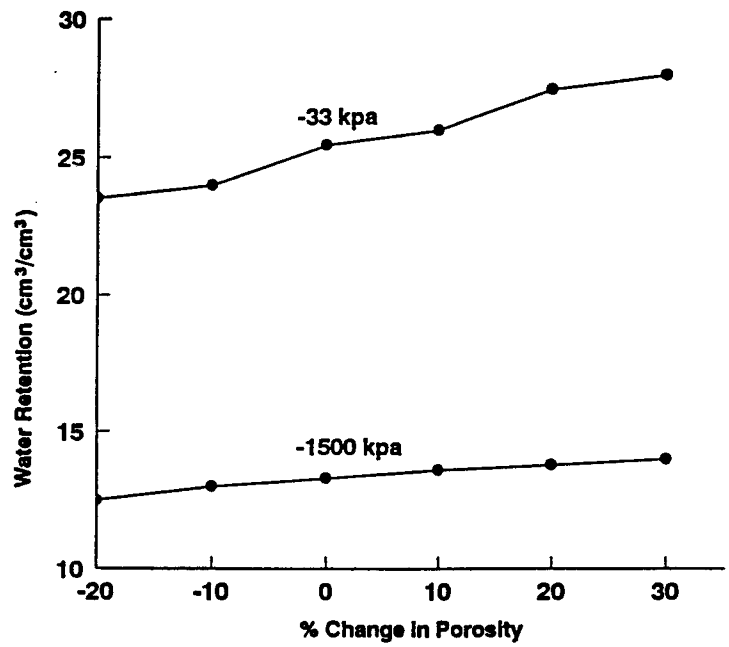

Fig. 1. Effect of changes in porosity on water retention.

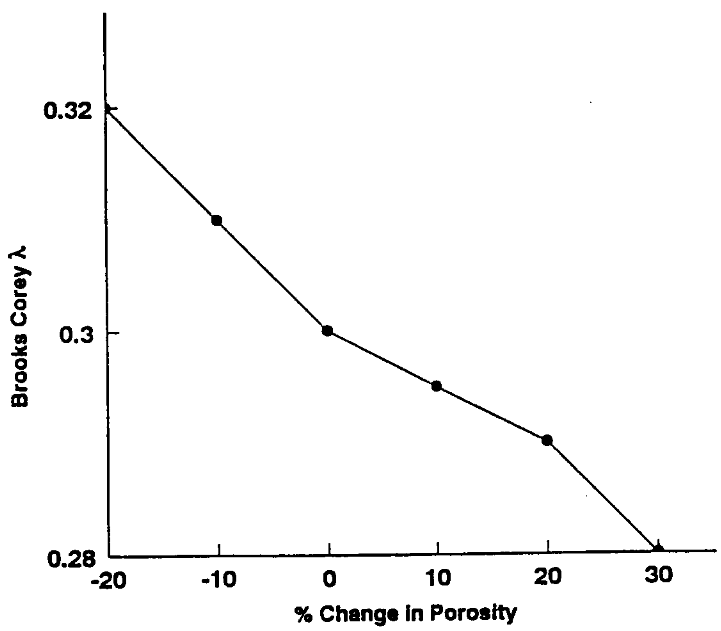

Fig. 2. Effect of changes in porosity on Brooks-Corey λ.

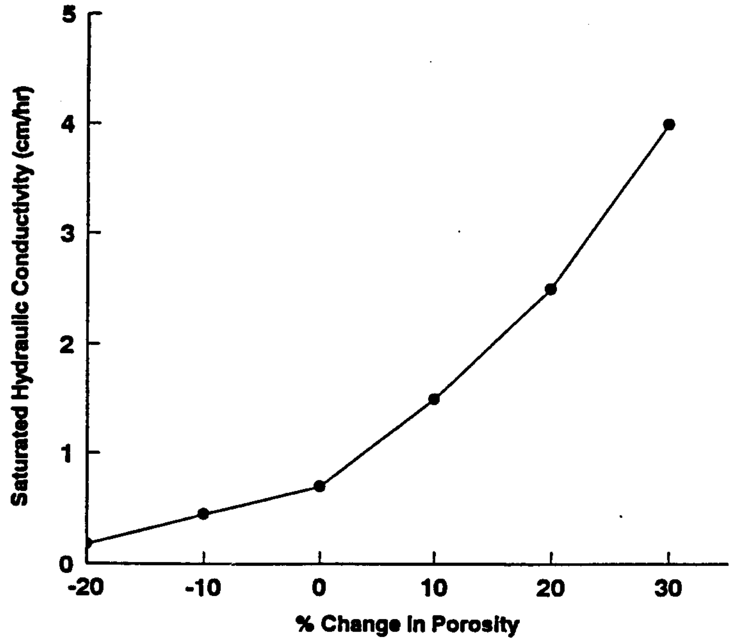

Fig. 3. Effect of changes in porosity on saturated hydraulic conductivity.

Future Research

Many infiltration experiments have been conducted from which inference has been made on the temporal effects of tillage on infiltration; however, infiltration measurements integrate many factors making it almost impossible to isolate the effects of bulk density (Radke and Berry, 1993). Another problem is that bulk density is not a good descriptor of soil structure related to water flow because it does not describe the pore size and distribution which affect the rate of water flow through the soil (Pikul, et al, 1990). The bulk density models which describe temporal variability are a combination of regression equations which have been developed from limited data. The following are the critical research needs for describing the temporal variability of tillage effects, compaction and reconsolidation on infiltration parameters:

- Develop a soil structure parameter which is related to water flow which incorporates the effects of freeze-thaw, biological activity; roots, depth, and compaction.
- Develop a data base on the effects of farming implements on soil structure.
- Incorporate into present models of bulk density process soil compaction models.
- Develop methods for integrating the spatial variability caused by tillage into a water related soil structure parameter.
- Develop methods for characterizing the temporal variation of macropores.

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The Characteristics and Temporal Variability of Infiltration as Affected by Macropores and Plants

J.L. Starr[1], D.R. Linden[2], S.D. Logsdon[3], and I.C. Paltineanu[1]

Abstract

Macroporosity and bioactivity, including plant growth, are inseparably intertwined with infiltration, and are the major sources of temporal variability in the rates of water infiltration. The larger pores of the soil fabric are dynamic and change with both natural and man made activity, causing most of the temporal variability of infiltration. These pores are largely created by bioactivity (roots and burrowing fauna), but may also be created by tillage operations, soil shrinkage, and aggregation. Numerous research papers have shown that temporal variability of the soil far exceeds spatial variability across a landscape and should be considered as a major area of future research. Vulnerability of the soil fabric to change and the agents which cause changes are important components of temporal variability. The less stable the soil fabric and the more exposed a soil is to the agents of change, the greater will be the observed temporal change. Some agents of change act quite rapidly such as a plow or a tractor, while others act quite slowly such as earthworms and roots. Recommended high priority research areas include to: improve and simplify measurements so that meaningful data can be collected; expand our infiltration data base to include the important temporal changes of soils and landscapes; and improve models to account for these temporal changes.

Introduction

Macroporosity and bioactivity, including plant growth are the major sources of temporal variability in infiltration. The larger pores of the soil fabric change and cause most temporal variability of infiltration. These pores are largely created temporarily by bioactivity (roots and burrowing fauna), but may also be created by tillage operations, soil shrinkage, and aggregation. They occur in forest, crop, and rangelands with varying degrees of importance. These pores are often destroyed, changed, or decreased in size due to animal or vehicle traffic and natural reconsolidation during wetting.

Temporal Variability of Infiltration

The most common temporal variability in infiltration rates is associated with the reduction in total porosity due to forces such as rainfall and traffic. At different times total soil porosity may also increase, with concomitant increased capacity to infiltrate water. Crop growth with

[1]USDA-ARS, ECL, Beltsville, MD 20705

[2]USDA-ARS, SWMU, 1991 Upper Buford Circle, St. Paul MN 55108

[3]USDA-ARS, NSTL, 2150 Pammel Drive, Ames, IA 50011

canopy development and root growth and decay are perhaps the most common of these. Shrinkage of the fabric to form cracks is also an important temporal change as well as the burrowing of soil fauna (earthworms, ants, termites, enchytrids, insects, etc).

There are also sources of variations not associated with the physical changes of the soil fabric such as the driving gradients for downward movement and the storage space available for infiltrating water. Temporal variability has sometimes been mistakenly reported as these natural time decrease in the infiltration rate associated with decreased potential gradients. These decreased rates have been described by both theoretical and empirical equations and are not the major limitation to our ability to describe and understand the temporal aspects real world infiltration. Natural soil drying (Van Es et al., 1991) and water repellency (Witter et al., 1991) may cause apparent or real temporal variability of infiltration. Spatially variable and temporary frost layers may also cause considerable variations of infiltration (Blackburn et al., 1990).

Infiltration variations, compared across time periods may also be due to events or seasons with more rainfall which results in a greater opportunity for infiltration and water movement into and or through the soil. Other important source of infiltration opportunity are variations in snow trapping, water storage in surface depressions, and stem flow.

Some spatial variability is often associated with differences in the "temporal state" of various positions in the system. For example, row interrow differences can be considered as major differences in the temporal scale instead. During a natural time course, these differences often become less obvious. The more vulnerable to change, the quicker the row becomes more like the interrow. Exceptions to these generalities occur when root or faunal activity of the row increases the fabric space for rapid infiltration and the canopy protects this region from change due to raindrop impact.

Vulnerability of the soil fabric to change and the agents which cause changes are important components of temporal variability (Van Es, 1993). Temporal variability will be most pronounced in systems or parts of systems that are most vulnerable to this fabric change. For example, a field that is freshly tilled is highly vulnerable to change and temporal variability of the soil fabric should be considered as the normal course of action. This same field after planting will have at least two zones with varying degrees of vulnerability. Behind the wheel tracks the vulnerability has been reduced because of the compressive reconsolidation, while the areas between tracks remain more vulnerable and will change during the course of the season. Some agents of change act quite rapidly such as a plow or tractor, while some agents act slowly such as earthworms and growing or decaying plant roots.

Numerous research papers have confirmed the idea that temporal variability of soil fabric far exceeds spatial variability across a landscape. Thus temporal variation should be considered as a major area of future research.

Characteristics of Macropores and the Nature of Their Variability

Several recent reviews have documented the relation of infiltration to soil properties, including macropores (Bouma, 1992; Kutilek and Nielsen, 1994; McCoy et al., 1994; Topp et al., 1992). Macropores have been given many definitions, including characterization by size, pores with cylindrical diameters >1 mm (Luxmoore, 1981); by tension, exhibiting channel flow at soil water pressures > -0.3 kPa (Luxmoore, 1981); by infiltration rate (Chen et al., 1993; Beven and Germann, 1982); and by their descriptive geometry. Macropores may be formed in soils by biological (e.g., roots, earthworms, ants, etc.) and physical (shrink/swell, freeze/thaw, and tillage) factors. Biopores tend to be fairly stable in time due to slow decay of plant roots and stabilizing secretions by earthworms along edges of burrows while porosity caused by tillage is quite unstable.

Soil porosity is often dynamic, being greatly affected by many external and internal factors. Primary tillage causes an increase in total pore space (Van Duin, 1956; Gantzer and Blake, 1978), not only increasing porosity near the soil surface, breaking the crust, but also by aiding in the formation of soil aggregates (Youngs, 1995; Van Duin, 1956). Freezing/thawing and wetting/drying also influence pore arrangement but not to the same extent as tillage (Voorhees and Lindstrom, 1984). Increased soil porosity due to tillage operations and temporally varying arrangement and size-distribution of soil pores, results in temporarily increased infiltration rates because greater space within the soil is available for water flow. These tillage-induced higher infiltration rates are of relatively short duration because the pore space begins to decrease as rearrangement and reconsolidation of soil particles occurs with exposure to rainfall or irrigation (Mapa et al., 1986). Infiltration studies on tilled and untilled soils indicate that infiltration in the early portion of a rain-storm is drastically affected by the cultural state but that infiltration after exposure to rainfall is affected much less (Unger, 1992; Burwell and Larson, 1969). After the initial tillage-induced higher infiltration rates decline and stabilize, additional tillage may decrease infiltration rates because the naturally occurring, quasi stable, large biopores have been disrupted (Quisenberry and Phillips, 1976; Ehlers, 1975).

The contribution of macropores in freshly tilled fields to infiltration can diminish rapidly as surface openings fill with washed-in soil and the soil settles (Ela et al., 1992; Dao, 1993). These surface seals can develop quickly if "heavy" rainfall occurs with sparse surface residue or before the plant canopy has closed. In contrast, in no-till crop land and pastures with an extensive network of earthworm channels and surface (corn brace roots) and subsurface roots, the macroporous contribution remains relatively constant throughout a typical rainfall event. In freshly tilled soils, closure is rapid while for sodded soil, closure may be small indicating a very stable soil structure. It is primarily this closure that causes temporal variation in macropore infiltration. In practice it has been observed that major changes in soil structure occurs as the soil near the surface approaches saturation. Such phenomena is probably also the cause for the closure of macropores (Berg and Ullersma, 1994).

Importance of Temporally Variable Macropores to Infiltration.

Tillage, residue placement, cover crops, land shaping (furrows, beds, etc.) and other cultural state conditions influence infiltration, (Unger, 1992; Dao, 1993; Freese et al., 1993; Duley and Kelly, 1941); yet a single physically based description of the effect of cultural state on infiltration remains unavailable. Tillage, as an example of cultural state, increases or decreases infiltration(Mwendera and Feyen, 1993; Unger, 1992; Ehlers, 1975; Logsdon et al., 1993a; Messing and Jarvis, 1993) and increases or decreases soil water storage (Hamblin, 1984; Wagger and Denten, 1992) depending on soil and climatic conditions.

Many soils, especially those that have not been disturbed for a period of time, have large diameter biochannels or cracks that occupy only a fraction of the total pore space of the soil (Gantzer and Blake, 1978; Luxmoore et al., 1990). These soils also have been shown to exhibit cultural state infiltration effects in the form of bypass infiltration. These soils also may exhibit a transitory nature to infiltration characteristics because of the changes that occur within the large void spaces or because of the non-uniform water supply and entry near the soil surface. Changes can occur which either close or open macropores to infiltrating water and that change the surface distribution of water available for infiltration (Bonell and Williams, 1986; Cassel and Nelson, 1985; Federov and Marunich, 1989; Jorgensen and Gardner, 1987; Logsdon, 1993; Neary et al., 1993; Thurow et al., 1993; Van Den Berg, 1989; Williams and Bonell, 1988). Experimental evidence is strong that suggests that bypass flow (infiltration which is not described by classical diffusion type flow theory) is important in real world infiltration (Quisenberry and Phillips, 1976; Dixon and Peterson, 1971; Thomas and Phillips, 1991). Early infiltration investigators observed macropores but minimized their importance by discarding samples with large voids as not representative, or by using disturbed samples (Free et al., 1940; Slater and Byers, 1931). Evidence also comes from laboratory studies that have shown high correlation of permeability with noncapillary porosity (Baver, 1953; Van Doren and Klingebiel, 1949). The influence of macropores on infiltration has also been obtained by: directly observing dye stained flow routes (Linden and Dixon, 1976; Ehlers, 1975), rapid reductions in soil water potential at soil depth , drastic reductions in infiltration rate with very small air pressures opposing the water flow (Linden and Dixon, 1976), and chemical tracer studies (Quisenberry and Phillips, 1976; Blake et al., 1973). A wide variety of soils and infiltration conditions, including artificial rainfall (Linden et al., 1977; Blake et al., 1973), have also exhibited macroporous flow features.

Infiltration into many soils is greatly affected by their multi-modal pore-size distribution, with the soil matrix representing the continuous part of the pore distribution and the macropores representing the somewhat abrupt increase in pore sizes at near zero soil water potentials. The soil's inherent matrix infiltration characteristic may involve intra- and inter-aggregate flow, which is largely controlled by the materials that make up the soil matrix. Superimposed on this system is a macroporous flow system which can be described by pipe type flow which can rapidly conduct water below the soil surface becoming available to the soil matrix for absorption by capillary type water movement.

A Conceptual Model of Infiltration

A conceptual representation of a structured soil system has geometrically-complex and irregular macroporous space into which infiltrating water can be introduced. This space may be simple cracks, nightcrawler burrows, or the more complex network of root biopores and interclod spaces induced by tillage. The flow geometry, the potential driving forces, and the boundary conditions may be extremely complex in the system and can often only be approximated by making many simplifying assumptions. The flow geometry of the macroporous system is as complex as the microporous system and must of necessity be represented by an equivalent system which can be described in terms of effective parameters. Simplified macroporous flow systems (single cylindrical pores) have been described and studied (Edwards et al., 1979). Although strong correlation has been shown between soil morphological features and saturated hydraulic conductivity (McKeague et al., 1982; Logsdon et al., 1990; Bouma, 1992), it is often necessary to describe the soil in terms of its effective ability to conduct water and not in terms of a specific geometry of the macropore system. This conceptual model is applicable to flow geometries, such as earthworm channels, exposed plant stems and roots in no-till, cracks, and the irregular network of macropores created during tillage.

This conceptual model includes the processes of microporous flow, surface sealing, macroporous inflow, outflow from macroporous space to matrix space and vice versa, macroporous closure and the interactions between each process. The generic macropore and those portions of the system which cause the temporal variations of infiltration are of particular concern to this discussion. The underlying assumption of this conceptual model is that the geometry of the soil fabric or macropores is complex so that it can be described only in terms of its effect. Macroporous infiltration is superimposed upon the microporous system. Infiltration via the macropore system is viewed as having a much higher conductivity, large void spaces in which flow occurs, higher flow velocities, and smaller potential losses across a small depth increment than adjacent and intermixed microporous infiltration.

Limitations to macroporous infiltration often involve the interactions of several factors and not just the geometry of the pore system. Macroporous infiltration is often only a small fraction of the possible flow because the space functions at full capacity only when surface ponded water can reach the macropores in sufficient volume to maintain flow rates. On gently sloping dry and macroporous soils, hydrophobicity of the soil may induce surface ponding during intense summer storms even before the matrix has thoroughly wetted (Edwards et al., 1989). Macropores are less effective at low intensity rainfalls (Trojan and Linden, 1992). Also at constant rainfall intensities macroporous space is less effective on soils with high microporous affinity for water (Edwards et al. 1989).

Macroporous infiltration may also be limited by geometrical consideration in the macropore space within soils. Constrictions, necks, partial closures, and abrupt ends to pores may limit the capacity for infiltration. Two conditions often credited with limiting macroporous infiltration are the lack of continuity near the surface, and the break in continuity at a tillage interface. The near soil surface layer becomes the most dynamic region for controlling temporal variations in infiltration.

Sources of Space-Time Variability

Aggregate Breakdown and Sealing

Kutilek and Nielsen (1994) state that the process and impact of sealing/crusting largely by rainfall/irrigation destroying surface soil aggregates was well documented. They also note that sealing depends more on drop intensity (energy) than rainfall amounts, and once the surface seal is developed the physical properties of the seal usually persist. Hence, measurements and data collection should be aimed primarily at the determination of the final saturated conductivity in relation to the energy of the rain, rather than on the changing infiltration rates during seal formation. Erosion during an event may increase infiltration because of the removal of the sealed layer (Gimenez et al.,1992).

Infiltration rates can be greatly affected by the measurement method. Ben-Hur et al., (1987), measured rates of infiltration at 30 random sites within a 1-ha field and reported nearly 7-fold greater infiltration rates from flood infiltrometers than with sprinkler infiltrometers. This difference was attributed to surface sealing of the soil from the impact of falling drops. They concluded for soils that seal it is impossible to predict infiltration rates under water drops from measurements under flooding conditions.

Cracking and Aggregation

Youngs (1995) reported a general lack of both infiltration theory and temporal observations of changes in soil structure. He stated that one consequence of swell and shrinkage in soils is the formation of cracks that separate micropore regions of aggregates from macropore channels. These “channels” can result in preferential flow through the soil when the macropores are filled with water, but when they are empty the aggregates become isolated with little water movement from one to the next. Water ponding on the soil surface may flow into these cracks without completely filling the cracks. This wall flow generally bypasses the aggregates. Youngs (1995) concludes that Richards’ equation cannot be applied to bulk soil in these circumstances so that classical infiltration solutions cannot describe the infiltration. Dual-porosity flow models are better suited for these soil conditions (Chen, 1993).

Agricultural Practices

Surface Residues and Amendments: Rawls et al., (1995) studied seasonal effects of canopy and residue cover on final infiltration rates of a loam soil by difference between rates of sprinkler application and plot runoff. They found that agricultural practices caused important seasonal effects on infiltration rates. For example, a 50% new residue cover minimized crusting resulting in large increases of early season infiltration rates compared to bare soils; the bare soil tended to recover infiltration rates under full canopy; residue benefits declined with time over the season; and residue + increasing canopy held infiltration rates steady from July to August. In an 8-yr study of the effects of three tillage and residue management practices with winter wheat on soil properties, including infiltration, Dao (1993) found that water infiltration was significantly higher under no-till than under plowed soil. Groundwater recharge generally occurred through macropores, directly wetting depths of 0.4 to 0.6 m of no-till soil, in contrast to a layered pattern in plowed soil. Martens and Frankenberger (1992) studied the effects of different organic amendments on soil physical parameters and water infiltration rates. After two-years of amendment, including straw, there was decreased soil bulk density (7-11%), and increased soil aggregate stability (22-59%), increased soil moisture content (3-25%), and increased infiltration rates (18-25%) in the organic-amended plots as compared with the unamended plots. Of the organic amendments, the straw amendment was statistically more effective in increasing soil aggregate stability and infiltration rates, and in decreasing bulk density in the tillage zone. Infiltration rates in the organic-amended soils were initially increased by stimulation of microbial activity, which increased the stability of soil aggregates. Cumulative infiltration rates were further increased by a decrease in soil bulk density with additional organic treatments to the tillage zone.

Crop and Crop Rotations: Meek et al., (1990) reported a 3-fold increase in infiltration rate for the first year in a sandy loam soil when alfalfa was grown following cotton, which continued to increase reaching a 4-fold increase (15 to 60 mm/h) after 3 years. Subsequent infiltration rates following planting of cotton by direct-plant into alfalfa under till and no-till systems were 4 to 9 times higher than normally measured for cotton in these soils (~12 mm/h). Water flow in the 5-yr-old alfalfa was determined to be mainly through the soil macropore system. Major infiltration variations are commonly observed to be caused by the amount and type of vegetation (Blackburn et al.,1992; Zhang et al., 1995; Mathier and Roy, 1993). Grazing intensity will also affect infiltration (Frazier et al.,1995).

Tillage, Cultivation and Traffic: A few studies have examined temporal variability in tillage systems using tension and ponded infiltration rates. Messing and Jarvis (1993) measured infiltration with tension infiltrometers on plowed and no-till plots on four dates. Generally $K(\Theta)$ was higher for June and October than for July or August due to recent tillage before the June and October measurements. Logsdon et al. (1993a) examined tension infiltration at four different times for four tillage systems and two crop sequences. The crop sequences (continuous corn vs. soybean/corn) had no influence on tension infiltration rates, and tillage only significantly influenced ponded infiltration rates. Tension infiltration rates were significantly affected by date; at a negative head of -30 mm, infiltration rates were significantly highest for September 1991, and significantly lowest on June 1991. Infiltration rates for July 1991 and May 1992 were intermediate. At a negative head of -60 mm, infiltration rates were higher for September 1991 than for June 1991; and at a negative head of -150 mm, infiltration rates were higher for May 1992 than for July 1991. Logsdon (1993) measured tension infiltration seven times during a growing season. Unsaturated hydraulic conductivity was significantly higher on 23 June, 11 August, and 1 September than on 14 July, and 24 July. The other dates were intermediate (17 June and 5 August). Increases and decreases were apparent over the growing season but were not directly related to initial soil water content. Logsdon and Jaynes (1996) observed variation in the spatial patterns of infiltration at four different measurement dates. Single ring ponded hydraulic conductivity was significantly higher for the late-season measurement and significantly lower for the post cultivation and pre-disk measurements. Hydraulic conductivity at a head of -30 mm was significantly highest for late-season and significantly lowest for pre-disk, whereas values for post-cultivation and post-disk were intermediate. Hydraulic conductivity at a head of -60 mm was significantly slower for pre-disk than for the other three measurement dates, and hydraulic conductivity at a head of -150 mm was significantly faster for late-season than for the other three dates. There was a spatial cycling period of 45.8, 146.9, and 96.2 m for post- cultivation, pre-disk, and post-disk measurements at a head of -150 mm.

Some studies have examined the temporal variability of ponded K in tillage systems. Mukhtar et al. (1985) measured double ring infiltration for four tillage systems at two or four different times on three locations. Treatment differences varied with measurement date. For the Central Iowa site infiltration was more rapid for the paraplow treatment than for chisel plow in May 1983, but in June 1983 was more rapid for paraplow than for moldboard plow and no-till.

For the South central Iowa site infiltration was more rapid for paraplow, moldboard, and chisel plow than for no-till in June 1983, but infiltration was more rapid for paraplow than for the other treatments in October 1983. For the Northeastern Iowa site infiltration on the paraplow treatment was more rapid than for moldboard plow in June 1983; infiltration was more rapid for paraplow and moldboard than for chisel plow or no-till in July 1983; infiltration on paraplow treatments was more rapid than for all other treatments in November 1983; and treatment differences were not significant one week later in November 1983. Jaynes and Hunsaker (1989) measured infiltration with single ring infiltrometers located within a flood irrigated field during four flood irrigations. Infiltration parameters were extremely variable and often uncorrelated over time. Starr (1990) measured double ring infiltration at eight different times. Messing and Jarvis (1990) observed that K_s changed seasonally as the soil shrank and swelled with highest rates in the summer when drying produced large cracks. The K_s was related to the volume of macropore cracks in the soil at a given time.

Intrinsic Plant and Canopy Effects

Stem Flow: Corn canopy has long been known to funnel water along the plant stem to the soil surface (Haynes, 1940; Glover and Gwynne, 1962; Paltineanu and Apostol, 1974; Parkin and Codling, 1990; Bui and Box, 1992). Direct quantification of this process (Paltineanu et al., 1995) revealed that preferential water flow down the plant stem at full canopy was highly correlated to the amount of rainfall (0.3 to 47 mm per event). Stem flow accounted for about 100% of the water reaching the soil at low rainfall events, decreasing to about 40% at 47 mm rainfall.

Brace Roots: Growing brace roots of corn, in both no-till and plowed soils, lifts up the crust of the silt loam soil, and opens the surface soil for receiving large amounts of water (Paltineanu et al., 1995). The total number of brace roots, confined into a 20-cm diameter area surrounding the corn stem, was twice as large under no-till vs plowed. Ponded infiltration rates on top of one-year-old corn stems was 10-fold greater than on adjacent traffic interrow positions. Similar results were observed Prieksat et al., (1994) for ponded infiltration directly over the base of a corn plant (cut off at the soil surface).

Short Distance Spatial Consideration for Temporal Variations

Short range spatial variations are typically due to cultural practices such as tillage, row spacing and wheel-traffic or intrinsic plant canopy effects such as surface protection and distance from plant (Ankeny et al., 1995; Prieksat et al., 1994; Freese et al., 1993; Starr, 1990, Starr and Paltineanu, 1996; Starr et al., 1995). Wherever short-range spatial dependencies occur, precise identification of the location of infiltration measurements is required. Long range spatial dependencies are largely due to soil formation factors or terrain and care must be taken to identify changes in three domains. Ankeny et al., (1995) found greater reduction of infiltration rates in traffic row vs nontraffic interrows of chisel plow tillage compared to no-till among five Midwestern locations. At two locations infiltration rates in untrafficked interrows under no-till were significantly lower than under chisel-plow system. Ankeny et al., (1995) also observed that ponded infiltration rates of the trafficked interrows were not only lower than other positions, but they were not different between tillage systems at any location. Data from no-till and plowed corn plots at in-row, traffic and nontraffic interrows locations, using real-time soil water dynamics with multi-level capacitance sensor probes (Paltineanu and Starr, 1996; Starr and Paltineanu, 1996) clearly show that water from rain or sprinkler irrigation penetrates faster and in larger quantities in the in-row position of no-till than in any other positions.

Some of the cause for the greater in-row infiltration rates that have been observed seems to be due to plant effects and not just planter-induced changes in soil physical properties within the row. For example, Prieksat et al., (1994) reports a generally increased infiltration rates over the growing season under chisel plow with rates over corn plants about twice those between plants. This is in contrast to the seasonal changes in infiltration rates under plow tillage reported by Starr (1990) in which the infiltration rates decreased from planting until mid-season, then increased abruptly in the fall of the year. The fall-season increase was attributed to increased worm activity resulting from rewetting of the soil profile.

Prieksat et al., (1994) also observed that infiltration rates measured in nontraffic interrows under chisel plowing remained fairly constant through the season while traffic interrows were consistently an order of magnitude lower than all other rates. Similar relationships between field positions occurred under no-till excepting the rates at each position were nearly constant over the growing season.

An important factor for positional effects on infiltration rates under no-till corn may also be due to the exposed corn stems/roots from the previous year. Paltineanu et al., (1995) observed that ponded infiltration rates over previous year's corn stems and brace roots under no-till increased ponded infiltration by ca. 10-fold.

These short distance variations can often be characterized by both "temporal condition" and the positions vulnerability to change. For example, the difference between a crop row and an interrow is often due to traffic changing the condition of the soil in the interrow while that within the row was not changed. The row position is also protected from change by the canopy but the soil is more susceptible to change. The compacted interrow becomes less vulnerable to further change because most of the change has already occurred. Wheel traffic on corn interrows increases the bulk density largely by eliminating large pore sequences, resulting in less spatial heterogeneity in infiltration rates.

Short and Long Time Scales for Temporal Changes

Increasing or decreasing infiltration rates may have short or long time dependencies ranging from minutes (e.g., soil crusting during rainstorm events, wheel compaction, plowing, cultivation-breaking up crusts, worm movement to the soil surface following certain rainfall or irrigation events), to days (e.g., reconsolidation following plowing, cracking due to drought;), to seasonal (e.g., changing crop canopies, root growth/decay, residue enhanced worm activity).

Measurement of Near Saturated Infiltration Characteristics

Direct Macropore Measurements

Owing to the impact of macropores on infiltration, a better understanding of soil systems can be derived by direct measurement of a variety of soil properties. Macropores can be characterized directly, but it remains difficult to identify which pores are hydrologically-active. For example, Constantz et al., (1988), found 5-10 fold reductions of ponded infiltration rates due to encapsulated air, suggesting that much of the encapsulated air resided in the interconnected pores (macropores) of the soil. For the time scale considered, this effectively nullified the effects of macropores in ponded/flooded infiltration rates.

Image analysis (Protz et al., 1987) can be done from resin-impregnated samples (Mackie-Dawson et al., 1988; Moran et al., 1989; McBratney and Moran, 1990; Singh et al., 1991; Mah et al., 1992), a soil peel (Smettem and Collis-George, 1985), a photo with proper contrast (Edwards et al., 1988), pore mapping (Logsdon et al., 1990), or a frozen soil thick section (Vermeul et al., 1993). Dyes (Smettem and Trudgill, 1983; Edwards et al., 1988; Logsdon et al., 1990) or plaster of paris (FitzPatrick et al., 1985) help to distinguish continuity. Tippkotter (1983) obtained a resin macropore cast of root channels which showed connectivity and root hairs.

Three-dimensional images are obtained from computer aided tomography (CT-scans) (Grevers et al., 1989; Tokunaga, 1988; Warner et al., 1989; Anderson et al., 1990), ground-penetrating radar (Kung and Donohue, 1991; Kung and Lu, 1993), or acoustical techniques. The CT-scans are obtained on samples, which are destructive, but all three techniques are expensive which precludes them from routine analysis. The ground- penetrating radar and acoustical techniques do not penetrate very far in fine- textured soils, and would thus miss important macropore information at soil depths. Direct pore data can help to understand the soil systems that effect infiltration rates, yet little has been achieved at predicting field-scale infiltration from such data.

Infiltration, Water Content, and Conductivity Measurement Methods

Disk or tension permeameters (Ankeny et al., 1988; Perroux and White, 1988) can be used to determine infiltration (or conductivity) as a function of negative head (Ankeny et al., 1991; Reynolds and Elrick, 1991; Cook, 1991; Logsdon and Jaynes, 1993; Hussen and Warrick, 1993; Cook and Broeben, 1994). They are usually used at the surface, but they can be used at subsurface depths if an area is cleared down to the depth of interest. Logsdon and Jaynes (1993) determine conductivity over the range of 0 to -150 mm head. As a first approximation, an exponential relationship was fitted to the data over this range, but ponded infiltration data had to be omitted to obtain a good fit. Preliminary data showed that conductivities measure from at ascending heads were significantly slower than conductivities measure from descending heads. Thus near saturated conductivities involve hysteresis.

In order to understand macropore influence on infiltration, a study of macropore volume changes would be helpful. Various methods are available to determine macropore volume and volumetric soil water content (Θ) as a function of negative head (Logsdon et al., 1993b). The rotated core method (McCoy, 1989; Logsdon et al., 1993b) is a direct water absorption/desorption method, which can be desorbed to -300 mm. Volume of air (i.e. macropore and mesopore volume, (Luxmoore, 1981)) can be determined from Θ at each applied head by subtraction. Traditional desorption of soil water in cores is inadequate in the range near saturation because of the gradient within the core during drainage. The rotated core method evens out this gradient because of the rotation during drainage. A laboratory measure of hydraulic conductivity (K(h)) over the range 0 to -150 mm was also used to determine K(Θ) by weighing the core after steady-state determination of K(Θ) (Logsdon et al., 1993b). Once Θ is known, then macropore volume can be determined by subtraction. Even field measurements of K(h) could be sampled afterward to determine Θ, but only at the last measured head at a location.

Modified borehole permeameter analysis can be used at soil depths to determine hydraulic properties with only a hole for access, but results are questionable for macroporous soils (Wu et al., 1992).

Timlin et al., 1994, assessed hydraulic characteristics of the soil by three field methods: tension infiltrometer (7.6-cm diam.), redistribution below a double-ring infiltrometer (50-cm diam.), and surface crust method in a 50-cm diam. inner ring. They found reduced rates of infiltration under confined one dimensional infiltrometer vs smaller unconfined infiltrometer. The lower rates were attributed to: entrapped air in the confined system; strong increase of bulk density with depth; and integrating over 0-20cm depth for one dimensional measurements. Advantages cited for the redistribution method were: 1) homogeneity with depth is not required; 2) macropore conductivity (and connectivity) can be determined for subsurface layers without soil disturbance; 3) the sampling area may better approximate the representative volume for a soil containing macropores.

A similar but simpler soil-water redistribution method has been proposed by Chen et al., 1993. This method assumes the soil to be a two-domain water flow system comprised of macropores which dominate the early drainage process, and the matrix pore space, which is responsible for drainage occurring after macropores are emptied. The unit hydraulic gradient approach for calculating K was extended and applied to the 2-domain system. In this approach, emphasis is placed on the hydraulic effectiveness of macropores in 2-domain flow systems rather than focusing on pore size and soil structure. The method requires only water content as a function of depth and time measured under free-drainage conditions from an initially saturated profile. Accuracy of results is highly dependent on the sensitivity and accuracy of the water content measurements.

Real-time water content data can be obtained with TDR probes inserted at several horizontal depths. This kind of data can also be obtained with the recent major improvement of the capacitance technique (EnviroSCAN® [4] by Sentek Pty Ltd, Kent Town, South Australia). Starr and Paltineanu (1996) have used this approach to measure changes in soil water content on a 10-minute interval at four depths (10 to 50 cm) in and between rows of corn under no-till and

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[4]Names are necessary to report factually on available data; however, the USDA neither guarantees nor warrants the standard of the product, and the use of the name by USDA implies no approval of the product to the exclusion of others that may be suitable.

plow. This methodology holds great promise for obtaining highly accurate *in situ* water contents data for obtaining the K values of macropores and the soil matrix that vary in time as well as in space.

Because of temporal variability discussed earlier, a one-time measurement is insufficient for characterizing infiltration rate. Measured temporal variability of infiltration rate does not always follow a pattern since it is influenced by many factors.

Subsurface Macropore Influence on Temporal Variability of Infiltration

Subsurface properties become important for infiltration processes when a subsurface impeding layer induced a perched water table. This is especially important when macropores are not continuous through the impeding layer. Future studies on temporal variability should include subsurface measurements. Direct infiltration measurement of temporal variability for subsurface hydraulic properties may not be possible because of the destructive nature of the measurements.

There is very little data on subsurface temporal variability of K_s. Cassel and Nelson (1985) observed temporal variation in K_s for three depths sampled at three dates for undisturbed cores of Norfolk loamy sand measured in the laboratory. Differences were not compared statistically, but K_s was generally lowest for the July measurement, and sometimes high for the June measurement with intermediate values for the May measurement.

A factor potentially influencing subsurface water flow properties is the continuity of pores through the topsoil into the subsoil. Tillage disrupts continuity of macropores (Logsdon et al., 1990; Roseberg and McCoy, 1992; Ehlers, 1975). Tillage creates unstable pores within the tillage layer itself and deeper tillage disrupts the natural, more stable macropores to a deeper depth, replacing them with unstable pores not connected with the natural macropores beneath the zone of tillage (Johnson and Erickson, 1991). In soils that do not contain natural macropores, slot tillage pores can be stabilized by the roots growing through the slots (Elkins et al., 1983). Blackwell et al. (1990) examined the stability of old root channels under traffic and observed that the macropores were more stable if they were larger, more vertically oriented, or deeper in the profile. If subsurface macropores or cracks are present a field should not be deep tilled because deep tillage would disrupt the existing stable pores and create unstable pores.

Because of the difficulty in obtaining repeated direct subsurface infiltration measurements over time, indirect measurements such as the instantaneous profile method (Watson, 1966; Green et al., 1986) could provide temporal information. Although this could be done in a field, it would be time consuming and tie up an area of land out of crop production (because crop water uptake interferes with the vertical gradient). Instead of conducting a temporal study in the field, large undisturbed columns could obtained from different tillage systems at different times. These columns could be instrumented with automated recording devices for soil water (using time domain reflectometry TDR and tensiometers. The columns could be used for drainage and /or evaporation instantaneous profile measurements. These columns, in addition to the instantaneous profile measurements could also be used for negative head and ponded hydraulic conductivity measurements using disk permeameters.

Current Status of Knowledge Base And Priorities For Research

Infiltration, the process by which water enters the soil, has been studied with various degrees of intensity since the publication of the algebraic infiltration equation by Green and Ampt in 1911. Yet, we are unable to predict with reasonable certainty the rate that water will infiltrate the soil, nor the subsequent fate of the water in the soil. Because of the magnitude of the temporal and spatial variability of infiltration, reliable data are difficult and time consuming to collect. Further, many of the measurement techniques alter the soil and affect of the process being studied. Most techniques impose conditions substantially different from that which occurs in nature. Hence, empirical equations are used more and more in computer simulations of infiltration, instead of using actual experimental data for the soil hydraulic functions (Youngs, 1995).

Difficulties in modeling infiltration of natural field soils is caused primarily by our lack of knowledge of the dynamic nature of soil: 1) infiltration increases through macropores from earthworm activity, brace root growth, root decay, desiccation cracks, and temporarily by tillage; 2) infiltration decreases from reconsolidation and surface seal formation, which is influenced by rain energy, residue and canopy cover, soil erodibility, tillage, and aggregate stability (Le Bissonnais et al., 1989; Loch, 1994; Reichert and Norton, 1994). Blackwell et al. (1990) have described cylindrical macropore destruction by traffic, but there is little, if any information on destruction of cracks by traffic. Shrink- swell influence on macropore infiltration has been incorporated into water flow and solute transport models (Messing and Jarvis, 1990; Jarvis and Leeds-Harrison, 1990; Jarvis, 1991; Kim and Chung, 1994). Traffic and tillage operations are seasonal; perhaps simple seasonal equations could describe these factors. Perhaps starting with a decision aids model (presence or absence of given factors), we could give a first cut quantitative estimate of how the macropore system varies with time. As additional temporal data is collected (and associated soil/climate/crop factors described), this approach could be refined.

This review has noted the increasing recognition that water infiltration rates are far from static in either space or time. To significantly increase our ability to predict infiltration of natural soil systems much more field infiltration data is needed and it must be collected in such a way that it adds to our global knowledge of the process rather than just becoming another interesting observation. It is apparent that one data point in the year is of little value unless it can be described by known spatial and time patterns. Further, to be of significant value, research reports must include field-crop positions, and infiltration conditions such as the presence of encapsulated air, surface seals, residues, crop canopy, soil profile description, etc. Finally, sufficient data need to be collected over space and time to characterize major soil systems. At first consideration, this data seems impossible to obtain, especially under shrinking budget conditions. Clearly, simplified techniques need to be developed and utilized, such that the primary assumptions of the methodology can be met within the limited resources available. A combination of the soil-water redistribution method of Chen et al.(1993), and the multisensor capacitance probes for measuring soil water content (Starr and Paltineanu, 1996) seems to overcome many of the short-comings of previous techniques. All field methods have limitations, including difficulty in obtaining adequate numbers of observations to statistically characterize a field at a point in time. Clearly, there is not only room for, but an urgent need for development of new methods. To better understand and to predict how macropores influence the temporal variability of infiltration, new and intensive research efforts should take place in the following three areas: modeling, methods, and data collection.

Improved predictive models should be developed which including: a) short-term temporally varying infiltration rates due to reconsolidation and surface sealing; b) seasonally varying infiltration rates due to plant growth earthworm activity; and c) systematic small-scale spatial variability such as crop-row position basal plant coverage and other effects.

Simplified infiltration methods should be developed that: a) mimics the primary water application at the site (rain/sprinkler, furrow/ponded); b) do not require simplifying assumptions that are difficult to obtain or maintain in the field; and c) may be conducted at a large enough scale to provide statistically meaningful results.

Sufficient data over space and time should be collected to characterize major soils and soil-cropping systems including the temporal changes which can predominate the source of variation. Since management and time often cause orders of magnitude variation in infiltration, data must include all relevant information necessary to describe the system and measurements. This database should be formatted and presented in such a way as to generalize the results so that in the future they can be applied to unknown, but described (soil type, crop, etc.), conditions.

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Temporal Character of Surface Seal/Crust: Influences of Tillage and Crop Residue

R R. Allmaras[1], J.L. Pikul Jr.[2], C.L. Douglas Jr.[3], and R.W. Rickman[3]

Abstract

Infiltration is fundamental to soil and water conservation and is sensitive to soil management. The temporal character of surface seal formation and reduced infiltration may well be sensitive to tillage and residue management because decomposition products can control the formation of water stable aggregates. The shift from moldboard plowing to other forms of primary tillage on 94% of the wheat, corn and soybean production in United States places the crop residue within 10 cm of the soil surface, where the decomposition products can most effectively modify infiltration. A cause and effect linkage (crop residues→aggregating agents→soil aggregation→seal formation→infiltration) was proposed for systematically evaluating the temporal character of surface seal formation and infiltration. Tillage control on crop residue placement and hydrothermal control on decomposition is generic and useful on a microsite/field/regional scale. Much is known about organic matter-mineral soil interaction in sieve sizes ranging from 2 to 2000μ. Wet aggregate stability is useful to predict K_{sat} in the simple saturated-flow model of the surface seal, but a uniform procedure (including field sampling, wetting, sieving, and calculated stability index) is needed and is suggested. The field procedure must also have supporting information on porosity distribution. This linkage scheme should be useful for resource assessment as well as site specific prediction.

Need for Study

Spatial and temporal variations of infiltration are commonplace in crop production fields and are often difficult to distinguish (Starr et al., 1996). Spatial variation can be a comparison of several different temporal variations of infiltration. For instance, Ankeny et al. (1990, 1991) tested infiltration in interrows with and without traffic during corn planting in different tillage systems. The final infiltration rates with and without traffic were different, but each position could have been at a different stage in the temporal transition to a seal-controlled infiltration. Crop residue management and the tillage system could have had a major impact on these temporal transitions if there were surface crusts/seals involved. Logsdon et al. (1993) concluded that temporal infiltration effects were greater than tillage or rotation effects. Surface seal formation and later disintegration due to shrink-swell and biological activity when dried were noted to produce the dominating temporal character. They discussed other sets of comparisons among tillages that showed significant temporal variation.

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[1]USDA-ARS, 439 Borlaug Hall, University of Minnesota, St. Paul, Minnesota 55108

[2]USDA-ARS, Northern Grain Insects Laboratory, Brookings, South Dakota 57007

[3]USDA-ARS, CPCRC, P.O. Box 370, Pendleton, Oregon 97801

Biologically controlled reactions are undoubtedly a major factor in the soil structure stability needed for prevention of surface seals/crusts. Biological components involved in these reactions are crop residues, rooting activity and debris, fauna and microbial flora, and soil organic matter. Resistance may be expressed by a slower rate of seal development but also a higher final infiltration rate due to a higher minimum saturated hydraulic conductivity in the seal. These reactions are highly sensitive to physical/chemical processes in the soil, especially those near the soil surface.

Infiltration studies in situ may compare treatments at a time not necessarily relevant to the most influential biological activity (as discussed by Logsdon et al., 1993). Other studies of infiltration have been made on disturbed soil, in which the biological component could easily have been perturbed out of the system. For instance, many tension infiltrometer measurements at the soil surface require soil manipulation or added sand for better contact, an experimental procedure which could easily destroy biological agents and the surface seal involved in control of infiltration. Morphology of the crust/seal may retain juxtaposition of inorganic components and soil structural features but the organic components are often lost during sample processing.

Infiltration and water storage benefits may have been changed drastically due to recent changes in tillage systems used by American farmers. Reduced tillage systems with and without 30% surface cover with residue at planting has replaced moldboard plow tillage (ERS, 1994; Allmaras et al., 1996b) on about 94% of land tilled for soybean, corn, and wheat. Instead of deeply buried crop residue, the crop residue is permanently retained on or within 10 cm of the soil surface. Undoubtedly there have been gains of technical efficiency and economy. A structured analytical system is needed to identify when and how biological factors in the crop residues can be expected to change infiltration during a season of arable agriculture. Because of our knowledge about crop production and tillage systems on an areal basis we ought to be able to assess nationally the resources that change soil and water conservation based upon management induced changes in infiltration.

Structured Relation: Crop Residue Control on Infiltration

The hypothesized structured relation between crop residue management and infiltration operates during the interim between tillage disturbance and steady state infiltration through a surface seal/crust. The bolded arrows (Fig. 1) highlight the desired cause and effect relations upon which a soil management scheme can be constructed using projected processes and estimable parameters. Some literature is available but additional studies are needed. The light lines indicate three of the most common linkages which are highly empirical because one or more of the underlying processes were not included in the measurements.

Supply and Position of Crop Residue

Recent interests in soil quality have focused on plant residue remaining after harvest. In most research and farm operations a measured harvest yield can be used with a harvest index (harvest yield÷total aboveground biomass) to project the amount of crop residue available for production of soil aggregating substances (Fig. 1). Prihar and Stewart (1990, 1991) discuss the theoretical basis of harvest index and give suggested values for various crop species; Allmaras et al. (1996b) review the use of harvest index for projecting the supply of biomass available for decomposition in the soil. Beauchamp and Voroney (1994) suggest that root material is a conservative 20% of the total shoot biomass and must be included in the supply of crop residue. This reservoir of plant residue does not include the exudates and rhizodeposition during active plant growth and microbial reactions coincident with root growth.

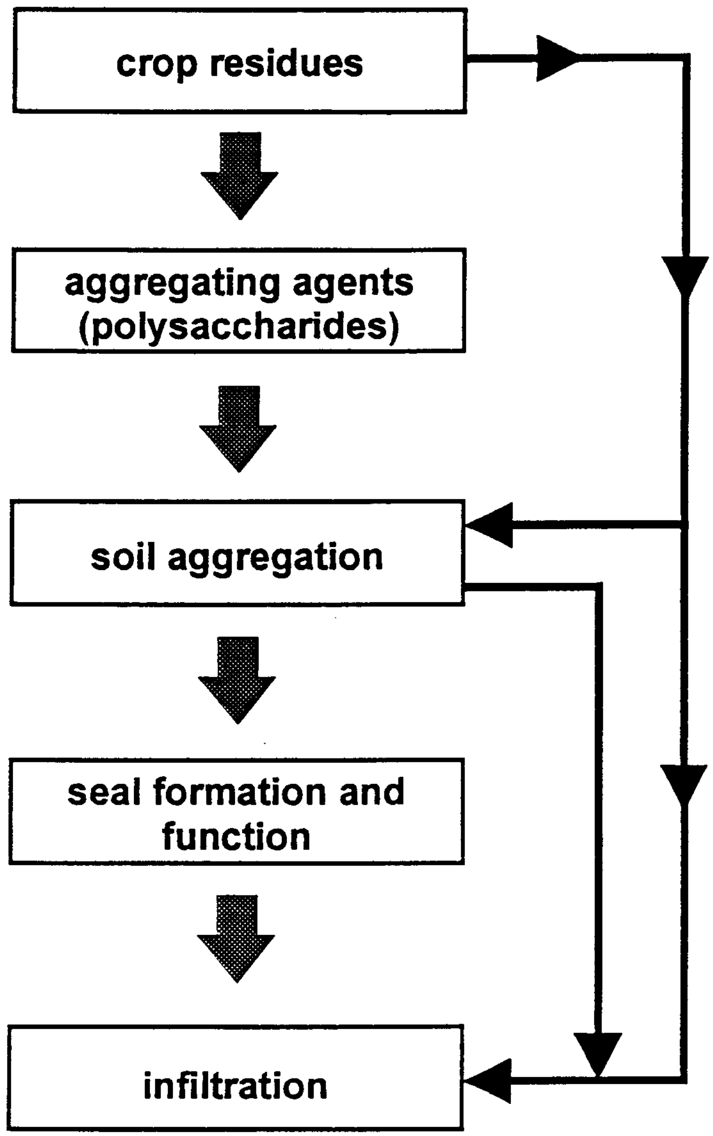

Fig. 1. Informational linkage for cause and effect between crop residue and infiltration through a surface seal/crust (adopted with revision from Boyle et al., 1989)

Not only is quantity of crop residue important, but also the position in the soil or on the surface must be specified as controlled by the tillage system. In an annual cropping system with spring sown crops, the primary tillage (if performed) usually occurs in the fall (September to November) and several secondary tillages (March through April) precede planting. There may or may not be postplant cultivation. Whenever moldboard plowed, the fresh or current residue is buried below 15 cm and old residue (after one year of decomposition) is deposited above 15 cm (Staricka et al., 1991, 1992; Allmaras et al., 1996a). When disc, sweep, or chisel are used for primary tillage, all buried residue is above 15 cm and often near the soil surface. The depth distribution within the top 15 cm shows sweep better than chisel better than disk for surface cover--this applies even when these tools are used in secondary tillage (Allmaras et al., 1988; Staricka et al., 1991). Fall sown crops show qualitative relations similar to spring sown crops and the positioning of the crop residue can be developed from a history of tillage operations. The control on buried residue positioning with tillage tools is equally specific to the control of surface residue cover now used in RUSLE (Renard et al., 1991). Almost complete farmer adoption of systems to keep residue in the top 15 cm of soil (ERS, 1994) could be a management change to use crop residue for better infiltration management. Because primary tillage with a moldboard inverts soil and included organic constituents about a 10 to 15-cm depth and other forms of primary tillage do not move soil and organic constituents out of the top 15 cm, a precise estimate of residue positioning would require knowledge of tillage system used after the harvest prior to the one that is producing the residue under test.

Residue Decomposition to Produce Aggregating Agents

The decomposition of crop residues controls the production of aggregating agents. Placement of the residue under control of tillage was discussed in an earlier section. Field investigations with mature crop residues show about 10 to 30% weight loss in the fall after harvest, an initiation of weight loss in early spring about when mean daily soil temperatures are above freezing, and about 20 to 40% of the original weight remaining at the end of the growing season (Douglas et al., 1980; Smith and Peckenpaugh, 1988; Ghidey and Alberts, 1993; Schomberg et al., 1994). Other field measures of residue weight loss are : Brown and Dickey (1970), Smith and Douglas (1971), Tanaka (1986), Collins et al. (1990), and Stott et al. (1990). Most test measurements were made on cereal residues but measurements were made on other crop residues, including corn and soybean. The glass cloth bag is the most used procedure, yet comparisons with other methods usually provide similar rates of weight loss. Measurements by Staricka et al. (1991) reveal a highly clustered distribution of buried crop residue not unlike that in the recoverable bags. Weight loss measurements over a 2-year period showed the rate of loss to be exceedingly slow after the end of the first season (Broder and Wagner, 1988; Ghidey and Alberts, 1993). It is presumed, that after a full year, the release of aggregate binding compounds would be unimportant, especially after a new sequence of tillage before the next growing season.

There are numerous models to predict the course of crop residue decomposition as related to placement (mainly above the surface, on the surface, and buried nearly always below 5 cm) and hydrothermal environment in the soil and soil-atmosphere interface (Gregory et al., 1985; Stroo et al., 1989; Stott et al., 1990; Douglas and Rickman, 1992; Steiner et al., 1994). Some of these models specify an initial C/N ratio, others used a N function (Douglas and Rickman, 1992) to improve the predicted decomposition rate. Broder and Wagner (1988) showed that soluble components of the crop residue hastened decomposition. Soil texture must also be considered because the storage of C in humic compounds increases as clay content increases (Martin and Haider, 1986; Hassink, 1995). Also available are microbial activity models for mineralization, which require simultaneous moisture and thermal optimums (Linn and Doran, 1984, a,b). Residue loading does not change the rate of weight loss within normal rates of crop residue return (Jenkinson, 1977), although the amount of soil aggregating agents produced will be proportional to the loading.

During residue decomposition, an array of soil aggregating agents are produced mainly as a result of microbial activity (Boyle et al., 1989). One of the most important of these aggregating agents is the polysaccharide. Weight loss techniques in the field are likely to overestimate decomposition because some biomass (rich in polysaccharides) may remain with the residue and be separated during the washing to remove soil contaminants; the new humic products are more likely to be retained in the crop residue because they are less water soluble. Laboratory studies with C^{14} label of residue (Martin and Haider, 1986) verify that CO_2 evolution accounts for about 60 to 70% of the original residue after 1 year, and that microbial biomass may account for as much as 10% of the remaining residue. Lignin content of residue after field exposure is needed to calibrate the yield of aggregating agents, because CO_2 loss cannot be conveniently measured during exposure.

Most mature crop residue contains the following materials: solubles 200g kg^{-1}, hemicelluloses 200g kg^{-1}, celluloses 300g kg^{-1}, lignins 200g kg^{-1}, proteins and inorganics 100g kg^{-1} (Oades, 1989). Aside from proteins and inorganics in the mature residue, the decomposition rate is inverse to the order mentioned. So the solubles C fraction is usually lost soon after harvest, while the lignin component remains within the 20% or more of C remaining in crop residue at the end of the growing season. Lignins may lose 15% of their C during decomposition, but the residual carbon is not present in biomass associated with the microbial proteins and polysaccharides (Martin and Haider, 1986). Hemicelluloses and celluloses constitute about 50% of the plant derived C and comprise the substrate most useful for microbial production of polysaccharides.

Soil Aggregating Agents and Soil Aggregation

Literature in soil and biochemical sciences is profuse with chemical descriptions of the various compounds in soil organic matter--some compounds (such as sugars and amino acids) are transitory, others (such as lignin and humic acids) are resistant or recalcitrant, and in between are the fulvic acids and polysaccharides (Greenland and Oades, 1975; Martin and Haider, 1986; Boyle et al., 1989; Oades, 1989). Polysaccharides are the organic constituents most active in soil aggregation; there are many molecular variations; and they generally make up 5 to 20% of soil organic matter. Five functional forms were outlined based upon resistance to chemical, physical or microbial degradation (Boyle et al., 1989). Evidence for polysaccharide involvement in soil aggregation has been demonstrated by: a) aggregation upon addition of polysaccharides to soils of poor structure, b) destruction of natural aggregates when treated with periodate or tetraborate, and c) measured adhesive properties of polysaccharides related to length and linear structure of the molecule (Boyle et al., 1989). Methods for analysis of polysaccharides in soil are an anthrone reactive carbon (DeLuca and Keeney, 1994), and a phenol-sulfuric hydrolysis of polysaccharide polymers (Lowe, 1993).

Morphology of the Water Stable Aggregate

A rationale for the water stable aggregate has a basis in the many fractionation studies of soil organic matter. Different binding processes function for different size classes of water stable aggregates (Tisdall and Oades, 1982; Oades, 1984) when organic matter is the binding agent. Stages of aggregation ($< 0.2\mu m \rightarrow 0.2$ to $2\mu m \rightarrow 2$ to $20\mu m \rightarrow 20$ to $250\mu m \rightarrow > 2$ mm) were suggested based upon thin section, chemical tests, and input of rupture energy. Hyphae and roots were the agents for binding 20 to 250μm into the >2 mm aggregate. The 20 to 250μm aggregates resisted rapid wetting, were not easily destroyed by tillage, had clay intimately associated with the organic matter, and consisted mainly of 2 to 20μm aggregates. The 2 to 20μm aggregates or particles consisted of < 2μm particles, contained large amounts of clay and organic carbon, resisted ultra sonic vibration, and were highly water stable. Thus the concept of macroaggregates (>250μm), microaggregates (<250μm), and the sieve size of 0.250 mm to separate markedly water stable material from material that is dependent on soil management for water stability (Oades, 1984).

Macroaggregates are sustained by the binding activity of root and fungal hyphae, and their slaking into microaggregates is likely when organic matter declines. Polysaccharides are normally associated with binding in aggregates <50μm, but, when decomposing residues dominate over root and fungal activity, polysaccharides may play a role in macroaggregate binding. Tisdall and Oades (1982) already suggested that rapid wetting ought to be a part of water stable aggregate procedures because of the slaking of dry soil at or near the soil surface to produce a surface seal. They also linked pore diameters to the aggregation stages of the water stable aggregate. The 20 to 250μm aggregate contains the 0.2 to 2.5μm diameter pores for available water retention; the 0.25 to 1.0 mm aggregates contain the 25 to 100μm pores important to capillarity and aeration, and the aggregates > 1.0 mm relate to the pores >0.1 mm needed for root growth, aeration, and normal infiltration. An increase in dispersed clay results from partial disintegration of the microaggregates formed during slaking of the macroaggregates (Oades, 1984)--later it will be shown that dispersed clay is indeed correlated with decreasing organic matter.

These earlier morphological descriptions of water stable aggregates are somewhat supported by the characterization of organic matter in light and heavy (density) fractions in a soil (Golchin et al., 1994, 1995; Hassink, 1995; Gregorich and Janzen, 1996). From the light fraction, Golchin et al. (1994) separated the free and particulate organic matter (mainly plant debris) and the occluded particulate organic matter; organic matter in the heavy fraction was definitely of microbial origin and was colloidally associated with clay. These separations indicated a major role for carbohydrate-rich organic matter in the formation and stabilization of the 50 to 250μm microaggregates. Golchin et al. (1995) found this 50 to 250 μm fraction to be highly correlated with water stable aggregation, but the free particulate fraction showed no correlation. The whole light fraction is from recent plant origin (Gregorich and Janzen, 1996), therefore there might be re-evaluation of the impact of the microaggregate fraction on water stable aggregation. Should a sieve smaller that 0.25 mm be used and should more emphasis be placed on dispersed clay? Again more evidence for the role of young and active rather that total soil organic matter in stabilizing soil structure (Golchin et al., 1994; Hassink, 1995). A long history of cereal chaff compared to alfalfa production enriched the free particulate fraction, but the enrichment in the occluded particulate fraction was less than with alfalfa. A long-time farmyard manure enriched the heavy or colloidal fraction relatively more than the other long-term treatments of cereal chaff or alfalfa production (Hassink, 1995).

Morphology and Processes within the Surface Seal/Crust

Morphology and processes (mostly physical and chemical) within the surface seal/crust have been reviewed in detail by Bradford and Huang (1992) and West et al. (1992). Respectively reviewed were mechanisms of crust formation and morphology of crusts. A structural crust can have microlayers, all with a characteristic porosity caused by rain-induced aggregate breakdown, particle/aggregate sorting, aggregate coalescence, micromass depleted skeleton grains, and micromass collection (West et al., 1992). Under field conditions a crust develops stagewise (West et al., 1992); a first stage is the local development of a disruptional layer (distinguished by a more dense layer less than 5 mm thick at the surface). As more rainfall (intensity and cumulative kinetic energy) occurs, particle or aggregate displacement from the microhighs may thicken the disruptional layer initiated in the microlows. Sedimentary (multiple layer of disruptional crusts) crusts are developed, and wash-in may deposit micromass below the original disruptional layer. After random or other roughness has no more influence spatially, a seal is expected to occur, in which the originally dispersed clay may be concentrated. So the final outcome will be a disruptional crust and a possible subadjacent layer formed by wash-in of micromass. This layer may be up to 10 mm thick unless there are many sedimentary layers.

Ewing and Gupta (1994) used a pore-level model to evaluate compaction and filtration (washing in) mechanisms on soil surface seal. Compaction reduced surface seal porosity more than filtration, but filtration reduced the final K_{sat} of the seal much more than compaction. Most particles available for filtration remained on the surface suggesting that surface deposition may be the most important mechanism for determining the final hydraulic conductivity in the surface seal. These predicted formation conditions may provide flow-related consequences of the morphology changes described by West et al. (1992). The water stable aggregate could play a role by delaying development of the disruptional layer and maintaining a larger pore size. Might it also reduce the particles available for filtration and surface deposition in the pore-level model of Ewing and Gupta (1994)?

Soil texture, antecedent moisture, aggregate stability, and surface roughness often control surface-crust formation (Bradford and Huang, 1992). Aggregate stability is often controlled by biological inputs. Soil textures with high silt content showed the greatest decrease in infiltration and the largest increase in strength of the crust. Gupta and Larson (1979) demonstrated that a particle packing index best represents the textural influence on surface seal and runoff production. When the soil surface consists of initially dry aggregates, breakdown is mainly a slaking process with partially slaked fragments and particles filling the interaggregate pores. This process rapidly decreases infiltration, and hence the importance of rainfall intensity and aggregate stability (under both dry and wet exposure). Sustained high infiltration and lower crust strength were most different between initially wet and dry aggregates when the aggregate stability was greater. Although aggregation is important for erosion control, there are few quantified relations between aggregate stability (water stable) and crust formation. However, Bradford and Huang (1992) report a good negative relation between a stability index (fraction of 2 to 8-mm aggregates retained on 250μm screen) and change in strength of the crust measured with a fall cone.

The structural crust, including the disruptional layer and adjacent layers, is conveniently formed and studied in the laboratory, where Huang and Bradford (1993) demonstrated the use of microtopographic roughness measurements on a scale to distinguish simultaneous surface sealing, erosion, and deposition all associated with aggregate breakdown, filling of surface voids and depressions, and erosion of a structural crust. Such a procedure might be tested on undisturbed surfaces containing different residue management treatments to mimic the typical sequences in a field with conservation tillage. Field scale of roughness may localize different stages of crust formation. Linden et al. (1988) demonstrated that random roughness not only retards crust formation but also reduces the crusted (structural and/or sedimentary crust) area; similar conclusions were made in the field study of Falayi and Bouma (1975). Resistance of surface roughness to rainfall energy is most important for delaying crust formation during which there may be a significant amount of infiltration, but, when erosion on sloping land is simultaneous with crust formation these processes and relations may change significantly as suggested by Huang and Bradford (1993) and Gimenez et al. (1992).

When aggregate beds (aggregates <5 mm diameter) of "stable" and "unstable" soils were subjected to alternating simulated rainfall and drying, the unstable soils showed a structure deterioration under the crust (Magunda, 1992). Final infiltration rates of the stable were >10 mm hr^{-1} compared to <10 mm hr^{-1} for unstable soils.

Water Flow in the Crusted Soil

Ahuja and Swartzendruber (1992) and Gupta et al. (1992) have discussed a one-dimensional, saturated water flow in crusted soils. This approach is logical in general because formation of the disruptional crust (West et al., 1992) is the beginning of the decline in infiltration after tillage; it is a convenient starting position for incorporation of the biological component into a physical model. A common form of the equation is to deal only with change of the saturated hydraulic conductivity, K_s (t), in the thin crust:

$$K_s(t) = K_{sf} + (K_{so} - K_{sf}) \exp(SE) \quad (1)$$

where K_{sf} and K_{so} are the final and initial saturated hydraulic conductivity, E is the cumulative combined intensity and kinetic energy of rainfall, and S is a parameter related to soil resistance to inputs of rainfall energy.

Modeling of saturated flow in the crust appears to accept a uniform value of K_{sf} (van Doren and Allmaras, 1978) but there is undoubtedly a textural control, while Rőmkens et al. (1990) indicate that K_{sf} may also change due to rainfall intensity. Gimenez et al. (1992) indicated that final hydraulic conductance was sensitive to rainfall intensity; these hydraulic conductances were related to seal formation and exposure of planar voids under the seal. Van Doren and Allmaras (1978) noted S variations related to tillage but these S variations did not measure residue decay effects near the surface as controlled by primary tillage--the field data all had been moldboard plowed since the last harvest. The morphology of a water-stable aggregate suggests a direct effect of their stability on the limiting value of K_{sf} but research is needed to evaluate simultaneously the chemical, physical, and biological factors.

Measured saturated hydraulic conductivities (K_{sf}) through crusts of different soil materials (Sharma et al., 1981; Moore, 1981; Chiang et al., 1993 a,b) were roughly about 5% of the initial values (K_{so}) of Eq. (1). These crusts ranged from 0.5 to 1 cm thick based upon changes of pressure gradient under the crust. Associated S values (cm^2/J) in Eq. (1) were computed to test the influence of soil properties on soil stability against slaking. Among 12 soils tested, Sharma et al. (1981) could not show a textural effect, but significant cultural contrasts implied an aggregation effect to reduce the magnitude of S and to increase K_{sf}. Chiang et al. (1993 a) demonstrated a higher value of S for southeastern Ultisols compared to Alfisols and Mollisols of the midwest, but Chiang et al. (1993 b) could not show an effect of water stable aggregates on the magnitude of S in southeastern soils. They did show that water dispersible clay was positively correlated with S.

Water Stability of Aggregates and Possible K_{sat} Relation

Methodology

A measure of water stable aggregates has long been tested for predicting the relative resistance to surface seal. Yoder (1936) detailed a method based upon a known or reproducible energy input and used five sieve sizes to compute some mean diameter. Kemper and Rosenau (1986) reviewed procedures and recommended a single sieve (0.25 mm) to separate stable from destabilized aggregates and merely state percent water stable. The amount of energy and the initial water content (oven dry, field moist, or a moist preconditioning exposure) should mimic the expected field exposure conditions. To mimic slaking during soil crusting, Tisdall and Oades (1982) suggested immersion of air dry aggregates. Irrespective of procedures to mimic initial condition and slaking mechanisms, a quantitative relation between water stable aggregates and K_{sf} and/or S in Eq. (1) requires measurements of water stable aggregation using multiple sieves and a measure of the soil remaining on each sieve.

Since 1986, there have been significant studies on methods. Pojasok and Kay (1990) developed a combination wet sieving and turbidimetry to estimate water stable aggregates and dispersible clay. Instead of a vertical stroke of the sieve nest the suspension of aggregates is shaken mildly end-over-end and then sieved. Measured water stable aggregates (percent retained on a 0.25-mm sieve) in the modified method agreed with the Yoder (1936) method, but the modified method was more sensitive to initial soil moisture in the aggregates. The water-stable aggregate methods and the dispersible clay were all three sensitive to grass vs arable management. An extension of the Pojasok and Kay (1990) procedure used water stable aggregates and dispersible clay both as a function of aggregate water content when sampled to predict the aggregate water content at maximum dispersible clay (Rasiah, 1994). Dispersible clay was a maximum in the wet range and increased as clay content increased and organic matter decreased. Wet aggregate stability (WAS) decreased as water content at sampling increased, and this sensitivity of WAS increased as organic matter and clay increased (Rasiah et al., 1992).

The rate of wetting aggregates in the modified Pojasok and Kay (1990) method provided information about an intrinsic wetting rate and the associated stability against slaking (Rasiah and Kay, 1995 a). Intrinsic wetting rates were controlled by organic matter content, clay, and cropping treatment (corn, forage crop, or grass); stability in water decreased with increased wetting rate; and the impact of wetting rate on stability was more intensive in larger aggregates. Air dried aggregates 2 to 4 mm were wetted for different time lengths up to 9 minutes before measuring water stable aggregates on 2, 1, 0.5, and 0.25 mm screens. This technique might be used to estimate S in Eq. (1).

Related water stability (or wet aggregate stability) and dispersed clay measurements have been made using the same field treatments of tillage, corn, forage, grass (Rasiah et al., 1992, 1993; Rasiah and Kay, 1994; Rasiah et al., 1995; Rasiah and Biederbeck, 1996). Aggregates sampled over the growing season had decreased wet stability and increased dispersed clay with increasing water content. Dispersed clay increased as clay increased and organic matter increased, but wet aggregate stability was a function of clay ÷ organic matter (Rasiah et al., 1992; Rasiah and Kay, 1994). When wet aggregate stability was measured repeatedly over 3 growing seasons and there was a change in management, the change of wet aggregate stability was estimated to have a half life of 5 years in a clay loam but 8 years in a sandy loam (Rasiah and Kay, 1994).

A fractal dimension of mass based wet-aggregate classes was developed and applied to a whole series of aggregate samples from field experiments with tillage and forage treatments in soils of variable texture and organic matter (Rasiah et al., 1993; Rasiah et al., 1995; Rasiah and Biederbeck, 1995). These fractals should provide a step in the estimation of K_{sf} and/or S in Eq (1), but estimates of large pore volume, aggregate density, and packing density will be needed along with the fractal of the mass aggregate distribution to estimate K_{sf} in Eq. (1) (see Giminez, 1995).

Zuzel et al. (1990) and Pikul and Zuzel (1994) developed an interesting set of data consisting of infiltration and pore size distribution (using mercury porosimetry) to estimate K_{sf} and/or S in Eq. (1). These measurements have pore size distributions and other measurements as related to long-term tillage, residue input, and fertilization; pore size distributions were made at various stages of crust formation.

Miscellaneous Water Stability Measurements

Numerous studies have followed the path (Fig. 1) from soil aggregation to infiltration with or without information on the seal/crust. Numerous of these studies have chosen to measure only the wet aggregates retained on a 0.25-mm screen or have computed the mean weight diameter when using more than one screen. Some have linked wet aggregate stability to aggregate binding agents.

Water stable aggregation (percent retained on 0.25-mm screen), infiltration, soil carbon, and other measurements demonstrated the value of a winter cover crop vs fallow in rotations with grain sorghum and soybean summer crops in an warm-humid climate in which crop residue from the warm season growth does not control crusting and soil erosion in the cold season even when there is only shallow or no post season tillage (Bruce et al., 1992, 1995).

Runoff, soil loss, and time-to-ponding were predicted by water stable aggregation (percent of 1 to 2 mm aggregates retained on 0.25 mm screen), dispersible clay; these relations were derived in simulated rainfall to produce soil detachment on plots with various tillage and crop rotation treatments (Rasiah and Kay, 1995 b). Wet aggregate stability ranged from 16 to 45% and an absolute 5% increase of wet aggregate stability increased the time to ponding by 27% (an increase in infiltration); only dispersible clay exerted a significant influence on runoff and sediment delivered.

Wet aggregate stability has been linked to aggregating agents to explain the influence of soil organic constituents (Haynes and Swift, 1990; Haynes et al., 1991). Wet aggregate stability increased (mean weight diameter obtained from aggregates on 2, 1, and 0.5 mm sieve) as acid hydrolyzable and cold water extractable carbohydrate increased in response to pasture vs arable culture on numerous soils in New Zealand. Wet aggregate stability increased with less years of arable cropping or more years of pasture. The water stable aggregates (<2.0 mm) contained more organic carbon and acid hydrolyzable carbohydrate than the unstable (<0.25) aggregates in all three types of management: arable, pasture and regrassed. Clod porosity increased as aggregate stability increased--at equal stability the arable clods had higher porosity. Gregorich et al. (1994) reviewed indices of changes in soil organic matter related to tillage and residue management associated with long term tillage treatments. Wet aggregate stability, hot water extractable carbohydrate, and acid actractable carbohydrate were shown to respond to residue positioning controlled by tillage systems in continuous barley--these changes occurred before any changes in total carbon could be detected.

Roberson et al. (1991) observed changes in polysaccharide mediated macroaggregation stable in water. Various forms of soil management (mowed barley, herbicide controlled vegetation, conventional tillage, permanent grass) changed crust vs no crust K_{sat}, polysaccharide in the heavy fraction, somewhat the mean weight diameter of water stable aggregates and strongly the slaking resistance. There were strong correlations between K_{sat}, slaking resistance, and heavy fraction carbohydrate, but light fraction carbohydrate was not well correlated with other measurements.

Concluding Remarks and Research Needs

American farmers are now using tillage systems that retain crop residues on the surface or in the upper 15 cm of the soil. A 1993 survey by the Economic Research Service (ERS, 1994) indicates that these tillage systems are being used on 95% of the planted corn, soybeans, wheat, and sorghum. This change is an abrupt disuse of the moldboard plow since 1980. Instead of fresh-crop-residue burial below 15 cm, all residues are available for soil structural improvements at the soil surface, where infiltration can be markedly changed and controlled by residue decomposition. Research needs have shifted and place a greater emphasis on the soil organic matter-soil structure-infiltration continuum both to search out opportunistic soil management and to improve natural resource assessments.

1. A much more comprehensive review is needed with some sensitivity projections to determine where pilot studies may be initiated in the cause and effect linkage of crop residue → soil aggregating agents → soil aggregation indicators → seal formation and function → infiltration. Some of these linkage elements have been researched but little attention has been given to the linkage elements simultaneously.

2. Field measurements of decomposition of crop residues needs to be revisited for checking out of the predictions of decomposition as related to hydrothermal environment, C/N ratio, and lignin content of residues before their field exposure and after various stages of their decomposition. Residue position as related to tillage has been worked out and perhaps is firmly implanted into RUSLE. The original objective of measured residue decomposition was residue available for erosion, but here the objective is determining the time when the aggregating agents are released from the residue.

3. Literature about aggregating agents is dominantly qualitative from laboratory observations. However, qualitative and quantitative characterization of aggregating compounds (resulting from crop residue decomposition) is expensive and perhaps cannot be routine. Rather the emphasis should be aggregation and related water stability. There is much research with organic-mineral interactions related to particle or sieve size, research that may support decisions about water stability and the status of organic materials in soil.

4. Although field accounts of surface seal formation and change suggests a multilayered surface seal, a simple-seal model with a thin disturbed (from rainfall energy) layer overlying the undisturbed soil (recently tilled or untilled) may sufficiently describe water flow processes during incipient crust/seal formation and decline of infiltration. This model is a one-dimensional saturated flow model. Parameters in the model relate to K_{sat} and the resistance to rainfall input of energy. There is currently much research activity to relate K_{sat} to soil structure and macroporosity.

5. Wet aggregate stability is the commonly measured soil characteristic responsive to organic matter interactions with soil structure. However, methods to measure water stable aggregation must be standardized to avoid cumbersome techniques and marginally useful reported results. Field sampling, method of wetting (without agitation in a nest of sieves), method of sieve separation (5 sieve separates recommended), and reported result must all be standardized. A mass fractal has been reported from a 5-sieve separation, yet a K_{sat} estimation cannot be made unless there is some associated large-pore estimate included in the measurement.

6. Seal formation and amelioration in the field as a sequence during a succession of rains and dry periods needs attention. The seal may be completely ameliorated or will form again faster

than before the interim dry period in the field. Does the presence of new crop residue above 10 cm have an effect on this function?

7. There are many new techniques to be applied making sure that the soil material and organic components are undisturbed. Some of these new techniques were discussed, however, accurate measures of crust formation need be made using new surface roughness (Huang and Bradford, 1993; Magunda, 1992) and porosimetry (Pikul and Zuzel, 1994) techniques. New techniques with TDR ought to be used (Baker and Spaans, 1994) to measure wetting and drying sequences in undisturbed soil either under controlled or natural ambient conditions. These new techniques and those related to K_{sat} estimated from soil structure are needed to make the link between water stable aggregation and the simple one dimensional saturated flow model.

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Shrinking/Swelling, Freezing/Thawing And Grazing Effects On Infiltration

G.N. Flerchinger[1], G.W. Frasier[2], R.B. Grossman[3], K.N. Potter[4],
B.S. Sharratt[5], and J.D. Williams[6]

Abstract

Infiltration of water into soils is the governing hydrologic process which partitions available precipitation into surface runoff, available soil moisture, and subsurface flow. The infiltration capacity of soils can have a large degree of temporal variability due to the poorly understood processes of shrinking/swelling, freezing/thawing, and grazing. A brief discussion of the effects these processes have on the temporal variability of infiltration is presented, and future research directions are proposed to better understand the impacts of these processes. Future research needs include: 1) Relating the shrinkage characteristic of soils to intrinsic soil properties and management practices; 2) Accurate characterization of the relationship between soil water content, ice content and frozen soil infiltrability, which may require better measurement and characterization of ice content and structure; and 3) Better definition of the factors controlling infiltration on rangelands and the tendency of grazing to influence these factors.

Introduction

Methods to quantify infiltration, including the Curve Number and Green-Ampt equation, have been developed and refined many times over. Indeed, infiltration has been studied for decades, and infiltration into normally consolidated porous medium is well understood and can be predicted with a high degree of accuracy with developed methodologies. However, infiltration into soils affected by processes which create a high degree of temporal variability in infiltration is poorly understood. Such processes include shrinking/swelling, freezing/thawing, and grazing. Progress in these areas have been hampered in part by the difficulty in measuring, quantifying and understanding the processes affecting infiltration. Before we can quantify the affect these processes have on infiltration, we must first understand the processes themselves. This paper presents a brief discussion of the effects that these processes have on the temporal variability of infiltration and proposes future research directions.

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[1]USDA-ARS, NWRC.800 Park Blvd., Ste 105, Boise, ID 83712

[2]USDA-ARS, RRR, 1701 Centre Avenue, Fort Collins, CO 80525

[3]USDA-NRCS, 100 Centennial Mall North, Lincoln, NE 68508

[4]USDA-ARS, NRSR, 808 E. Blackland Rd., Temple, TX 76502

[5]USDA-ARS, NCSCRL, 803 Iowa Ave., Morris MN 56267

[6]USDA-ARS, CPCRC, Box 370, Pendleton, OR 97801

Shrink/Swell Effects On Infiltration

Soil shrinkage is the act or process of soil material contracting to a lesser volume when subjected to the loss of water (Mitchell, 1996). Soil swelling is the increase in volume with the addition of water. When the volume change is sufficiently large, the result is the formation of soil cracks on drying or the closure of cracks on wetting. In extremely dry, cracking soils, cracks may extend deep into the soil profile.

Three dimensional shrinkage has served as the basis for modeling soil crack development and resultant water infiltration in heavy clay soils (Bronswijk, 1990). Vertical height differences have been used to estimate the soil profile water content using functions to estimate total volume change from the change in vertical height (Mitchell, 1991; Bronswijk, 1991).

Characterization of Shrink/Swell Impacts

Potential of soils to change volume with changing water contents have been characterized with the coefficient of linear extensibility (COLE) index (Grossman et al., 1968) which is a measure of one-dimensional shrinkage in a soil between its field capacity and oven dryness. More recently, soils have been characterized by their shrinkage characteristic, i.e. the change in volume of soil relative to the change in volume of water extracted (Mitchell, 1991).

The shrinkage characteristic is best explained when combined with the conceptual model proposed by McGarry and Malafant (1987). In this model, soil shrinkage has been divided into three separate phases with loss of water.

(1) Structural shrinkage - Shrinkage that is less than water loss due to water draining from large, stable pores at high water content.
(2) Basic shrinkage - The middle phase of soil shrinkage between structural and residual shrinkage: it refers to the fundamental shrinkage process. Often this phase is known as unitary or normal shrinkage if the change in soil volume is equivalent to the change in water volume. Soil pores are collapsing as water is lost in this phase.
(3) Residual shrinkage - Shrinkage that is less than water loss during the final stages of drying. Air is entering the soil to occupy some but not all of the drained pores.

An additional phase may be included:

(4) Zero Shrinkage - No volume change with change in water content.

Soil volume change is often linear with change in water content within a phase. Soils vary in the extent that each shrinkage phase occurs and the shrinkage characteristic of a soil may vary with different phases of shrinkage. Shrinkage characteristics are affected by soil properties such as clay type and size distribution (McCormack and Wilding, 1975), sodicity and salinity (Smith et al., 1985), and plant distribution and root structure (Mitchell and van Genuchten, 1992). Grossman et al. (1990) have subsequently developed a formulation for an intermediate coefficient of linear extensibility (COLEi) for intermediate water contents. Intermediate water contents can be estimated and COLEi can be placed on a basis related to weather and vegetation.

Temporal Variability of Infiltration Related to Shrink/Swell

Temporal variability of infiltration can be extremely large in soils with large shrink/swell potential. Infiltration in high clay soils which exhibit large shrink/swell is generally characterized as very rapid when the soils are dry and very slow when the soils are wet. Water infiltration in these soils is determined to a great extent by the presence of soil cracks. Obviously, when these soils are near saturation and the cracks are swollen closed, the effect of cracks is less than when the cracks are open. Ritchie et al. (1972) found that while field-measured hydraulic conductivity was low, about 2.5 cm d^{-1} in a saturated Houston Black Clay, this was much greater than that measured in soil cores. This was attributed to the loss of continuity of sloping soil cracks at the walls of the soil cores. Still, even when the soils are wet and cracks in the surface are swollen shut, the soil cracks provide continuous soil pores available to conduct water. Jarvis and Leeds-Harrison (1987) noted the importance of continuous macropores provided by cracks in recharging the soil profile even when the soil is wet and swollen near the maximum.

As the soil drains and dries, either from evaporation or transpiration, shrinkage occurs and soil cracking may take place. Cracking may occur at soil water contents below the wilting point. Bronswijk (1991) noted that the whole pressure range in which water uptake by plant roots takes place lies within the unitary shrinkage phase of a Bruchem heavy clay soil (very fine clayey, mixed, illitic-montmorillonitic, Typic Fluvaquent) in the Netherlands. Stirk (1954) found for several cracking clay soils that half the total shrinkage measured occurred on the dry side of the wilting point. Evaporation from the cracks increase the drying rate of the soils (Adams and Hanks, 1964; Adams et al., 1969).

Profile recharge can be quite different in a cracked soil than in an uncracked soil. When the rainfall rate is greater than the infiltration rate of the bulk soil, water will flow into the cracks and deep into the soil profile without passing through the soil matrix. This has been termed by-pass flow (Bouma and Dekker, 1978). The phrase internal catchment has been used to describe the water stored within the crack volume (Van Stiphout et al., 1987). The soil profile is re-wet almost simultaneously at all depths as water is absorbed by soil aggregates forming the crack wall (Jarvis and Leeds-Harrison, 1990; Van Stiphout et al., 1987). The amount of water infiltrated is related to the volume of the cracks (Stirk, 1954). Water entering cracks has been reported as making up to 74% of the total water infiltrated in some studies (Mitchell and van Genuchten, 1993).

The crack volume may be estimated if the soil shrinkage characteristic is known and soil water contents are monitored. The change in soil volume associated with a change in water content can be separated into crack volume and soil subsidence with the following equations:

$$\Delta z = z - z\left(\frac{V - \Delta V}{V}\right)^{1/rs} \tag{1}$$

$$\Delta V_{CR} = \Delta V - z^2 \cdot \Delta z \tag{2}$$

with: Δz = change in layer thickness due to shrinkage/swelling (m)
ΔV_{CR} = change in crack volume (m^3)
V = volume of cube of soil matrix before shrinkage/swelling
ΔV = change in volume of soil matrix due to shrinking/swelling (m^3)
z = layer thickness before shrinkage/swelling (m)
rs = geometry factor equal to 3 for isotropic shrinkage

As noted earlier, rapid infiltration occurs when soil cracks are open, and very slow rates of infiltration occur when the soils are wet. Thus, any factor affecting soil water contents can influence the infiltration rate. This may include weather, vegetative cover and associated evapotranspiration, and management. A high-intensity rain occurring under dry conditions may result in by-pass flow and internal catchment which could quickly recharge the soil profile. Conversely, an extended low-intensity rain may induce soil swelling which could close surface cracks, limiting infiltration without resulting in by-pass flow and internal catchment. It is conceivable that a series of storms could result in very different infiltration depending upon the sequence and intensity of storms.

Soil shrinkage may have limited effect upon irrigated soils because most cracking will occur at water contents drier than allowed in irrigated agriculture. Tillage of cracking soils may also limit water infiltration by forming a mulch which will inhibit water flow.

Expression of surface-connected cracks is strongly dependent on the mechanical continuity of the soil fabric to permit propagation of strain. If soil is freshly tilled to 15 or 20 cm and not subject to water compaction by wetting, the propagation of strain is small and cracks are weak or nonexistent. If, on the other hand, the same soil is subject to compaction while moist, such as a tractor wheel, the strain propagation is greatly increased and cracks are apparent. Thus, because of this control by the degree of near surface compaction, the pattern of surface connected cracks in a field shows large spatial and temporal variation.

Future Directions

Studies have shown the high degree of temporal variability depending upon soil water content. This is even more important in soils with large shrink-swell potential. The shrinkage characteristic of soils needs to be determined and related to intrinsic soil properties as well as the effects of management practices. Combining this information with estimates of soil water content will aid in predicting infiltration.

Freeze/Thaw Impacts On Infiltration

Soil freezing can dramatically reduce the permeability of frozen soil. Ice blocks the soil pores resulting in large runoff events from otherwise mild rainfall or snowmelt events. Runoff and erosion from frozen soil are widespread phenomena. Rain and/or rapid snowmelt on impermeably frozen soil is the leading cause of severe flooding and erosion in many areas of the world (Johnson and McArthur, 1973).

The blockage of pores by ice increases the tortuosity of flow paths through the soil matrix and reduces the permeability and infiltration of soils. Frozen soil infiltration rates generally decrease with an increase in soil water content (Motovilov, 1979; Larin, 1963). Kane and Stein (1983) noted that infiltration decreased from 0.7 cm/h to 0.01 cm/h as the water content of a relatively dry silt loam was increased to near saturation. They also noted little difference in infiltration rates between the unfrozen and frozen, but relatively dry, soil. Despite reductions in infiltration that typically occur as wet soils freeze, Thunholm et al. (1989) noted that rates may exceed 5 cm/h in heavy clay soils that are susceptible to cracking upon freezing.

Water potential and temperature gradients within a freezing soil profile induce water migration from the unfrozen subsoil toward the freezing front and result in the formation of ice lenses. These lenses, typically formed in wet soils having a high proportion of silt, are often a barrier to infiltration (Kane, 1980). Melting of ice lenses during infiltration experiments is evident from the temporal variation in infiltration data (Kane and Stein, 1983). Indeed, temporal trends in the rate of infiltration into frozen soils are different from trends in unfrozen soils (Kane and Stein, 1983). In the absence of ice lenses, infiltration rates in fine-textured versus coarse-textured soils may be higher owing to the greater amount of unfrozen water in the fine-textured soil (Burt and Williams, 1976).

Tillage and surface characteristics influence the porosity as well as the water and heat transport processes in soils. Infiltration is typically greater in tilled versus untilled soils when frozen, due to the creation of macropores by tillage (Zuzel and Pikul, 1987). However, tillage is likely to have little effect on infiltration as the freezing front descends below the depth of tillage (Pikul et al., 1992). Infiltration is enhanced in frozen soil with greater roughness. Gor'kov (1983) suggests that rough surfaces create more localized variations in snow cover as well as pore ice content, both of which affect infiltration. Surface cover also impacts infiltration into frozen soils (Zuzel and Pikul, 1987; Haupt, 1967; Stoeckeler and Weitzman, 1960).

Characterization of Freeze/Thaw Impacts

Several techniques have been used for in situ determinations of infiltration into frozen soils. These techniques assess water flux into soils using a bore hole (Kane, 1980), annular cylinder (Stoeckeler and Weitzman, 1960; Harris, 1972), double-annular cylinder (Kane and Stein, 1983), or square metal plate (Haupt, 1967; Wilkins and Zuzel, 1994; Zuzel and Pikul, 1987). One- dimensional water flow has also been achieved in the field using large soil monoliths lined with plastic (Kuznik and Bezmenov, 1963; Thunholm et al., 1989) and in the laboratory using plastic pipe (Engelmark, 1988). Water at a known temperature close to 0°C is typically applied using a rainfall simulator (Haupt, 1967; Zuzel and Pikul, 1987) or by flood irrigation (Kane, 1980; Wilkins and Zuzel, 1994) to assess infiltration. Granger et al. (1984) assessed snowmelt infiltration on the prairie soils of Canada by measuring the temporal changes in soil water content during snow ablation.

Accurate quantitative descriptions of frozen soil infiltration is lacking, partly due to experimental difficulties in measuring infiltration into frozen soil and characterizing the ice content and structure within the frozen soil. Presently, there is no quantitative means of directly measuring ice content of the soil, which is the single most important factor to reduce infiltration potential upon freezing. With the advent of time-domain reflectometry (TDR) to measure liquid water content, ice content has been computed as the residual between liquid water content and total water content measured by neutron probe or gravimetric samples. Even so, ice content, pore blockage and infiltration rate changes as water infiltrates into frozen soil. Introducing water into frozen soil causes freezing of the infiltrating water, thawing of the ice contained within the soil, or both. Thus, there is no steady-state infiltration rate analogous to that in unfrozen soil. An approach to circumvent this problem is to use an alternate fluid that remains viscous at subfreezing temperatures. Fluids such as ethylene glycol (Harris, 1972) as well as air (Saxton et al., 1993; Seyfried and Flerchinger, 1992) have been used as test fluids for characterizing infiltration of frozen soils. Measured permeability for these alternate fluids can be related to hydraulic conductivities by accounting for differences in density and viscosity (Cary et al., 1989).

Most approaches for estimating infiltration into frozen soil make use of some adjustment to the saturated hydraulic conductivity or curve number when the soil is frozen. However, models differ considerably in the theory and sophistication used to determine whether the soil is frozen and the adjustment to unfrozen conditions. Adjustments for frozen conditions may be based on: simply whether the soil is below freezing temperatures; the amount of ice present in the soil; or the available porosity remaining in the frozen soil. Very simple approaches use essentially a simple on/off switch for accounting for frozen soil effects, in which the curve number or hydraulic conductivity is set to an arbitrary value to cause reduced infiltration when the soil is frozen (Knisel et al., 1985). Slightly more sophisticated methods use a adjustment factor to hydraulic conductivity based on antecedent water content or ice content of the soil (Savabi et al., 1995). Many detailed approaches for estimating hydraulic conductivity in frozen soils assume the hydraulic conductivity and water retention characteristics are the same for frozen and unfrozen soils (Flerchinger and Saxton, 1989; Lundin, 1990; and Grant, 1992). Thus, hydraulic conductivity for infiltration is based on the unsaturated hydraulic conductivity computed from the available porosity (total porosity less volumetric ice content).

Temporal Variability of Infiltration Related to Freeze/Thaw

The permeability of soil can vary dramatically depending on whether the soil is frozen or not. The permeability of frozen soil is affected by the occurrence, depth, and ice content of the soil, which is dependent on the interrelated processes of heat and water transfer at the soil surface and within the soil profile. Heat and water transfer at the soil surface are governed by the meteorological and environmental conditions at the soil-atmosphere interface. Soil freezing and thawing can also alter soil physical properties or structure that also impact infiltration. Changes in aggregate stability (Hinman and Bisal, 1973) or stress fractures (Saxton et al., 1993) caused by freezing affect soil structure and pore continuity and thus affect infiltration even after the soil is thawed.

Future Directions

Prediction of infiltration into frozen soil requires knowledge of frozen soil occurrence depth, and effect of soil freezing on infiltration. Methodologies for describing frozen soil depth and occurrence have been developed with reasonable accuracy (Flerchinger and Saxton, 1987;

Lundin, 1990; and Grant, 1992), but this is not the case for frozen soil infiltration. Accurate characterization of the relationship between soil water content, ice content and frozen soil infiltrability is a difficult, but vital first step. Better measurement and characterization of ice content and structure (i.e. the presence of ice lenses) will likely be necessary before great strides in understanding infiltration into frozen soils are achieved.

Grazing Impacts On Infiltration

Some researchers have reported that high intensity grazing by cattle can reduce the infiltration rates of soils causing increased runoff and possibly increased soil erosion on semiarid rangelands (Rauzi and Hanson 1966; Rauzi and Smith 1973; Hart et al. 1988; Gamougoun et al. 1984; Warren et al. 1986). It has been stated that reducing the intensity of cattle grazing will allow the areas to return to pristine conditions with respect to water infiltration (Wald and Alberswerth 1985).

Evaluating the hydrologic impact of removing cattle from areas of historic long-term grazing is difficult using natural precipitation events. It is virtually impossible to find identical watersheds for rainfall-runoff instrumentation where replicated studies can be conducted. Obtaining a representative distribution of rainfall intensities and quantities for quantitative evaluation of the effects of grazing treatments on infiltration processes under natural rainfall events usually requires many years of data collection.

Characterization of Grazing Impacts

Rainfall, runoff and infiltration characteristics can be estimated using rainfall simulation. Rainfall simulators apply a reproducible rate and quantity of water to an area allowing the evaluation of factors such as vegetation density and composition, soil bulk density, and surface roughness on infiltration and runoff.

Frasier et al. (1995, 1996) presented results of infiltration/runoff studies using a rotating boom rainfall simulator on two native shortgrass rangeland sites to evaluate the effects of cattle grazing for 53 and 12 years, respectively. As expected, the total runoff quantities and rates were higher in the heavy grazed treatment than in the lightly grazed treatment. There was also a decline in the dry period equilibrium runoff rates with succeeding years. The declines in runoff rates were most apparent in the heavy grazed treatment.

Runoff in a wet equilibrium state (wet run) was less affected by initial soil moisture conditions and differences may be more a factor of the soil physical properties such as bulk density in the top layers of the soil profile. Soil bulk density in this zone can be affected by grazing cattle trampling and/or soil compaction by raindrop impact following removal of the canopy cover by the cattle. The results showed significant differences in equilibrium runoff among all grazing treatments during the wet run period ranging from 46% in the heavy grazed exclosure to 10% in the light grazed treatment. Only the heavy grazing intensity showed a major decrease with years. There was no correlation with changes in wet run equilibrium runoff ratios to the soil bulk density.

Runoff data from wet soil under a high precipitation intensity (wet-wet run) minimizes the effect of surface roughness. During the wet-wet run the soil surface is saturated and the simulator application rate substantially exceeds the soil infiltration rate. The infiltration rate approaches a final steady state level. The wet-wet mean equilibrium runoff ratios were 65%, 60%, and 35% for the heavy, moderate, and light grazing intensities, respectively indicating differences in the infiltration parameters at the deeper depths with the moderate and heavy grazing intensities.

It is possible that there were some changes in the surface bulk densities with time as a result of the elimination of cattle traffic and an increased protection of the soil surface from raindrop impact with increased canopy cover. Abdel-Magid et al. (1987) found that very small increases in bulk density produced large changes in infiltration rate. Since there was no change in runoff on the heavy grazed area during the wet-wet period it was assumed that parameters affecting changes in the runoff ratios in the wet and dry run did not change in the underlying soil layers which typically controls the infiltration rate during the wet-wet run.

Temporal Variability of Infiltration Related to Grazing

Frasier et al. (1995, 1996) found that immediately following removal of the cattle, equilibrium runoff ratios for the dry run ranged from 61% for the heavy grazing intensity to 11% for the light grazing intensity treatment. Two years later the dry run equilibrium runoff ratio in all treatments had decreased to less than 50% of the initial values. The heavy grazed treatment had the greatest decrease. It had been anticipated that biomass production would increase following the cattle removal with a corresponding increase in infiltration and a decrease in runoff. However, there was actually a greater increase in biomass production in the light and moderate grazed exclosures compared to the heavy grazed exclosure. This implies that biomass production alone is not a good indicator of water infiltration/runoff rates at the sites evaluated. The data also indicates that antecedent soil moisture at the time of the studies was not a controlling factor in the dry run equilibrium runoff ratios. The results indicated that changes in the infiltration parameters may occur as quickly as 2 years after removal of cattle from shortgrass prairie. The decreases in the runoff from the dry run and wet run on the heavy grazed area during the 3 year study suggest an improvement in soil surface infiltration properties but, these changes do not seem to have been influenced in the deeper soil layers.

The effect of grazing on surface sealing and microbiotic crusts can also influence infiltration. This can have either positive or negative effects on infiltration depending on the nature of the microbiotic crust. Effective saturated hydraulic conductivity decreased as lichen-dominated microbiotic crust increased in soil cores collected from a variety of disturbed sites in southern Utah (Loope and Gifford 1972). Lee (1977) hypothesized that algal crusts reduce permeability in arid and semiarid soils of Australia, but did not present data to support his contention. However, various lichens are capable of rapidly absorbing from 1.5 to 13 times their dry weight in water (Galun et al. 1982). In a semiarid woodland in Australia, steady-state infiltration increased with microbiotic development at a grazed site and tended to decrease with microbiotic development at an ungrazed site (Eldridge 1993a), which was attributed to differences in soil physical properties, e.g., the degree of macroporosity development. Eldridge (1993b) argued that soil physical properties, porosity and aggregate stability are controlling infiltrability factors and that as these properties are damaged, the presence of microbiotic crust becomes more important. Meyer and García-Moya (1989) measured microbiotic crust cover on ungrazed and grazed sites on the semi-arid gypsum plains in northern San Luis Potosí, Mexico. Grazed sites had less microbiotic crust cover that ungrazed sites, (30.0:5.0%, 85.0:22.5%, and 47.5:7.5%, respectively), but more soil moisture.

Future Directions

Studies have shown a high spatial variability of infiltration among the plant bases and the interspaces (Blackburn et al., 1992; Link et al., 1994; and Frasier, 1996). This variability coupled with surface roughness presents problems in relating measured runoff from simulator studies to surface measured parameters. There are also indications of a surface soil water repellency that reduces infiltration during the initial wetting periods. This water repellency gradually dissipates in a time function decay. These two problems (spatial variability of infiltration and surface water repellency) need to be quantified to better describe the infiltration process on undisturbed rangeland soils.

The impact of grazing on infiltration has limited data available and is contradictory, i.e. grazing can increase or decrease infiltration. This is influenced by the impact which grazing has on the factors controlling infiltration at a particular site, i.e. soil physical factors versus the presence of surface seals or microbiotic crusts. Grazing can break up infiltration-inhibiting surface crusts, destroy infiltration-enhancing macroporosity, or inhibit the presence of lichens which have tremendous capacity to absorb water. An important question which needs addressed is how extensively developed microbial crusts in a range of soil textures influence infiltration, and if the components of microbial crusts, including algae, cyanobacteria, lichens, and mosses, affect infiltration on disturbed soils. Thus, future research should attempt to quantify the controlling factors affecting infiltration on rangeland sites, and the tendency of grazing to influence these factors.

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Specific Goal 4

Innovations in Computer Modeling of Variabilities

Topic 4 Summary: Status and Future of Modeling Spatial and Temporal Variability of Infiltration

R. E. Smith[1] and D.C. Goodrich[2]

Introduction

One important objective of research on the variability of infiltration is to be able to incorporate a measure of variability in hydrologic models to make them valid over a variety of time scales and a variety of spatial scales. Models can be thought of as symbolic representations of our knowledge, and it has long been recognized, as reflected in the motivation for this workshop, that our representation of infiltration variability is often crude in comparison with the variations seen in nature. Temporal variability may be considered as manifesting itself most dramatically in the effects that tillage may have on the intake properties of a soil. Temporal variability also can be observed, albeit to a much smaller extent, in seasonal changes that can be observed on natural watersheds. Spatial variability is the dominant problem in dealing with scale effects in simulating runoff on watersheds, and in extrapolating plot measurements to estimate watershed runoff. We will briefly summarize our interpretation of the state of current knowledge (i.e., models) and the major challenges facing us in these areas. We will assume a general point model and describe various treatments for applications to larger scales.

Definitions. In the following we will assume that at a point, and if current soil conditions are known, local infiltration behavior for a uniform simple soil can be described in relation to local soil hydraulic properties. Infiltration rate, f, equals the soil-limited infiltration capacity, f_c, when water is supplied at a rate r exceeding that capacity. For lower values or r, $f = r$. In either case, infiltrated depth, I, is defined as

$$I = \int_0^t f dt \tag{1}$$

Two basic infiltration parameters describe infiltration capacity: First, effective, saturated hydraulic conductivity, K_s constitutes the asymptotic value of f_c if the profile is homogeneous. Second, capillary drive, G (often termed H_c), is a basic soil parameter defined as

$$G = \int_{-\infty}^{0} \frac{K(h)}{K_s} dh \tag{2}$$

in which h is soil capillary head, K(h) is the conductivity-capillary head relation. G [units of L] is effectively a K - weighted value of h. Also, infiltration is sensitive to the soil water deficit, $\Delta\theta$, defined as $\theta_s - \theta_i$, where θ_s is the maximum soil water holding capacity, by volume, and θ_i is the soil water content at the beginning of rainfall. Given these parameters, a quite general, basic

[1] ARS-USDA, AERC Foothills Campus, CSU, Fort Collins, CO

[2] ARS-USDA, Southwest Watershed Res. Center, 2000 E. Allen Rd., Tucson, AZ

infiltration relation can be described by defining the following dimensionless variables:

$$f_c^* = \frac{f_c - K_s}{K_s}$$
$$I^* = \frac{I}{G\Delta\theta} \tag{3}$$

Then the infiltration equation can be given as

$$f_c^* = \frac{\alpha}{\exp(\alpha I^*) - 1} \tag{4}$$

For α approaching 1, Eq. (4) is the Smith-Parlange (1978) relation, and in the limit as α approaches 0, Eq. (4) becomes the Green-Ampt infiltration relation. Employing I^* rather than t as the independent variable eliminates need for a separate computation of ponding time.

Models of Temporal Variability

Agricultural soils undergo a variety of changes in time, caused by both mechanical actions and the natural actions of weather. Tillage can cause enormous changes in bulk porosity and create a dual porosity medium composed of soil "clods", and can at the same time create wheel track compaction of soil in a few furrows. The effect of a given type of tillage implement further will vary with the specific tillage history, the soil texture, and the soil water content at the time of tillage. Significant amounts remain to be learned about interrelations of all these factors before a robust tillage infiltration model can be proposed.

Rainfall energy and rewetting on loose soil will often cause particle dispersion and structural reformation at the surface which can create a crust layer. The factors influencing this crust development and its properties in relation to the parent soil are poorly understood. In the Opus model (Smith, 1992), this transition is modeled as a function of soil clay content and cumulative rainfall energy. Clay content is assumed to govern the ultimate reduction ratio for K_s which can be attained in a crust. This ratio is assumed largest for moderate clay amounts.

Further wetting can cause a slow reformation of tillage induced "clods" and a reduction to more natural bulk porosity. Soils high in clay may be subject to swelling during wetting, and subsequent cracking upon drying. There has been some modeling of this process, which creates and destroys a special kind of macroporosity and a two-dimensional infiltration opportunity. The WEPP model contains a simulation of shrinkage cracks for clay soils, assuming the fraction of area cracked is a function of clay content, swelling ratio, and water content. Cracks both act as macropores and induce two-dimensional water intake.

Frozen Soils: On any soil, especially in climates subject to annual freezing, there are seasonal changes that are complex and difficult to model. In many northern latitudes, rain or snowmelt on seasonally frozen soils is the single leading cause of severe runoff and erosive events. A separate paper by Flerchinger (this volume) discusses modeling infiltration in frozen soil.

Infiltration Models for Heterogeneous Soil Conditions

Layering

Soil heterogeneity as it affects infiltration may be viewed as two-dimensional: both vertical and horizontal variations. Infiltration into layered soils has received some attention in the past. Early work focused on the application of the Green-Ampt model to special cases of layering, where the soil K_s varied monotonically (Bouwer, 1976). Often it was assumed that K_s could be obtained by use of saturated flow computations. In fact, however, for an arbitrary layering it cannot be assumed that the soil is saturated above a wetting front in a layered soil. Moore(1981) and Smith(1990) have published models that rigorously treat infiltration into a two-layer soil. Any number of layers can be treated by assuming the general functional relation such as Smith-Parlange or Green-Ampt with a capillary parameter G (whose effective value changes with wetting front position) and finding an effective asymptotic value of $\hat{K}_s$ for the current wetting front by solving the steady flow equation through all wetted layers. Internal boundary conditions at each layer interface must be satisfied, and saturation of layers above the wetting front cannot be assumed. This is not a trivial exercise, but does provide a general infiltration model for layered profiles.

Infiltration modeling through layers includes the case of temporally changing crusts mentioned above. Where crusts are significant infiltration controls, the ideas of Mualem and Assouline (1989) concerning a gradation from surface properties to subsoil properties, rather than a distinct layer, deserve further study. Mualem and Assouline have only studied this type of crust under steady flow.

Spatial Soil Heterogeneity

The treatment of natural heterogeneity is one of the greatest challenge in hydrologic modeling at larger scales is (Smith et al., 1994; Bloschl and Sivapalan, 1995). For the purpose of this discussion, "large scale" implies the field or hillslope scale (characteristic lengths > 100m) and beyond. Meter and sub-meter scale soils and infiltration heterogeneity can be classified as random while larger scale variability due to changes in soil type can be classified as organized variability (Bloschl and Sivapalan, 1995). For purposes of large scale modeling, it is assumed that organized variability can be resolved with the use of geographic information systems and treated within a distributed hydrologic model via variation in infiltration parameters from one model element to another. Small scale (sub-grid or sub-model element) random infiltration variability is often be treated via a statistical or probability distribution (Smith and Hebbert, 1979; Woolhiser and Goodrich, 1988; Binley et al., 1989; Smith et al., 1990; Wood et al., 1990). It is crucial to understand infiltration phenomena at the field and small watershed scale as this is the typical size of land areas subject to management, and is the scale of our major source of field data.

Considerable research has been published on the treatment of flow, both saturated and unsaturated, in heterogeneous porous media. Most of this work is not directly relevant to infiltration. The perturbation approach for stochastic differential equations, (Mantoglu and Gelhar, 1987) for example, is not applicable either near soil saturation or near a boundary. Further, most work on unsaturated flow near the soil boundary has focused on the areal mean wetting fronts and their moments, and not looked at areal infiltration heterogeneities (e.g. Bresler and Dagan, 1983; Chen *et al.*, 1994).

Models for Soil Variability: There have been many studies to evaluate a statistical model for the random spatial heterogeneity of soil properties, which are directly related to the spatial properties of infiltration. K_s as a spatially random variable has often been a subject of inquiry (e.g., Nielsen *et al.*, 1973). The question is not only one of establishing the distribution function (log-normal is almost always found applicable) but also the spatial structure: correlation scale, semivariogram, or other measure of spatial relation. A further question for hydrology is the relation of spatial statistical structure of a soil property to the scale of hydrologic interest. There are more theoretical questions involved than can be discussed here. What is important to remember is that one cannot arbitrarily average over heterogeneous areas as one considers larger scales. Small scale variations may have a significant effect for larger scale behavior. To model infiltration over a heterogeneous area, one may either reproduce the small scale variations, or establish by detailed simulation or by theory a method of lumping that is valid for the purpose.

Infiltration Model to Account for Spatial Heterogeneity: It should be understood that the parameters for the net behavior of an ensemble of nonlinear processes cannot be obtained by the average of the parameters of each member of the ensemble. Moreover, there may be no effective stationary parameter value that will allow a single process realization to mimic the behavior of the ensemble. These two facts must be kept in mind in modeling infiltration for areas containing infiltration heterogeneity. A further complication is involved in spatial interactions. The important spatial interrelations for adjacent infiltrating areas are upstream/downstream relations, not simple spatial correlation lengths. If an area producing early runoff flows away from an adjacent area that is pre-ponding, the distance or correlation between those points is unimportant. On the other hand, runoff onto a preponding area will act as an increase in rainfall rate and cause accelerated ponding and runoff production.

An overall picture of the ensemble behavior of a heterogeneous watershed is diagramed in Fig. 1. The areal distribution is diagramed at two successive times, during which rainfall is assumed to have dropped. Parts of the catchment (where $r > K_s$) have ponded and infiltration is controlled by capacity, f_c. The overall area has an expected value for $K_s = \xi(K_s)$, and an effective value, $K_e(r)$, which depends on rainfall. Hawkins and Cundy (1987) were one of the earliest to

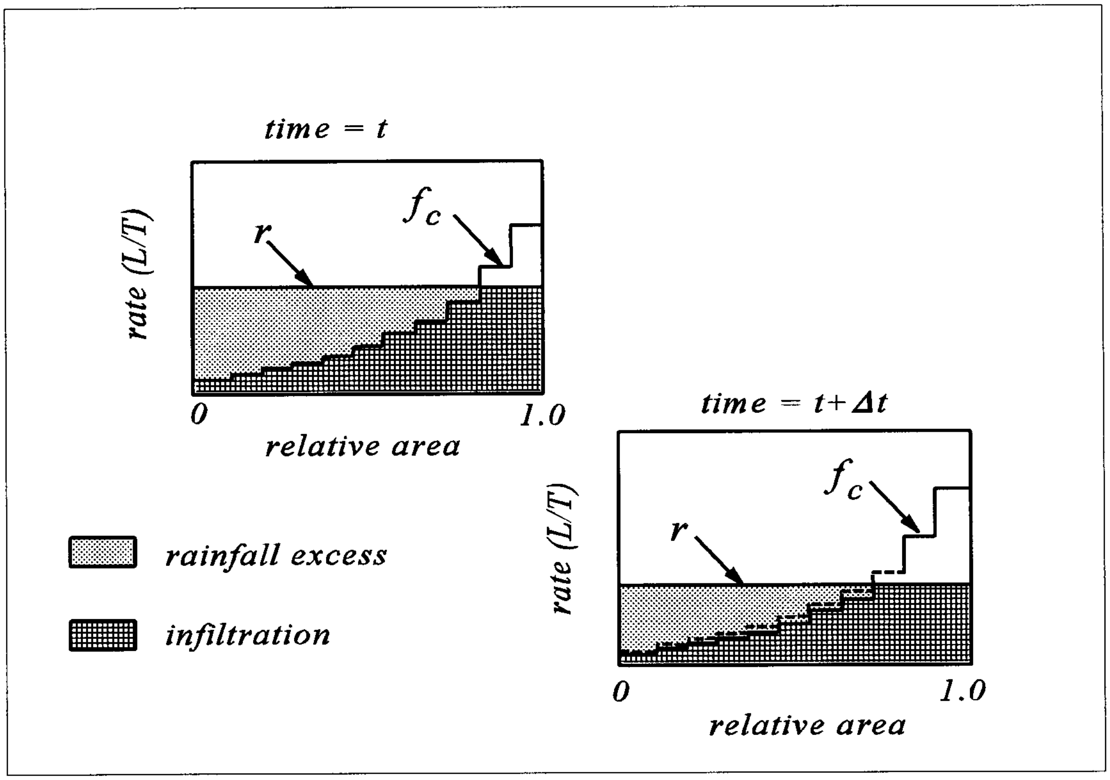

Fig. 1. Illustration of infiltration capacity and infiltration rate changes in space and time during a rainfall event of rate *r*. Note that infiltration is always limited by r, and f_c deceases in time when limited by soil rather than r. The dotted line indicates f_c at the previous time.

point out that areal effective K_e can be described as

$$K_e = \left[1 - P_k(r)\right] r = \int_0^r k p_k(k)\, dk \tag{5}$$

given the probability distribution p_k and cumulative distribution P_k for K_s . Using K_e in place of K_s in Eq. (3), the areal infiltration equation may be expressed (Smith, unpublished) as

$$f^* = r^* \left\{ 1 + \left[\frac{r^*}{\alpha} \left(e^{\alpha I^*} - 1 \right) \right]^c \right\}^{-1/c} \tag{6}$$

Note that this is not an infiltration capacity equation, because with a randomly distributed K_s, a ponding time is not defined, and for small values of I this equation properly depicts f ~ r, as shown in Fig. 2. The parameter c in Eq. (6) is a function of r^* and the coefficient of variation of K_s.

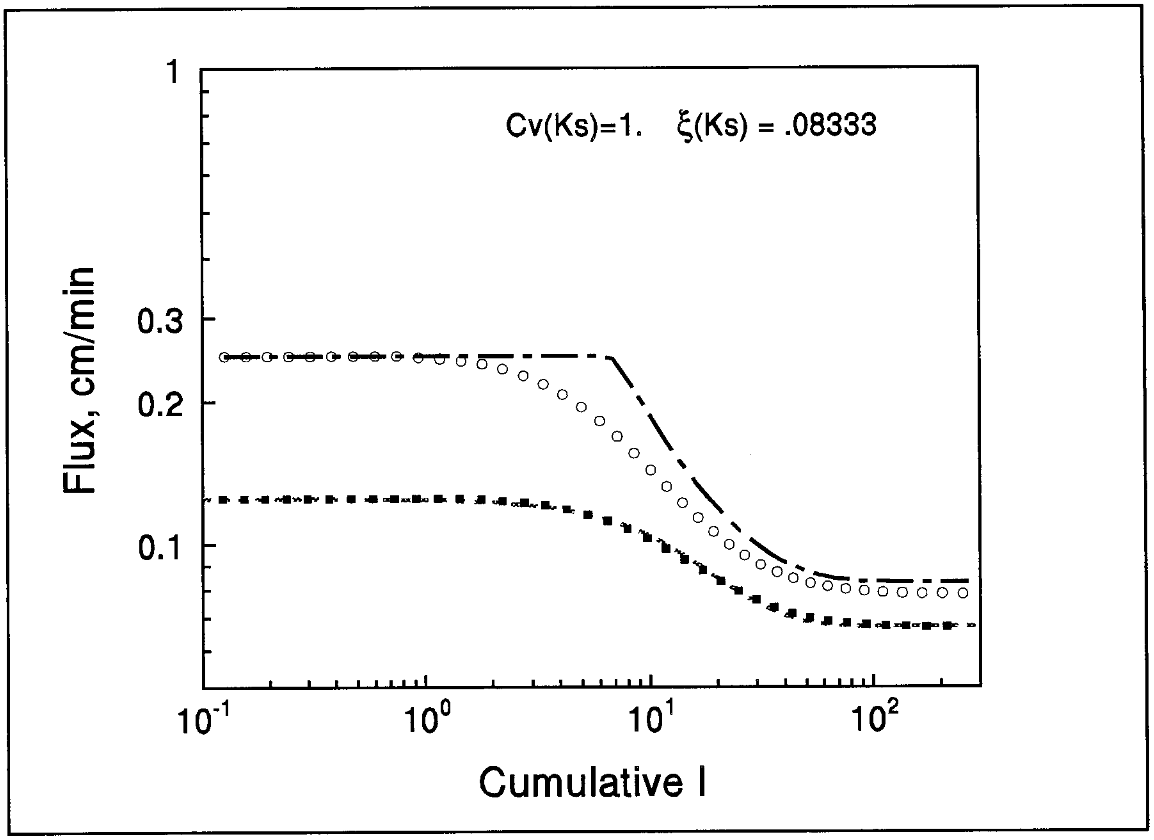

Fig. 2. Areal ensemble infiltration relations for two intensities, compared with the point infiltration function for the expected value of K_s (dot and dashed line), when the coefficient of variation of K_s is 1.

Smith et al. (1990) demonstrated three methods to estimate infiltration and runoff on a small watershed, including ensemble net infiltration (ignoring spatial interaction), stratified (Latin Hypercube) sampling simulation using parallel strips (as in Woolhiser and Goodrich, 1988), and two dimensional sampling over the watershed with simulated upstream/downstream interactions. These were shown to have different degrees of accuracy, but all simulated peak flows were significantly larger than for the uniform, average infiltration assumption. It was also demonstrated that the effect of areal heterogeneity on runoff is most significant for the common case where runoff is a small portion of total rainfall.

Latin Hypercube sampling with parallel strips can be applied at scales from plots to small catchment surfaces. The method is illustrated in Fig. 3. Each strip represents an equally likely value of K_s, taken from the cumulative distribution as shown. The strip arrangement illustrated will not simulate runon-runoff phenomena, but they increase significantly the ability of the model to treat the effects of heterogeneity on runoff.

Surface Microtopography: The interaction of runoff and soil infiltration should in all cases involve the actual surface shape. Microtopography has been shown (Woolhiser, et al., 1996) to

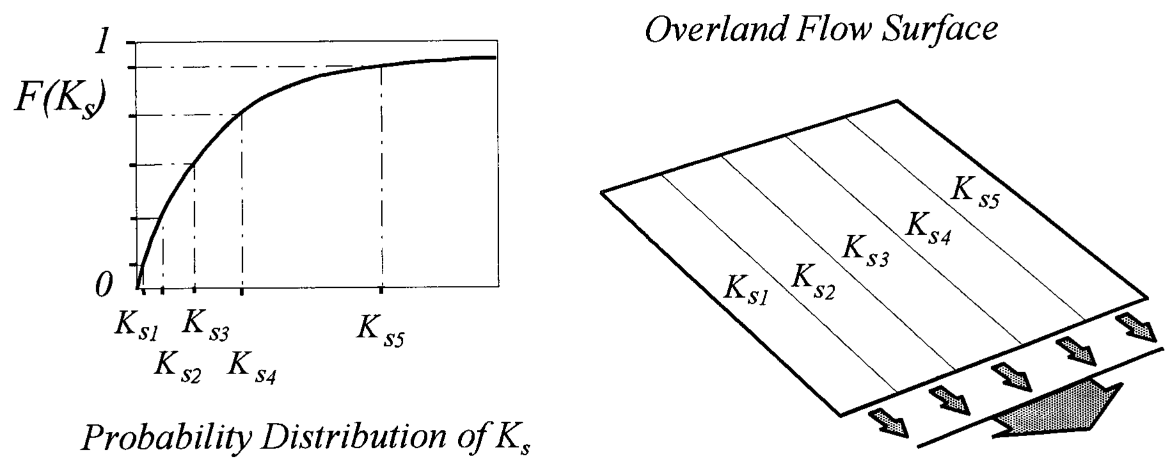

Fig. 3. Latin hypercube parallel strip method for simulating a distribution of infiltration parameters on a runoff surface. Each strip contains an equally likely value of K_s.

have dramatic effects on runoff and infiltration that cannot be ignored at larger scales. On the other hand, such interactions need not be treated by simulation at the microscale. Rather, it appears that a statistical model relating extent of soil covered with mean depth of surface flow can suffice to model many of the interactions that are important. These interactions concern the loss of runoff water during recession and the successful travel of runoff to the stream after rainfall excess has turned negative. In other cases, there may be a correlation of infiltration characteristics with local micro-elevation. Examples of this include the shortgrass rangeland microtopography, composed of hillocks of grass clumps interspersed with crusty bare areas, or the higher infiltration rates under rangeland shrubs. Modeling an interaction between water flow depth and infiltration rate is rarely undertaken but is feasible in current models (e.g., KINEROS, Smith et al. 1995).

Macropore Flow Models: Distinct cracks or channels through the soil which distinguish a real soil from an ideal porous media have received considerable attention lately, and have collectively become known as macropores. This topic is covered in a separate paper in this volume.

Rainfall Heterogeneity

Variations in rainfall intensity at the local scale can have a significant effect on infiltration heterogeneity and should not be ignored. Faurès et al. (1995) and Goodrich et al. (1995) observed rainfall gradients up to 2.5 mm/100m within a 4.4 ha watershed. This spatial rainfall variation resulted in modeled peak runoff rates which varied by a factor of almost three (8 to 23 mm/hr) when two different recording rain gauges in the proximity were used independently with the uniform spatial rainfall assumption. While this issue is not one of infiltration modeling, it is important not to forget the role that such variability plays in our treatment of heterogeneities. It is also important to remember that for modeling a balance is necessary between the treatment of process complexity and the data (rainfall or soil) availability.

Research Challenges

There are several significant areas of ignorance that should be addressed before a robust model of soil infiltration can be formulated to deal with spatial and temporal variability. Some of these areas are:

1. Much remains to be learned before a model for the statistical character of soil areal heterogeneity can be used with confidence across major soil types. Ultimately, some measure of inherent randomness and spatial scaling, such as correlation length and coefficient of variability of major soil hydraulic characteristics, should be part of our description of a soil, just as we now classify soils (albeit qualitatively) in terms of drainability and texture.
2. Probably the largest area of uncertainty in infiltration modeling is the changes that the infiltration function undergoes as a result of mechanical modifications, as indicated above. While some progress has been made in modeling the changes that a swelling, cracking clay undergoes with time, we have very little confidence in our ability to predict the formation of surface crust for a give soil texture, and our ability to anticipate the change in infiltrability of a soil at a given state caused by a given mechanical treatment.
3. There is progress being made in modeling the aggregate behavior of an area containing internal infiltration variability, but there remain significant challenges in their application in "management" models, and in understanding the conditions under which a variety of possible simplifications are acceptable. Given the preponderance of daily rainfall data, much needs to be done to improve our knowledge of disaggregation statistics and rainfall intensity distributions so that infiltration models can be used to improve the simulation of daily runoff: a physically and statistically sound lumping, rather than empirical lumping.
4. At larger scales (e.g. 10 ha +), modeling areally variable infiltration should not be done independently of the surface runoff, itself with considerable organized and random heterogeneity, nor should it be modeled without consideration of small-scale rainfall rate heterogeneities. One promising approach for larger areas might be a joint statistical/ deterministic representation of the probability of local rainfall exceeding local infiltration capacity, integrated over the area.

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Infiltration - Scale Interactions

D.C. Goodrich[1], R.E. Smith[2], D.D. Bosch[3], J.J. Stone[1], G. Paige[1], J.R. Simanton[1], W.E. Emmerich[1], T.O. Keefer[1], R.A. Shillito[1], and C.L. Unkrich[1]

It is crucial to understand infiltration phenomena over range of spatial scales to fully assess its impacts on both soil-plant-water interactions at the small scale and runoff and erosion at large scales. In particular, the field and small watershed scale is important as this is a typical area subject to management as well as the scale at which observed hydrographs and associated runoff water quality samples are available. These measurements are often our only realistic means to evaluate the impacts of management decisions on hydrologic response, and therefore indirectly, on infiltration behavior. Therefore a coupled understanding of runoff and infiltration processes is required. Two primary issues in obtaining accurate estimates of infiltration and runoff over a range of scales will be discussed. The first involves a very preliminary assessment of the comparability of infiltration estimates obtained from different techniques applied over a wide range of spatial scales. The second discusses the necessity of accurate temporal and spatial estimates of rainfall to obtain infiltration estimates at both small and large scales.

Infiltration Estimates Over a Range of Spatial Scales

A consistent set of economically feasible measurement or modeling procedures does not exist for obtaining accurate estimates of infiltration over a range of scales. At the small scale direct measures of infiltration fluxes can be obtained rapidly with disk permeameters. To obtain large scale infiltration estimates using this procedure, the daunting challenge of heterogeneity and the requisite large number of measurements must be faced. At larger scales, such as rainfall simulator plots or small watersheds, direct measures of infiltration flux cannot be obtained but must be calculated as a residual from measured rainfall less measured runoff. For rainfall simulators, this approach requires more elaborate instrumentation, and for small watershed, long-term monitoring of natural events. If a relatively small number of rapid disk permeameter measurements could provide comparable infiltration estimates to those obtained from the large scale residual approaches, a cost effective method would be available to provide large area estimates that could be used as initial infiltration parameters for simulation models. To assess the feasibility of this approach, steady-state infiltration estimates from unponded disk permeameter measurements, rainfall simulator measurements, and those estimated from a calibrated and verified rainfall-runoff model were obtained at the Lucky Hills catchments within the USDA-ARS Walnut Gulch Experimental Watershed in southeastern Arizona. Table 1 contains a summary of these preliminary estimates and the areas over which they apply. It is interesting to note the relative consistency over areas spanning seven orders of magnitude. For these watersheds, these results

[1]USDA-ARS, SWRC, 2000 E. Allen Rd.,Tucson, AZ 85719

[2]USDA-ARS, WMR, Ft. Collins, CO, 80523

[3]USDA-ARS, SEWR Lab., Tifton, GA, 31793

indicate that several disk permeameter measurements could provide the initial infiltration parameter estimates required to apply runoff models.

Table 1. Steady-State Infiltration Estimates for Lucky Hills from Three Different Methods

Method	Area (sq. m)	Steady-State Infil. Est. (mm/hr)
Unponded Disk Permeameter[1]	0.033	8.1
Rainfall Simulator[2]	30	8.1
KINEROS R-R Model[3]		
Catchment: LH-106	3,600	10.9
LH-102	14,000	8.1
LH-104	44,000	10.9

[1] Average of 2 measurements, -5 cm pressure head
[2] Average of 2 simulator runs (Green-Ampt Ks optimized using the IRIS-KW program)
[3] Estimates are based on area weighted averages from multiple distributed model elements optimized over 10 calibration events (Smith-Parlange Ks)

The Importance of the Rainfall Boundary Condition for Infiltration Estimation

The importance of temporal rainfall variability for infiltration is well known and explicitly treated in most infiltration models. The importance of spatial variability at large scales is also well known. However, recent findings in the thunderstorm dominated Walnut Gulch environment indicate that spatial rainfall variability is significant even at the 50-100m scale where the spatially uniform rainfall assumption (single rain gauge) is commonly used. Faures et al. (1995) observed rainfall gradients up to 2.5 mm/100m within a 4.4 ha watershed. This spatial rainfall variation resulted in modeled peak runoff rates which varied by a factor of almost three (8 to 23 mm/hr) when two different recording rain gauges in the basin were used independently with the spatially uniform rainfall assumption. The dominance of rainfall variability was also pointed out by Goodrich et al. (1993). In this study it was found that far greater impacts on runoff model performance resulted from using one versus two rain gauges (spatially uniform rainfall assumption) than by simplifying the entire watershed geometry (and soils variability) from over 200 modeling elements to 1 element with uniform soil/infiltration parameters. The point is that for environments dominated by infiltration excess we must be able to properly define rainfall inputs in time and space. If this critical boundary condition cannot be specified, sophisticated small-scale analysis and treatment of spatial soil variability impacting infiltration variability may not be warranted.

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Macropore Modeling: State of the Science

S. D. Logsdon[1], M. H. Nachabe[2], and L. R. Ahuja[2]

Abstract

Mechanistic infiltration models exist, but flow equations appropriate for the soil matrix do not apply for the macropore domain. This review discusses several macropore flow models. Some of the models emphasize macropore geometry whereas other models use series of equations for the macropore domain. The models also need to include analysis of the exchange between the macropore and matrix domains. The recommendations are: 1) select a model (or models) with the minimum degree of complexity required for a specific situation, to reduce the number of unmeasurable parameters; 2) complex models should provide a range of values for those parameters that cannot be easily measured; 3) sensitivity analyses are essential to determine which parameters need to be measured or carefully estimated; and 4) comparative studies are needed to evaluate these models under a variety of field conditions.

Introduction

Macropores allow rapid water flow through the soil, increasing infiltration rate and decreasing runoff and erosion. Agro-chemicals transported by infiltrating water may move through a soil layer via macropore with little altered concentration. Traditional infiltration and flow equations like Richards' equation, or Green and Ampt and Philip infiltration models, usually apply to flow through the soil matrix, but do <u>not</u> apply well for infiltration in structured soils with macropores. The flow system in macroporous soils should be divided into at least two domains, the macropore flow domain, and the soil matrix (e.g., Nachabe 1995). The traditional infiltration equations can still be used for flow in the matrix domain, but additional equations are needed to describe (i) channeling of flow through macropores, and (ii) lateral flow between macropore/soil matrix. In this article, we summarize how macropore flow is described in existing models. The objective is to emphasize the concepts behind the models, and discuss some of their features and limitations. More details can be obtained from the literature cited.

The models discussed are primarily water drainage and solute leaching models. Only two (RZWQM and MACRO) contain a runoff term, but no runon component. The importance of macropore flow in runoff prediction depends on rainfall rate relative to saturated hydraulic conductivity, but macropore flow is very critical for correct prediction of solute transport in structured soils. The solute leaching components of these models will not be discussed in this review.

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[1]USDA-ARS, NSTL, 2150 Pammel Dr., Ames, IA 5001

[2]USDA-ARS, GPSR, P.O. Box E, Ft. Collins, CO 80522

List of Abbreviations:

a = dimensionless exponent
b = source/sink term between macropore and matrix region
c = kinematic wave velocity
D_w = water diffusivity
g = acceleration due to gravity
h_b = boundary water pressure head between regions
h_{mi} = actual pressure head in micropore region
$h_{\theta\alpha}$ = function parameter
H_c = capillary drive
I_{ma} = infiltration rate in macropores
K_{ma} = hydraulic conductivity in macropore region
$K_{s(ma)}$ = saturated hydraulic conductivity in macropore region
K_s = saturated hydraulic conductivity
K_a = effective hydraulic conductivity
$K_{s(min)}$ = minimum hydraulic conductivity
K_b = boundary hydraulic conductivity between regions
$K_{r\alpha}$ = maximum K for region
$K_{k\alpha}$ = minimum K for region
L = layer thickness
n^* = empirical exponent
m^* = empirical exponent
P_{ma} = macropore fraction of soil volume
P_s = minimum volume of macroporosity
q = flux
r_p = mean radius of cylindrical macropores
r_{wf} = wetting front radius
s = slope of shrinkage characteristic
t = time
t_{cum} = cumulative time
V_{cr} = volume of crack
V_r = radial absorption
V_l = absorption from cracks
w = mean width of crack
α_{ma} = macropore drainage parameter
β = dimensionless geometry coefficient
γ_α = function parameter
γ_w = scaling factor
θ = water content
θ_{ma} = water content fraction in macropore region
θ_b = boundary water content between regions
θ_{mi} = actual water content in micropore region
θ_i = initial water content
θ_s = saturated water content
ξ_α = function parameter
κ = function parameter
ρ_w = density of water
υ = dynamic viscosity of water

Models for Macropore Flow

RZWQM

The RZWQM (Root Zone Water Quality Model; Ahuja and Hebson 1992) considers heterogeneity in the soil matrix, i.e., soil layering, and macropores. Infiltration into the soil matrix is simulated with the Green and Ampt equation using a harmonic average of conductivity of the wetted depth of the soil profile. The top soil horizons are assumed to contain vertical cylindrical macropores, and the bottom horizons contain planar cracks. Infiltration into macropores is simulated with Poiseuille's law, assuming unit hydraulic gradient (gravitational flow). Thus the infiltration rate for cylindrical macropores is

$$I_{ma} = P_{ma}\, \rho g \pi r_p^{\,4}/8\upsilon \tag{1}$$

and for planar cracks is

$$I_{ma} = P_{ma}\, \rho g w^2/12\upsilon \tag{2}$$

Macropores can either extend to the water table or dead end at a specified depth. Variation of macroporosity with depth is a useful feature in RZWQM because field experience suggests that soil macroporosity often changes with depth and time due to management effects (Starr et al. 1996 in the proceedings of this workshop). RZWQM considers that water can flow into macropores only at the soil surface after surface ponding. The ponded water at the surface is channeled through macropores and cracks and can be adsorbed slowly by lateral flow into the soil matrix.

The RZWQM, including its macropore flow component, is currently undergoing extensive evaluation at several locations around the country. Most importantly, the GPSR (Great Plains Systems Research) unit is cooperating with researchers in Nebraska, Ohio, Iowa, Minnesota, and Missouri to evaluate RZWQM for use in integrating experimental information and technology (Johnsen et al. 1993). Preliminary evaluation and testing of the macropore component of the RZWQM was conducted by Ahuja et al. (1993) and Ahuja et al. (1995). Ahuja et al. (1993) compared the performance of the model for several macropore radii, dis-continuous macropores with depth, different initial soil moisture, and several storm durations and intensity. The model provides new insights into the macropore flow behavior which are difficult to obtain in field experiments. In their evaluation of RZWQM, Ahuja et al. (1995) suggested that (i) the viscous resistance which reduces Green and Ampt infiltration rate varies between 2 and 3 depending on wetting history, (ii) lateral absorption at the macropore wall needs to be adjusted for compaction and air entrapment, (iii) the model should mix macropore flow with soil solution within $\approx$0.5 mm of the macropore wall. The model was recently modified to include this mixing with solution in the matrix. Also, the GPSR hopes to release a new version that models flow into buried macropores below a soil layer.

In the original version of RZWQM (Ahuja and Hebson 1992), shrinking and swelling effects were not included; however, Hua (1995) modified the RZWQM to simulate shrinking/swelling as affected by change in water content. Based on field measurements, Hua (1995) recommended a second order polynomial to relate the volume of crack to θ. Also, the new version by Hua (1995) allows varying the macropore/matrix transfer relationships for soil layers. This appears to be an improvement because there is strong relationship between swelling-shrinking and clay type, plant root distribution, and salinity in a soil horizon (Flerchinger et al. 1996 in the proceedings of this workshop).

MACRO

MACRO (A Model of Water Movement and Solute Movement through Macroporous Soils; Jarvis 1991a, 1994) is a two flow domain (macropore/matrix) water flow model that includes shrink-swell influence on macropore volume and crop water uptake. The water retention in the matrix region, $h(\theta)$, is described by the Brooks and Corey (1964) equation, and the conductivity, $K(\theta)$, by Mualem's (1976) equation. Instead of saturated θ_s and K_s, the matrix region's upper limit is a boundary water content θ_b and conductivity K_b. The difference is the minimum macroporosity of the soil, typically a small value (e.g. Nachabe 1995). The hydraulic conductivity for the macropore region is:

$$K_{ma} = K_{s\,(ma)}\,(\theta_{ma}/P_{ma})^{n^*} \tag{3}$$

It is assumed that water cannot flow into the macropore region until the matrix region is saturated. Like RZWQM, water can flow into macropore region at the soil surface as long as the matrix at the surface is saturated (even if the subsurface matrix is not saturated). Water that enters the macropore region at the soil surface can move laterally and vertically into subsurface matrix pores.

Shrink-swell changes are accounted for in the macropore region by the following equations:

$$P_{ma} = P_s + s(\theta_b - \theta_{mi}) \tag{4}$$

$$K_{s(ma)} = (K_{s(min)} - K_b)(P_{ma}/P_s)^{m^*} \tag{5}$$

It is assumed that total soil volume does not change, but that changes in microporosity with shrinking/swelling result in changes in macroporosity.

Experience using this model has produced spurious results unless h_b is -30 mm or larger (Logsdon, 1996, not published). Also MACRO has been evaluated by Jarvis et al. (1991a, b, 1994) and Jabro et al. (1994). Jarvis et al. (1991b) concluded that the two domain-approach in MACRO reproduced better than the one-domain approach the water discharge and chemical breakthrough in a soil monolith. On the other hand, Jabro et al. (1994) used field data to compare and evaluate the predictions of early version of MACRO and SLIM (SLIM is another preferential flow model described later in this article). They found that SLIM predicted leaching successfully whereas MACRO required calibration for successful in prediction. Using statistical criteria for comparison, Jabro et al. (1994) concluded that SLIM provided slightly better predictions than

MACRO

Kinematic Wave Model of Macropore Flow

Germann (1995) and Germann and Beven (1985) proposed the kinematic wave approximation to model flow into macropores. The governing equation for macropore flow including a sink function for water sorbance into the soil matrix is

$$\partial\theta_{ma}/\partial t + c\ \partial q/\partial z + c\ b\ \theta_{ma} = 0 \tag{6}$$

The flux through macropores, q, is related to θ_{ma} by the relation (Beven and Germann 1981)

$$q = K_{ma}\ \theta_{ma}{}^{a} \tag{7}$$

Germann and Beven (1985) used these three equations to introduce a macropore flow model for variable macropore input. The parameters of this model, a, b, and K_{ma} need to be estimated a priori from experience or calibrated with field data. Germann and Beven (1985) found that a and K_{ma} are more sensitive parameters than b. In evaluating this model, Germann and Beven (1995) coupled the kinematic wave approximation for macropore with Philip's two term infiltration model for the soil matrix. They showed a fairly good agreement between the modeled and observed soil water hydrographs for 19 Dutch soils.

MURF

MURF (MUltiple-Region-Flow; Gwo et al. 1995) is a three-flow domain water flow model in two dimensions that does not include shrink-swell or plant water uptake. The three regions are macropore, mesopore, and micropore. For the mesopore and micropore region, h(θ) is described by the van Genuchten (1980) equation. The macropore region is described by a Fermi function as follows:

$$\theta_{\alpha} = (\theta_{s\alpha}-\theta_{sr})/\{1 + \exp[-\gamma_{\alpha}(h_{\alpha} - h_{\theta\alpha})]\} + \theta_{sr} \tag{8}$$

For each of the regions a relative K can be determined as follows:

$$\log_{10}(K_{r\alpha}/K_{s\alpha}) = \xi/\{1 + \exp[-\kappa(h_{\alpha} - h_{k\alpha})]\} - \xi_{\alpha} \tag{9}$$

Mass transfer coefficients need to be given for flow between any two of the three regions. For unsaturated conditions, it is possible for all regions to be partially unsaturated at the same time.

Tipping Bucket Model of Macropore Flow

Emermann (1995) describes a tipping bucket model for macropore flow. Micropore flow is given by Richards equation where macropore water is considered to be at air-entry pressure (so micropore flow is independent of macropore flow). Brooks and Corey equations were used to describe water retention and hydraulic conductivity in the matrix domain. It is assumed that precipitation is divided between micropore and macropore regions according to their relative proportions in the soil. All water in excess of micropore infiltration is channeled to the macropores. It is also assumed that the total porosity in the macropore region is equal to the total porosity in the micropore region. Macropore flow is assumed to be only due to gravity, and the macropore domain is characterized by P_{ma} and α_{ma}. Drainage from the macropore domain for layers of varying thickness was derived starting with the simple relationship:

$$d\theta_{ma}/dt = -\alpha_{ma}\theta_{ma} \tag{10}$$

Emermann (1995) suggested that the tipping bucket model is a particular case of the kinematic wave model for exponent a=1 (see section on kinematic water model, Eq. 7), and constant thickness. Sensitivity analysis using measured water content data in Brazilian Oxisols showed that P_{ma} is the most sensitive and important parameter for the tipping bucket model.

Other Macropore flow Models

SLIM (Solute Leaching Intermediate Model) developed by Addiscott et al. (1986) and updated by Addiscott and Whitmore (1991) is a simple two-domain flow model. The model divides the profile into layers, and the soil water within each layer is partitioned into mobile and immobile phases. The boundary between mobile and immobile soil water is assumed at the potential of -0.03 Mpa. The two main parameters of the model are a measure of the capacity of the soil to hold immobile water and a permeability parameter. These parameters can be estimated from water retention and hydraulic conductivity data. SLIM does not account for runoff.

Beven and Clarke (1986) introduced a model of infiltration into a population of macropores. They developed Green and Ampt type of solutions for a single macropore-soil matrix flow. These solutions were then integrated for a population of macropores assuming a Poisson distribution of macropore radii and depth.

Steenhuis and Parlange (1988) and Steenhuis et al. (1990) describe a "piecemeal linear K function" which could be called a multi-flow domain approach. Each domain has a separate conductivity.

Chen and Wagenet (1992a,b) described the macropore, matrix, and exchange regions. Water flow in the macropore region is assumed to be turbulent flow. Campbell's equation (Campbell, 1974) is used to describe K of the matrix region.

FLOCR (FLOw in CRacking soils; Oostindie and Bronswick, 1992) is another one dimensional model that includes shrinking-swelling effects of clay on infiltration and drainage in an unsaturated soil. The model can also be used to calculate subsidence of clay soils. FLOCR does not simulate chemical and biological processes in the root zone.

Hosang (1993) described a two-domain approach with Richards' equation applying to both the macropore and matrix regions. Preferential flow does not occur unless the application rate is greater than matrix K. Water which moves through the macropores is absorbed into the matrix at the base of the channel. He also assumes van Genuchten (1980) equations of $h(\theta)$ and $K(h)$ for both the macropore and matrix regions. He assumes that preferential flow is immediately started and stopped by intense rain events.

Gerke and van Genuchten (1993a) introduced a two-flow domain model which adopts a van Genchten (1980) characterization of $h(\theta)$ and $K(h)$ for both the macropore and matrix domain. Richards' equations for the two domains are coupled through a water transfer term and solved simultaneously for the flow system.

Water Transfer Between Macropore/Matrix

One of the critical components in most dual-porosity models is the source/sink term describing the exchange of water between domains. The challenge is to capture these local-scale processes and incorporate them into a macroscopic dual-porosity model. Several assumptions have been introduced to simplify the modeling of water transfer. For example, if the transfer of water between domains is assumed fast and there is instantaneous equilibrium between pressures, then the governing equations for flow between domains can be combined (Gerke and van Genuchten 1993b; Dykhuizen 1990). Water transfer between domains can be modeled with numerical techniques (Hoogmoed and Bouma 1980), Green and Ampt approximations (Ahuja and Hebson 1992; Beven and Clarke 1985), or Philip's sorptivity solution (Chen and Wagenet 1992a). In addition, Darcian type water transfer terms based on water pressure gradient were used in models by Workman and Skaggs (1990) and Othmer et al. (1991). Gerke and van Genuchten (1993b), on the other hand, proposed a first-order mass transfer term of the form:

$$b = (\beta/d^2)\, K_a\, \gamma_w\, (h_b - h_{mi}) \tag{11}$$

Equations for the mass transfer coefficients can be determined for different geometries. For example, in the case of parallel rectangular slabs $\beta = 3$.

Both MACRO and MURF are based on parallel rectangular slab geometry with inter-domain affected directly or indirectly by slab width. In MACRO flow between regions is very similar to that by Gerke and van Genuchten (1993b), but is linearly related to the difference in water content:

$$b = [(3D_w\gamma_w)/d^2](\theta_b - \theta_{mi}) \tag{12}$$

The macropore sorbance term of the kinematic wave model is assumed of the form:

$$b = -1/\theta_{ma}\, d\theta_{ma}/dt \tag{13}$$

In RZWQM the radial absorption rate from radial macropores is calculated with a Green and Ampt type of equation given as

$$V_r = 2\pi K_s H_c \,/\, \ln(r_{wf}/r_p) \tag{14}$$

The absorption rate, V_l, per unit width of cracks is:

$$V_l = [2H_cK_s(\theta_s - \theta_i)/t_{cum}]^{1/2}. \tag{15}$$

Recommendations and Future Research

Simplicity vs. Flexibility

The macropore system is complex (geometry, depth, distribution, connectivity, etc.). The first type of models attempt to describe the macropore system by simplified geometry. Examples are the RZWQM and the model of Beven and Clarke (1986). The second type of models rely on separate flow equations for the two or multi-domains without specifically describing the geometry. Examples of this second type of models are MACRO, MURF, tipping bucket model, and Gerke and van Genuchten (1993). Whether or not the geometry is specified for macropore flow, geometry is usually included in the exchange term between domains. Advantages and drawbacks exist in both types of models.

There are limitations to adequate descriptions of macropore and fracture geometry, especially for irregularly shaped pores and for continuity and connectivity of pores. We do not completely understand the conversion of geometry to flow, i.e., laminar vs. turbulent, interconnected flow, etc.

It may also be difficult to obtain parameters needed for flow equations in multi-regions type of models. We can obtain K for saturated condition and negative head (-3 cm cutoff for macropores according to Luxmoore, 1981), but two points do not describe the shape of the curve. Assumptions of unit gradient (Jarvis, 1994; Emermann, 1995) in the macropore region may not be valid.

Simplified models may have greater utility for field studies because they use easy to measure parameters and minimize the reliance on unmeasured parameters. Calibrated unmeasured parameters of complex models may not have physical significance. Complex models may still be useful as research tools because they are more flexible, and they may help us understand the relative importance of various factors for infiltration and solute leaching. Models should provide a range of values for these parameters that cannot be easily measured, and should minimize the number of unmeasured parameters.

Equations for noninteraction, cylindrical pores are not applicable to fractures and irregular macropores usually present in soil due to different capillary rise for noncylindrical pore shape, pore interaction, tortuosity, pore necks, turbulent flow, and nonfuctionality of some pores. Simple flow equations should be used to describe the macropore flow system rather than the pore geometry. The flow system is easily measured with disk permeameters, but the geometry description may require resin impregnation along with image analysis. Traditional flow equations developed for soil matrix usually apply to homogeneous sand, but do not apply well to structured, fine-textured soil within the macropore range. Because the macropore flow system is not well understood, complicated equations with many parameters are not very useful descriptions.

Comparative Studies

Given the abundance of macropore flow models, we need comparative studies to evaluate these models under a variety of field conditions. Many of the models discussed above have limitations in relation to macropore geometry, depth, and distribution. Comparative studies will allow the user to choose a model that is most appropriate for a field situation.

In summary the recommendations are: 1) select a model (or models) with the minimum degree of complexity required for a specific situation, to reduce the number of unmeasurable parameters; 2) complex models should provide a range of values for those parameters that cannot be easily measured; 3) sensitivity analyses are essential to determine which parameters need to be measured or carefully estimated; and 4) comparative studies are needed to evaluate these models under a variety of field conditions.

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Modeling Infiltration Into Frozen Soils

G.N. Flerchinger[1]

Abstract

The occurrence of frozen soil can result in significant runoff and erosion events from otherwise mild rainfall or snowmelt events. However most hydrologic models, including most snowmelt runoff models, include no provisions for soil freezing and thawing, and thus cannot address these extreme, yet common hydrologic events. A wide array of approaches exist among different models for compensating for the effects of frozen soil on infiltration. The purpose of this paper is to present the state-of-the-science in modeling frozen soil infiltration and to propose future research directions to improve our prediction of infiltration into frozen soils. Future research needs include: better quantification of the interrelation between water content, ice content, and texture; better quantification of the effects that these have on infiltration; and development of a methodology for accounting for which pores are filled with ice and which are available for infiltration.

Introduction

Seasonally frozen soil plays an important role in the hydrology in northern latitudes. In many areas, rain or snowmelt on seasonally frozen soil is the single leading cause of severe runoff events, but most hydrologic models do not address the effects of soil freezing on infiltration. Efforts to predict frozen soil infiltration and runoff have had limited success, which is reflective of the current knowledge of frozen soil infiltration processes. Frozen soil processes lag considerably behind non-frozen processes due partly to the difficulty in quantifying and measuring water and ice conditions in frozen soil. An array of modeling approaches for describing infiltration into frozen soils is presented, and future directions for modeling infiltration into frozen soil is discussed.

State-of-the-Science

Most approaches for estimating infiltration into frozen soil make use of some adjustment to the saturated hydraulic conductivity or curve number when the soil is frozen. However, models differ considerably in the theory and sophistication used to determine whether the soil is frozen and to adjust infiltration rates for frozen conditions. Adjustments for frozen conditions may be based on: simply whether the soil is below freezing temperatures; the amount of ice present in the soil; or the available porosity remaining in the frozen soil.

Very simple approaches use essentially a simple on/off switch for accounting for frozen soil effects. Models which use this approach typically use empirical methods based on air temperature to determine whether the soil is frozen, and if so, adjust infiltration parameters regardless of antecedent water content. Such an approach has been incorporated into the

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[1]USDA-ARS, NWRC, 800 Park Blvd., Ste 105, Boise, ID 83712

CREAMS and GLEAMS models (Knisel et al., 1985). In these models, mean monthly temperatures and radiation are used to estimate the occurrence and duration of frozen soil conditions. Daily temperature estimated by fitting a Fourier series to mean monthly data are used to determine when frozen soil conditions are initiated. The number of days with frozen soil after mean daily temperature rises above freezing, *DFS*, is computed from:

$$DFS = 35.4 - 0.154\ R \tag{1}$$

where R the December solar radiation in gm cal cm^{-2} d^{-1}. When the soil is frozen, the base Curve Number is set to 95 or 98 depending on whether the unfrozen base Curve Number is less than or greater than 80. Unfortunately, the model has been unable to simulate runoff for short-term melt events because mean monthly temperature data does not provide for the mid-winter warm days that are responsible for many major runoff events. Jamieson and Clausen (1988) reported that CREAMS underestimated runoff during high flow months and suggested that monthly input data does not represent actual field conditions well enough to allow accurate predictions of snowmelt and runoff.

A slight improvement over the CREAMS/GLEAMS approach was presented by Gray et al. (1985). This approach categorizes snowmelt into frozen soil as follows: (1) restricted where the soil is assumed impervious; (2) unlimited where the soil is capable of infiltrating the snowcover water equivalent; and (3) limited where infiltration is governed by the snowcover water equivalent and the ice content of the soil at the time of melt. For the above classification system, infiltration for the "limited" category was computed from:

$$I = 5(1 - \theta_p)\ SWE^{0.584} \tag{2}$$

where I is infiltration capacity (mm), θ_p is degree of pore saturation, and SWE is snowcover water equivalent (mm).

More sophisticated models estimate soil frost depth using bulk heat transfer coefficients and assuming linear temperature gradients through snow, frozen soil, residue and unfrozen soil layers (Vehviläinen and Motovilov, 1989; and Savabi et al., 1995). Surface temperature in these models are typically estimated from air temperature or a simple surface energy balance, and all water within the frozen soil layer is assumed to be ice. The hydraulic conductivity adjustment factor for frozen soil in the WEPP (Water Erosion Prediction Project) model is computed from

$$FS_a = 3.75\ e^{(-0.26 F_\theta)} \tag{3}$$

where frozen soil hydraulic conductivity is estimated by multiplying saturated hydraulic conductivity by FS_a. F_θ is predicted from:

$$F_\theta = \frac{\theta_i}{\theta_{fc}} 100 \tag{4}$$

where θ_i is the volumetric ice content (taken as the soil water content at freezing) and θ_{fc} is the volumetric field capacity (Alberts et al., 1995). A similar adjustment factor is computed by Vehviläinen and Motovilov (1989).

Models by Flerchinger and Saxton (1989), Lundin (1990) and Grant (1992) take a detailed approach by simulating the coupled heat and water transfer in the frozen soil system. These models use finite difference or finite element methods to solve the heat and water flux equations within the soil. Liquid water content is computed from soil water potential, which is related to temperature by the Clapeyron equation. These models typically assume the hydraulic conductivity and water retention characteristics are the same for frozen and unfrozen soils. Thus, hydraulic conductivity for infiltration is based on the unsaturated hydraulic conductivity computed from the available porosity. In the SHAW model (Flerchinger and Saxton, 1989), which uses the Brooks-Corey type approach for unsaturated relations, effective hydraulic conductivity for use in the Green-Ampt infiltration equation is computed from:

$$K_e = K_s \left(\frac{\theta_w}{\theta_s} \right)^{2b+3} \tag{5}$$

where K_s is saturated conductivity, θ_w is the available porosity for infiltration (total porosity less ice content), θ_s is saturated water content of the unfrozen soil, and b is the Brooks-Corey pore-size distribution parameter.

Future Directions

Before significant advances in modeling infiltration into frozen soils can come about, a better understanding and quantification of water and ice conditions in the soil needs to be developed and related to the impacts on infiltration of water into the soil. Better quantification of the impact of the interrelation between water, ice content and texture on infiltration would be helpful in both the simpler and more complex models.

Models currently have no accounting of which pores are filled with ice and which are available for infiltration. This is a potential area for improvement in the more sophisticated models. The size of the pores filled with ice depends on the water content and water potential of the soil at the time of freezing and can greatly affect the infiltration capacity of the soil. Approaches to track the range of pore sizes filled with ice have been developed, but these need to be incorporated into hydrologic models.

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Poster Abstracts

Spatial Variability of Runoff From Small Agricultural Watersheds On Similar Soil-Map Units

James V. Bonta[1]

Introduction

Soil hydraulic property and other characterization information are important for watershed modeling of runoff, water quality, and erosion. The source of soils information is often soil-survey maps. Runoff data from two small, adjacent experimental agricultural watersheds at the USDA-Agricultural Research Service experimental watershed facility near Coshocton, Ohio, that are visually similar, on opposite sides of an isolated hilltop, and share an upper boundary, show that annual runoff can be significantly higher on one watershed compared with the other (Fig. 1). However, the soil map available to a practitioner shows the soils for both watersheds to be mapped in the same soil series (Fig. 2).

Summary

The experimental watersheds are located on an isolated hilltop on which a clay layer outcrops. Available precipitation, runoff, ground-water, soil moisture, and soil characterization data were analyzed and suggest that the soil characteristics and moisture differences in areas upstream from each watershed outlet are responsible for at least some of the difference in watershed response to precipitation. Available clay-configuration data suggests that a geologic clay layer in the higher-yielding watershed may be collecting water from outside the watershed. The results suggest that spatial variability of soil information within soil map units and geologic information would be helpful to adequately model watershed runoff, provided a suitable model that used these types of inputs was used. This information would enable persons using GIS data bases to include this variability and sources of variability for watershed model inputs.

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[1]USDA - ARS, North Appalachian Experimental Watershed, P. O. Box 488, Coshocton, OH 43812

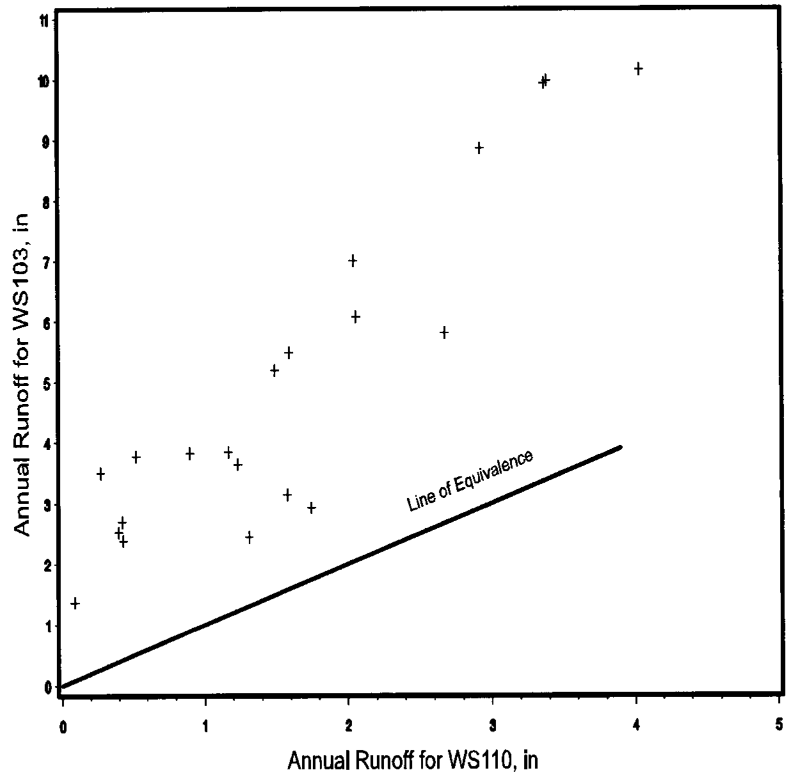

Fig. 1. Comparison of total annual runoff for Watersheds WS103 and WS110.

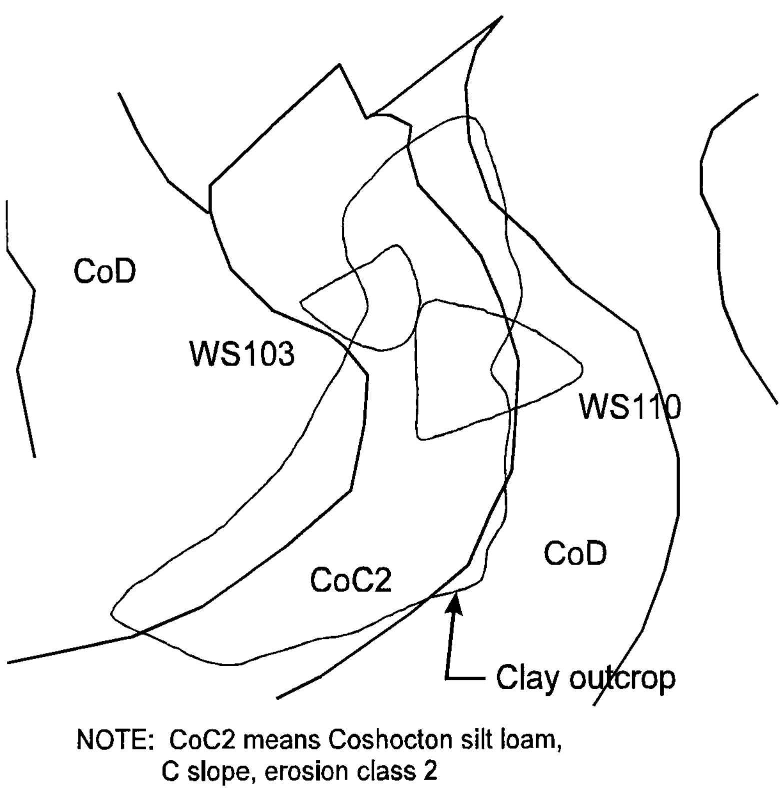

Fig. 2. Generally available NRCS soil map for WS103 and WS110.

"Real World" Precipitation

James V. Bonta[1] and Virginia Ferreira[2]

Introduction

Models that simulate natural hydrologic processes are needed for evaluating the effects of agricultural land-management practices on runoff, erosion, and water quality at a variety of areal scales. The infiltration process is affected by antecedent moisture conditions (dry times between storms) and rainfall intensity variations during a storm. Newer models utilize infiltration algorithms that require information on statistically-correct storm occurrence and short-time increment precipitation that mimic these critical conditions. However, short-time-increment data are generally not available and, therefore, must be synthesized. Various types of design storms that have been used in the past for engineering design capture neither the information needed for quantification of antecedent moisture nor the variability of intensities observed in natural rainfall.

Summary

The poster presents a summary of investigations that have been made into various aspects of storm generation using Huff curves. Huff curves are a probablistic representation of within-storm intensity variations. They have the potential to be used for stochastic simulation of within-storm intensities. Huff curves are developed from a storm-identification technique that utilizes the exponential distribution between independent storms, yielding a critical duration. The critical duration is the maximum dry-period duration that, on average, separates storms. The resulting storms are nondimensionalized and superimposed. From this graph, isopleths of probability are developed. The results from past studies show that there is essentially no effect of sampling interval of data on Huff curves. However, season, number of storms analyzed, and storm size are factors that must be considered in developing Huff curves. Regionalization of curves has been successfully demonstrated for distances of about 400 miles.

A preliminary investigation of regionalization of critical duration showed that the logarithm of critical duration could be mapped for the Colorado, Wyoming, Nebraska, and Kansas areas. The resulting patterns of critical duration followed the general patterns of monthly precipitation, an important linkage that is applicable to the storm-generation project.

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[1]USDA - ARS, North Appalachian Experimental Watershed, P. O. Box 488, Coshocton, OH 43812

[2]USDA - ARS, Great Plains System Research, 301 S. Howes, P. O. Box E, Ft. Collins, CO 80522

Subsurface Lateral Transport in Glacial Till Soils During a Wet Year

S. D. Logsdon[1] and D. B. Jaynes[1]

Water and solute movement in a landscape is not all vertical, but rather laterally down a hillslope. The Des Moines lobe area has shallow (<3 m) groundwater and gently rolling topography that often results in lateral movement of water and solutes down hillslopes. The objective of this study is to measure the occurrence of lateral flow of water and solutes in Central Iowa.

On 20 May 1993 Br- tracer was applied in a 6 m long transect 0.15 m deep (disk pan) near the top of a hill in a farmer's field in Iowa. The total amount of Br- applied in the transect was 171 g applied as a 0.25 M solution. The soil on the slope is a typical well-drained Clarion (Fine-loamy, mixed, mesic Typic Hapludoll). Disking and soybean (*Glycine max*) planting were done on 9 June 1993.

During June nests of tensiometers at four depths (0.33, 0.50, 0.66, and 0.95 m center point of ceramic) were installed at four locations down the slope from the transect (1.47, 2.4, 4.8, 7.8 m). Tensiometers were manually read with carry-along transducers connected to needles inserted through the septa for measurement.

Neutron probe access tubes were installed at three locations downslope (0, 3.0, and 5.29 m) in June 1993. For calibration, soil samples were collected to 2.5 m at the same time as measurements were made with the neutron probe. After subsamples were removed for soil water content measurements, the rest of each profile were examined for soil characterization.

Shallow wells were installed at three locations downslope (1.9, 3.22, and 4.20 m) at depths of 3.56, 3.65, and 3.62 m, respectively. (Rain prevented further well installations.) Depth to the water table was measured referenced to the top of each well with a battery operated well depth indicator. The well tops were surveyed in for elevation and location.

Four times (11 June, 23 June, 21 July, and 27 October, 1993) soil samples were taken to 1.4 m for analysis of bromide. On 11 June the samples were taken 0.30, 1.40, and 2.92 m downslope. On 23 June and 21 July the samples were also taken 5.81 m downslope. After harvest on 20 October the samples were taken at 0, 1.4, 2.92, 5.81, and 14.95 m downslope. The soil samples were extracted and measured for bromide on a Dionex Series 4500i ion chromatograph (West Mont, IL). Soil water content was measured on subsamples.

After harvest a pit was dug perpendicular to and extending downslope from the transect on 20 October 1993. Undisturbed soil samples were taken in thin-walled stainless steel cores (73 mm diameter by 76 mm long) both vertically and angled with the soil surface to examine anisotropy of K_{sat} and for bulk density measurements. The cores were taken at depths of 0.15, 0.41, 0.62, and 0.77 m along the east side of the pit, and depths of 0.17, 0.32, 0.57, and 0.74 m along the west side of the pit. The cores were used for measurement of bulk density and saturated hydraulic conductivity. Additional deep core data for Clarion soils in the same or nearly fields was included for deep bulk density and saturated hydraulic conductivity data.

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[1]USDA-ARS, NSTL, 2150 Pammel Dr., Ames, IA 50011

The vertical depth of bromide peak was calculated for each position downslope by summing (all depths) the quotient of the depth by the fraction of bromide at that depth. Extrapolated percent of vertical, lateral, and surface movement was calculated by extrapolating between measured values. The lateral transport of bromide was calculated by using the maximum downslope detection of bromide for each sampling date and fitting by linear regression to get the mean lateral transport per day. Similarly regression of lateral transport as a function of rain was used to determine lateral transport per mm rain.

Cumulative rainfall amounts of 84, 172, 534, and 959 mm had fallen between application and the first, second, third, and fourth sampling dates. In this wet year, 25% of applied bromide mass moved laterally (21 July, 1993 sampling) in the subsurface extending to 6 m downslope, and minute amounts were detected near the surface extending to 15 m downslope (23 October, 1993 sampling). Over 70% moved vertically beneath the applied transect (23 October, 1993 sampling) but only to the 0.9 m depth.

The peak bromide concentrations were at 0.7 m depth underneath the application transect, and increased to 1.1 m at 5.8 m downslope. The center of bromide mass could indicate vertical movement within the application transect, then lateral movement downslope with only a small vertical component in the downslope positions. Vertical movement of bromide was not detected below the highest measured depth of the water table (0.9 m) in the transect but trace amounts (<0.5 ppm) were observed within the water table for downslope positions. By the end of the season, the water table depth was 1.8 m, and sampling after harvest showed that bromide depth was deeper than the 1.4 m of sampling, but only for downslope positions. Mean lateral tensiometer gradients were 7% or slightly more vertical than the slope. A sand layer was noted under the tracer placement at 0.72 to 0.87 m below the soil surface, but only in the middle of the transect. This corresponded to about the bottom depth of bromide detection. No sand lenses were observed downslope. Anisotropy of Ksat was not apparent from undisturbed cores taken horizontally (6-15 μm/s) and vertically (12-34 μm/s), but Ksat generally decreased with depth, and bulk density generally increased with depth. The mean lateral bromide transport distance was 93 mm/day downslope, or 15 mm/mm rain.

In summary, lateral movement of bromide occurred primarily in the unsaturated zone, and movement may have been concentrated in the capillary fringe. Crops were growing which usually resulted in an upward vertical gradient. The occurrence of lateral water and solute movement should be considered in contaminant transport.

Scaling Analysis of Infiltration at R-5 Catchment

Huan Meng[1], Jorge A. Ramirez[1], Jose D. Salas[1], Lajpat R. Ahuja[2]

Abstract

Both steady state and transient infiltration rates are studied in this study using scaling approach based on two sets of data collected from R-5 catchment in Oklahoma. The emphasis is on the spatial aspect for the former and is both on spatial and temporal aspects for the later. It is concluded that infiltration is multiscaling. Both precipitation and soil properties appear to affect the scaling properties of infiltration.

Introduction

Scaling theory has been used to study some hydrologic fields and processes in the past decade or so. But little has been done on the infiltration process in terms of its scaling properties. In this study, we focuses on the scaling analysis of both steady state infiltration (SSI) and transient infiltration rate (TI). The data we use are composed of 157 SSI measurements taken from R-5 catchment in Oklahoma (area 0.1 km^2). There are two kinds of scaling processes: simple scaling (Eq. (1)) and multiscaling (Eq. (2)).

$$\log(E[X_\lambda^n]) = n\theta \log(\lambda) + \log(E[X_1^n]) \tag{1}$$

$$\log(E[X_\lambda^n]) = \theta(n) \log(\lambda) + \log(E[X_1^n]) \tag{2}$$

where X is a random field or process; X_λ is the arithmetic mean of X at spatial scale λ; n is order of moment; θ is scaling parameter, constant for simple scaling and variable for multiscaling.

Scaling analysis of SSI

We start with SSI. The arithmetic means of SSI measurements are calculated on five consecutive aggregation levels. Up to the 6th order moments are computed for these averaged SSI at each level. The results display a log-log linearity between the moments of SSI and scale, λ (eq. (1) and (2)). SSI is thus scaling. It is also observed that the slopes of the moments are a convex function of their order - a phenomena that reveals multiscaling property.

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[1]Department of Civil Engineering, Colorado State University, Ft. Collins, CO 80523

[2]USDA-ARS, GPSR, Ft. Collins, CO 80523

Scaling analysis of TI

Next, we study the scaling properties of TI. A set of comprehensive measurements from R-5 catchment is used to derive the empirical regression relation between saturated hydraulic conductivity, K_s and capillary drive, H_c and between K_s and deficit soil moisture content, θ_d. These relations along with the 157 K_s data are then used to compute the H_c and θ_d data required in the Green-Ampt infiltration model. Four cases with steady, spatially uniform rainfall rates, R, are designed to explore the effect of soil property and precipitation on the scaling properties of TI. The first three cases have original K_s and small, medium and large rain rates, respectively. The last case corresponds to a small R and a K_s set with 10-fold smaller values than the original K_s. We compare the slopes of the moments at 20 minutes after rain starts for different R. Also compared are the different scaling properties of TI at t = 20 min. with same R but different K_s.

The following conclusions are drawn from our study. i) Steady state infiltration is multiscaling at R-5; ii) Transient infiltration is also multiscaling except in the trivial case when majority locations having K_s>R; iii) Prior to steady state conditions, infiltration becomes more and more "multiscaling" as time increases; iv) Slopes of moments become stable after a certain period of time depending on R and K_s; v) As R increases, slopes of moments decrease while intercepts increase; vi) As R increases, infiltration becomes more "multiscaling" because more locations having K_s>R; vii) The smaller the magnitude of K_s, the more "multiscaling" infiltration is for the same rainfall rate; viii) Both precipitation and soil properties affect the scaling properties of infiltration; ix) Results prove similar scaling properties for cumulative infiltration as for infiltration rate (not shown).

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Roughness Influences on Soil Acoustic Impedance

Ron W. Rickman[1], James M. Sabatier[2], and K. Attenborough[3]

Abstract

Surface roughness of agricultural fields is utilized to control both wind and water erosion. There is no method that is both simple and accurate for measuring the roughness of a soil surface. Since both roughness and porosity of a surface control the reflection of sound from it, sound pressure was measured above surfaces of known roughness and porosity in an attempt to develop a nondestructive, rapid procedure for characterizing soil surface roughness. Acoustic level difference measurements (Sabatier et al. 1990), were collected above dry fine (< 0.2 mm) and coarse (0.8 mm) textured sands with a variety of roughened surfaces. Both sands were 45 cm deep and contained in 2m x 2m wooden frames that were isolated by over 200m from adjacent structures that would reflect sound. For the fine sand a flat surface, a surface with triangular furrows 8 cm center-to-center and 2.0 cm from furrow bottom to ridge top, and a surface with triangular furrows 17 cm center-to-center and 4.5 cm from furrow bottom to ridge top were observed. For the coarse sand a flat surface and a surface with rounded furrows 8 cm center-to-center and 1.5 cm from furrow bottom to ridge top were observed. All furrows were perpendicular to the line from the speaker to the microphones, filled the 1.5 m space between them, and extended at least 0.75 m on either side of the center line of the speaker and microphones.

The theoretical pattern of the sound pressure above each surface was computed as described by Sabatier et al. 1993. Expected roughness effects were computed according to relationships derived by Howe, 1985 and provided by Attenborough,1995. The theoretical level difference curves were fitted to the observed data by selecting flow resistivity and tortuosity values that provided the best fit when measured soil porosity, roughness element size and source-receiver distances for each surface condition were utilized.

For the flat surfaces, agreement between theory and observation was excellent. For the fine sand which was quite reflective to sound, observed and computed effects of the furrows also matched quite well. In order to match the observed effects of roughness for the coarse sand, which did not reflect sound well, a lower value of flow resistivity had to be used for the coarse sand when it was furrowed than when it was flat. (Table 1).

Furrow patterns in the soil surface can be incorporated into the computed frequency response of level difference measurements. Furrow height and spacing must be measured and provided for the computations. Changes in flow resistivity of soils caused by tillage are detectable with level difference measurements. The results of the reflection computations were very

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[1]USDA Columbia Plateau Conservation Research Center. Box 370 Pendleton OR 97801

[2]Natl. Center for Physical Acoustics, Box 847, Univ. of Mississippi, University, MS 38677

[3]Dept. of Engineering Mechanics, The Open Univ., Walton Hall, Milton Keynes MK7 6AA, England.

sensitive to the distances between the microphones, speaker, and soil surfaces. Accurate measures of the distances between system components were critical. Simultaneous changes of both surface roughness and porosity could not be detected from a single observation.

Table 1. Parameter values for matching observed and computed level difference curves.

Sand texture	Surface condition	SR	U mic.	L mic.	P	T	F R	R E height	R E length	R E
		m	m	m			mks	m	m	f/m
fine	flat	1.50	0.34	0.09	0.43	3	500,000	0	-	-
fine	small furrows	1.50	0.34	0.09	0.43	3	500,000	0.025	0.07	12.5
fine	large furrows	1.50	0.34	0.09	0.43	3	500,000	0.045	0.14	7.1
coarse	flat	1.50	0.34	0.09	0.39	1.5	140,000	0	-	-
coarse	small furrows	1.50	0.34	0.09	0.39	1.5	80,000	0.01	0.07	12.5

SR - Source to Receiver distance
U mic. - Upper microphone and speaker elevation
L mic. - Lower microphone elevation
P - Porosity (volume ratio)
T - Tortuosity
F R - Flow resistivity (mks units)
R E - Roughness Element (f/m - frequency of occurrence per meter)

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"Real World" Infiltration Research at the National Sedimentation Laboratory

M. J. M. Römkens[1] and S. N. Prasad [2]

Several projects on "Real World" infiltration are being pursued at the National Sedimentation Laboratory. They consist of: (i) the effect of surface sealing on infiltration and (ii) the role of cracks on infiltration in swelling/cracking soils. In the first category, both the effect of rainstorm characteristics and the role of sealing susceptible soil properties are investigated. In these projects, the research procedures used reflect a high degree of artificiality, yet the methodology chosen is designed to single out specific and commonly occurring factors that differ from the customary, fixed matrix system that usually is assumed in many analytically based infiltration equations. The experiments indicated the need for detailed and systematic studies of soils with a changing soil matrix upon wetting or drying.

Surface Sealing In Relation To Rainfall Intensity

The effect of rainfall intensity on infiltration was studied on two soils. (i) Grenada Ap material, and (ii) Glauconitic sediment. In the glauconitic sediment, a consistent pattern of higher cumulative infiltration with higher rainfall intensities were observed at corresponding times during simulated rainstorms. On the other hand, the Grenada soil showed "cross-over" points in which the cumulative infiltration at higher rainfall intensities was less than that of storms with less intensity. The results suggest that the Grenada soil was much less stable than the glauconitic sediment under raindrop impact.

Surface Sealing In Relation To Soil Properties

The effect of liming on cumulative infiltration was studied for glauconitic sediment under simulated rainfall conditions. The data show that liming substantially reduced infiltration with a maximum reduction at about 5 mmole of $Ca(OH)_2$ per 1 kg of soil. Reduced infiltration following liming is a commonly occurring phenomenon on ultisols and oxysols and is related to the physico-chemical characteristics of these iron oxide rich soils which have a pH dependent surface charge. Liming eliminates the positive surface charges of the low pH soil which leads to a more dispersible condition that enhances surface sealing with the impact of raindrops.

Infiltration Into Swelling/Cracking Soils

Infiltration into expansive soils is appreciably affected by the degree of drying (or crack development) and initial wetness of the soil profile. For an initially packed, air dry soil sieved to pass a 2 mm screen and subjected to a series of identical rainstorms, followed by drying, infiltration and incipient ponding increased during each subsequent storm event. These findings

[1] USDA-ARS, P. O. Box 1157, Oxford, Mississippi 38655

[2] University of Mississippi, University, Mississippi 38677

were attributed to increasing crack opening following each successive storm. On the other hand, the infiltration rate during the post ponding phase decreased with each successive rainstorm, suggesting that water intake through the soil matrix itself had virtually ceased. Also, the cracking pattern in successive storms appeared to be very similar.

Appendix

Partnership with NRCS

T. M. Sobecki and R. D. Yeck[1]

Introduction

We appreciate the opportunity to participate in this conference that will set the course for ARS infiltration research for the next decade. The discussions of current infiltration research that we heard underscore the shared roles of our two agencies for understanding and applying that research to address water-related needs for agriculture. We recognize ARS and NRCS as separate agencies that can bring their own unique expertise together to solve problems with the one objective of bringing the best possible stewardship to the nation's soil and water resources.

Our shared objectives are no accident but stem from a shared parentage. Both agencies have part of their organizational roots in the former Bureau of Plant Industry and Soils. From this heritage, we remain each others important customers. A recent example was the cooperative WEPP effort. The NRCS provided analysis of the project soils to ARS, who in turn developed a model that will serve as the cornerstone of erosion prediction used by the NRCS in the years to come.

Partnership Opportunities

There are several broad areas of potential partnership between ARS and NRCS in the area of infiltration research that were presented in our discussions.

Soil Variability

The first is achieving a better understanding of soil variability at a variety of scales. Knowledge gained in this area enhances the ability to quantitatively, and with a known degree of error, predict soil property distribution in the landscape that directly influences water movement in soils. The need to make such quantitative predictions is growing as the use of simulation models and other decision support tools increases.

In order to spatially distribute soil properties in natural landscapes in lieu of site-specific field studies, a combination of parameter estimation techniques such as pedo-transfer functions (Tietje and Tapkenhinrichs, 1993) and techniques to estimate the spatial distribution of soil properties at various scales from soil survey data (Bouma, 1973a; Bouma, et al., 1996; Grossman, et al., 1992; Wilson, et al., 1996; Wosten, et al., 1985; Wu, et al., 1996) have been used. Two recent symposia indicate the amount of research being directed at these two areas, respectively (van Genuchten, et al., 1992; Corwin and Wagnet, 1995).

If NRCS and others wish to apply ARS models and decision support technology in the many areas where NRCS is obliged to operate under its mandate to provide technical assistance to local soil and water conservation districts, we must have the ability to generate reasonable

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[1]USDA-NRCS, Federal Building, Rm 152, 100 Centennial Mall N., Lincoln, NE 68508-3866

estimates of soil properties in a variety of soils and landscapes at a variety of scales. The typical scenario is that there is often very little information except a standard soil survey and topographic data available, and additional data collection is time consuming and costly.

In the process of making a soil survey a model establishing soil-landscape relationships is developed to aid in rapidly and accurately delineating boundaries between natural soil bodies. This is what makes production of soil surveys feasible at a reasonable cost. The NRCS has a cadre of soil scientists and geomorphologists at the National Soil Survey Center (NSSC) whose main area of expertise and research center around such topics.

During the course of making a standard soil survey that conforms to National Cooperative Soil Survey (NCSS) standards, much information is gathered by observing soils and soil behavior. Much more information is gathered than is routinely displayed in the typical published soil survey. This includes everything from field descriptions to soils and landscapes, and insitu and laboratory measurements made of soil properties and water movement. As a result, NCSS soil scientists are frequently able to describe more detailed spatial and temporal soil variability in particular landscapes than they are normally able to present in the current format of published soil surveys.

New database technology and designs incorporated in the National Soil Information System (NASIS) (Soil Survey Division Staff, 1995), particularly the separation of map unit names from map unit data, and innovative display and publishing techniques will eventually allow users to access and display much of this data. It is conceivable, then, that in representative watersheds NRCS soil scientists can be called upon to describe in detail the soil components present, develop soil-geomorphic landscape models depicting where specific components are in the landscape, and conveying the relevant stratigraphy or bedrock geology affecting water flow in the landscape or watershed.

ARS researchers might find such detailed observations and studies of soils as a part of the landscape useful as ground truth in developing and testing predictive models, and in evaluating ways to predict soil variability in similar landscapes. NRCS soil scientists, in turn, would be able to use the resulting models over a much more extensive area to provide additional information to NRCS planners and engineers and to other agency customers.

Soil Porosity and Structure

A second area of mutually beneficial research centers around soil porosity. Soil aggregation and structural development impact porosity, and are often related to genetic soil factors and anthropogenic alteration of surface soil characteristics under different types of land use. The NRCS has a descriptive protocol used to describe soil structure and porosity on a macro scale in natural soil bodies (Soil Survey Division Staff, 1993).

The connection between basic concepts of elemental and secondary structure impacting water movement and standard soil survey descriptive protocols has been established for a number of years (Bouma, 1973b). Researchers have established connections between soil structure and saturated hydraulic conductivity and porosity values (McKeague, et al., 1982; McKeague, 1987). Concepts of macropores and macropore flow are an accepted part of the soil science literature (Beven and Germann, 1982). Some countries have instituted protocols for adjusting K_{sat} to take into account the effects of soil structure on porosity (Griffiths, 1985). Grossman (1993) has proposed some tentative structural adjustment protocols that address macropores to place soils in K_{sat} classes for use in NRCS.

Structural concepts as outlined above fall under the general category of soil fabric (Brewer, 1976). The NSSC has on staff scientists with expertise in soil fabric analysis from the field to the microscopic scale. Experienced soil scientists could identify major soil structure and fabrics and work with ARS scientists to quantify porosity and other features that effect infiltration and water movements in soils. This in turn might lead to the development of conceptual and quantitative models to enhance the prediction and description of soil structure and porosity impact on water movement in a variety of soils.

The above is not new work, but it has been applied in limited geographic areas. The NSSC, however, annually samples soils and undertakes scientific soil investigations from all parts of the U.S. and usually several foreign countries. Thus, the opportunity exists to study a wide variety of soil fabrics in relation to water movement. This information would enable NRCS scientists to provide more useful information on water flow in soils to users. It would allow ARS to extrapolate research results to areas where detailed, site specific studies are not feasible or unavailable.

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Research Needs Selected and Prioritized by the Cross-Goal Interest Groups

Group 1: Mechanical/management/tillage effects - process knowledge, space-time-causal factor relationships, modeling . . .

Specific Goal 1:

High 5.[*] Address spatial variations in infiltration and runoff processes by measuring runoff measurements not only at the end of the plot, but also at points within the plot.

Medium 7. Quantify temporal responses: Temporal responses may be larger than spatial responses, but little plot and watershed data exists coupled with soil and vegetation information to explain such responses.

Specific Goal 2:

High 1. Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.

8. Test multifractal techniques for modeling spatial dependence of properties/processes.

Specific Goal 3:

High 16. The whole area of seal formation and amelioration as a sequence during secession of rains and dry periods needs attention. [Temporal changes, cover, and chemistry effects on crusts/seals; Process of flow related to crust morphology.]

Med/high 8. Develop simplified infiltration methods to measure macropore flow and its continuity with depth.

9. Collect sufficient data over space and time to characterize macropores in major soils and soil-cropping systems. [Characterize subsoil pedology in respect to macropores.]

11. Study the physics of macropore flow, in complex macropore geometries.

Medium 6. Test and further develop functions for the changes in soil bulk density due to tillage, reconsolidation, and compaction.

7. Test and further develop methods to describe effects of tillage, reconsolidation, and compaction on soil water retention and hydraulic conductivity, and their spatial variabilities.

13. The field work on decomposition of crop residues needs to be revisited.

15. An area of great need is the measure of water stable aggregation and dispersed clay as related to decomposition and supply of the aggregating agents.

18. Determine shrink-swell characteristics of important soil types.

Low 2. Develop a data base on the effects of farming implements on soil structure parameter.

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[*]These numbers refer to the serial numbers of the original consolidated list of research needs for the respective Specific Goal.

Specific Goal 4:

High 2. Probably the largest area of uncertainty in infiltration modeling is the changes that the infiltration behavior or real soil undergoes as a result of mechanical modifications.

6. Macropore flow models need considerable testing to determine appropriate conditions for their application and parameter estimates to help users in their application.

Group 2: Experimental methods/measurements for field-scale quantification of spatial and temporal variabilities

Specific Goal 1:

High +** Need one FTP site for all data, in standardized format and information such as georeference, size, soil, methods, etc.

\+ A new book on standardized methods of measurement and analysis.

Medium 4. Investigate partial area responses: Address observations that apparent infiltration rate increases with increasing rainfall application rate.

Low 3. Define correspondence among simulator types at point, small plot, and large plot scales.

7. Quantify temporal responses: Temporal responses may be larger than spatial responses, but little plot and watershed data exists coupled with soil and vegetation information to explain such responses.

Specific Goal 2:

Medium 6. Relate hydraulic properties on rangelands to vegetation type and other methods of indirect parameter estimation.

7. Establish relationships between soil hydraulic/infiltration characteristics and remotely sensed data using theoretical and experimental analysis.

Low 4. Test and improve multiple geostatistical analysis (e.g. Co-kriging) using slope, vegetation and cover characteristics, and soil structural properties.

5. Consider multi-variable geostatistical methods as a framework for measuring the spatial variability of infiltration on rangelands.

Specific Goal 3:

Medium 8. Develop simplified infiltration methods to measure macropore flow and its continuity with depth.

Low 1. Develop a soil structure parameter which is related to water flow which incorporates the effects of tillage and reconsolidation, freeze-thaw, biological activity; roots, depth, compaction. [Rated low due to wording - important topic but not as stated.]

**Indicates a research need added to the original consolidated list.

Specific Goal 4:

High
1. Experimental measurements be made to develop a general model for the statistical character of soil area heterogeneity across [and within] major soil types.
4. Given the preponderance of daily rainfall data, much needs to be done to improve our knowledge of disaggregation statistics and rainfall intensity distributions. [Disaggregate weather service data to shorter time frames.]
5. At larger scales (e.g. 10 ha +), modeling a really variable infiltration should not be done independently of the surface runoff, itself with considerable organized and random heterogeneity, nor should it be modeled without consideration of small-scale rainfall rate heterogeneities.

Low
6. Macropore flow models need considerable testing to determine appropriate conditions for their application and parameter estimates to help users in their application.

Additional Recommendations/Comments (not rated):

+ Management impacts, macropores, seals, and root effects are national problem areas; grazing, freezing-thawing, swelling-shrinking, compaction, and stone effects are regional problems.
+ Measuring infiltration of natural rains is the highest national priority; rain simulation comes second.

Group 3: Biological effects --roots, canopy, worms, . . .process knowledge, space-time-causal factor relationships, modeling, . . .

Specific Goal 2:

High
4. Test and improve multiple geostatistical analysis (e.g. Co-kriging) using slope, vegetation and cover characteristics, and soil structural properties.
6. Relate hydraulic properties on rangelands to vegetation type and other methods of indirect parameter estimation.

Specific Goal 3:

High
10. Develop improved predictive models by including:
    b) seasonally varying infiltration rates due to plant growth and (fall) worm activity, shrink/swell;
    c) systematic small- scale spatial variability such as crop- row position effects.
12. A much more comprehensive review and evaluation of existing data and creation of new data to quantify/estimate the temporal/spatial character of seal/crust as influenced by residue cover, soil type, and crops.
13. The field work on decomposition of crop residues [and animal waste] needs to be revisited.
23. Measure and quantify spatial variability of infiltration parameters between plant bases and interspaces, as a function of soil type, grazing intensity, and other factors.

Medium
1. Develop a soil structure parameter which is related to water flow which incorporates the effects of tillage and reconsolidation, freeze-thaw, biological activity; roots, depth, compaction [,and worm holes].
9. Collect sufficient data over space and time to characterize macropores in major soils and soil-cropping systems.

Additional Recommendations/Comments (not rated):

- + The end product of above research should be a model; look at the above needs form this view point.
- + Bring in other disciplines as needed - microbiologist, plant physiologist, chemist, etc.
- + Set up a team to develop a coordinated plan.

Group 4: Physical effects--freezing-thawing, hydrophobicity, swelling-shrinking, . . .

Specific Goal 1:

Medium 4. Investigate partial area responses: Address observations that apparent infiltration rate increases with increasing rainfall application rate.

Specific Goal 2:

High 5. Consider multi-variable geostatistical methods as a framework for measuring the spatial variability of infiltration on rangelands.
8. Test multifractal techniques for modeling spatial dependence of properties/processes.

Specific Goal 3:

High 12. A much more comprehensive review and evaluation of existing data and creation of new data to quantify/estimate the temporal/spatial character of seal/crust as influenced by residue cover, soil type, and crops.
15. An area of great need is the measure of water stable aggregation and dispersed clay as related to decomposition and supply of the aggregating agents.
16. The whole area of seal formation and amelioration as a sequence during secession of rains and dry periods needs attention.
17. Develop methods to measure/estimate and quantify the seal/crust conductivity and distribution on a field scale.
25. Study the occurrence and effects of microbial crusts from different soil textures, climates, and grazing management on infiltration.

Medium 18. Determine shrink-swell characteristics of important soil types.
19. Relate shrink-swell characteristics to soil properties as well as management practices.
20. Test and refine methods for prediction of soil water content with time in swell-shrink soils.
21. Characterize relationships between soil water content, ice content, hydraulic conductivity, and infiltration in frozen soils.
22. Better measurement and characterization of ice content and structure (i.e., ice lenses).
23. Measure and quantify spatial variability of infiltration parameters between plant bases and interspaces, as a function of soil type, grazing intensity, and other factors.
24. Characterize the occurrence of water repellency in rangelands as a function of soil and grazing.
\+ Determine shrink/swell effect on aggregation.

Low 1. Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.

Specific Goal 4:

Medium 7. More knowledge is needed on the physics of the freezing and thawing of soils to determine the relation of partial freezing conditions to the reduction of hydraulic conductivity.

Group 5: Parameterization/estimation***

Specific Goal 1:

High
7. Quantify temporal responses: Temporal responses may be larger than spatial responses, but little plot and watershed data exists coupled with soil and vegetation information to explain such responses.
8. Design an ARS rainfall simulation experimental procedure which would maximize the information collected and provided for consistency in data between experiments.
9. Develop consolidated database of existing plot and watershed infiltration data. Similarities and differences in data sets need to be identified and documented.
+ Develop a data collection protocol - minimum data set appropriate methodology, and data format.

Medium
+ Develop a data collection network - instrumented natural rainfall sites.

Specific Goal 2:

High
3. Incorporate terrain attributes into relationships for surface soil attributes, soil hydraulic properties and soil textures to obtain regional-scale prediction equations.
+ Move toward mechanistic (process)-based parameter estimation (long term).
+ Develop methods to incorporate parameter variability into parameter estimation.

Medium
6. Relate hydraulic properties on rangelands to vegetation type and other methods of indirect parameter estimation.
7. Establish relationships between soil hydraulic/infiltration characteristics and remotely sensed data using theoretical and experimental analysis.
+ Develop "indices" from new data sources (e.g., remote sensing) for parameter estimation.

Specific Goal 3:

High
12. A much more comprehensive review and evaluation of existing data and creation of new data to quantify/estimate the temporal/spatial character of seal/crust as influenced by residue cover, soil type, and crops.
17. Develop methods to measure/estimate and quantify the seal/crust conductivity and distribution on a field scale.

Specific Goal 4:

Medium
4. Given the preponderance of daily rainfall data, much needs to be done to improve our knowledge of disaggregation statistics and rainfall intensity distributions.

***This group did not specifically prioritize the consolidated research needs, but based on the priority areas deliniated the following research needs were identified.

Group 6: Stochastic statistical characterizations and modeling

Specific Goal 1:

Medium + Design intensities/variations/storms for infiltration plot studies [combined and revised from 1 & 2 below].
1. Better define relationship between simulated rainfall and natural rainfall and storm characteristics, particularly the western U.S.
2. Establish relationship between natural and artificial rainfall simulations plot response. Locations exists where both data exists, but little analysis has been completed.

Specific Goal 2:

High
1. Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.
3. Incorporate terrain attributes into relationships for surface soil attributes, soil hydraulic properties and soil textures to obtain regional-scale prediction equations.
7. Establish relationships between soil hydraulic/infiltration characteristics and remotely sensed data using theoretical and experimental analysis.

Medium
4. Test and improve multiple geostatistical analysis (e.g. Co-kriging) using slope, vegetation and cover characteristics, and soil structural properties.
5. Consider multi-variable geostatistical methods as a framework for measuring the spatial variability of infiltration on rangelands.
8. Test multifractal techniques for modeling spatial dependence of properties/processes.

Specific Goal 3:

Medium
10. Develop improved predictive models by including:
    a) short- term temporally varying infiltration rates due to reconsolidation and surface sealing;
    b) seasonally varying infiltration rates due to plant growth and (fall) worm activity, shrink/swell;
    c) systematic small- scale spatial variability such as crop- row position effects.

\+ Evaluate the statistical properties of effective size, geometry, and distribution of macropores toward developing a deterministic/stochastic field model.

Specific Goal 4:

High
4. Given the preponderance of daily rainfall data, much needs to be done to improve our knowledge of disaggregation statistics and rainfall intensity distributions.
5. At larger scales (e.g. 10 ha +), modeling a really variable infiltration should not be done independently of the surface runoff, itself with considerable organized and random heterogeneity, nor should it be modeled without consideration of small-scale rainfall rate heterogeneities.

Medium
3. There is progress being made in modeling the aggregate behavior of an area containing internal infiltration variability, but there remain significant challenges in their application in "management" modeling.

Group 7: Spatial characterization needs for precision farming

Specific Goal 1:

Low

4. Investigate partial area responses: Address observations that apparent infiltration rate increases with increasing rainfall application rate.
5. Address spatial variations in infiltration and runoff processes by measuring runoff measurements not only at the end of the plot, but also at points within the plot.
6. Examine all components of measured hydrographs. Progress in infiltration research is dependent on being able to define the change in depression and surface storage with time as well as being able to compute runoff.
7. Quantify temporal responses: Temporal responses may be larger than spatial responses, but little plot and watershed data exists coupled with soil and vegetation information to explain such responses. [Temporal responses are a response to process in nature and we must understand the processes.]

Specific Goal 2:

Medium

1. Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.
3. Incorporate terrain attributes into relationships for surface soil attributes, soil hydraulic properties and soil textures to obtain regional-scale prediction equations.
4. Test and improve multiple geostatistical analysis (e.g. Co-kriging) using slope, vegetation and cover characteristics, and soil structural properties [because we need a stat's method of relating soils and other attributes; sampling is a major issue, will need new stat's].
7. Establish relationships between soil hydraulic/infiltration characteristics and remotely sensed data using theoretical and experimental analysis. [Sampling is a problem, we must do it cheaply, co-krig with something, that is expensive to collect for ground truth.]

Specific Goal 3:

Medium

1. Develop a soil structure parameter which is related to water flow which incorporates the effects of tillage and reconsolidation, freeze-thaw, biological activity; roots, depth, compaction.

\+ The combination of the following are important:
   3. Incorporate into present models of bulk density the process soil compaction models.
   6. Test and further develop functions for the changes in soil bulk density due to tillage, reconsolidation, and compaction.
   7. Test and further develop methods to describe effects of tillage, reconsolidation, and compaction on soil water retention and hydraulic conductivity, and their spatial variabilities.

9. Collect sufficient data over space and time to characterize macropores in major soils and soil-cropping systems. [We need to understand the development of macropores and their persistence under different management methods and how they influence infiltration and the movement of nutrients and pesticides. Macropores will be different between no-till and tillage management, and different again below the area of influence of the crop zone.]
13. The field work on decomposition of crop residues needs to be revisited. [Idea of evaluating residue production, how it's incorporated, and how it becomes functional infiltration variation and is under the control of management.]
19. Relate shrink-swell characteristics to soil properties as well as management practices.

Specific Goal 4:

High 3. There is progress being made in modeling the aggregate behavior of an area containing internal infiltration variability, but there remain significant challenges in their application in "management" modeling.
+ Decision support system for precision farming. [We need to combine GIS data bases, remote sensing, and available predictive simulation/management models for producers to use as a management tool.]

Group 8: Special topics: Minimum data set, methodologies, amending infiltration, and some general items

Specific Goal 1:

High + A working group should be formed to develop a standard list of minimum data set. Revise Hydrology Handbook.

Specific Goal 2:

Medium 1. Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.
3. Incorporate terrain attributes into relationships for surface soil attributes, soil hydraulic properties and soil textures to obtain regional-scale prediction equations.

Specific Goal 3:

High 1. Develop a soil structure parameter which is related to water flow which incorporates the effects of tillage and reconsolidation, freeze-thaw, biological activity; roots, depth, compaction.
8. Develop simplified infiltration methods to measure macropore flow and its continuity with depth.

Medium 10. Develop improved predictive models by including:
a) short- term temporally varying infiltration rates due to reconsolidation and surface sealing;
b) seasonally varying infiltration rates due to plant growth and (fall) worm activity, shrink/swell;
c) systematic small- scale spatial variability such as crop- row position effects.
+ Amend and manage soils to increase infiltration.
+ A quick, simple, and affordable method of estimating infiltration across a field. Set up a working group.

Specific Goal 4:

High + Decision support system for precision farming.
+ Physically-based, simple infiltration model - convertible to time basis (2D G&A).

Medium 3. There is progress being made in modeling the aggregate behavior of an area containing internal infiltration variability, but there remain significant challenges in their application in "management" modeling. [Management techniques to raise/lower (tillage, residue, chemical, "surge," amendments).]

Additional Recommendations/Comments:

Medium + Develop a network mechanism to communicate innovative methodologies and equipment.

Enhanced Consolidated Research Needs and Priorities Assigned by Cross-Goal Groups

Specific Goal 1: Field Research

1. Better define relationship between simulated rainfall and natural rainfall and storm characteristics, particularly the western U.S.

2. Establish relationship between natural and artificial rainfall simulations plot response. Locations exists where both data exists, but little analysis has been completed.

+[*] Design intensities/variations/storms for infiltration plot studies [combined and revised from 1 & 2 above].
M(6)[**]

3. Define correspondence among simulator types at point, small plot, and large plot scales.
L(2)
4. Investigate partial area responses: Address observations that apparent infiltration rate increases with increasing rainfall application rate.
M(2),M(4),L(7)
5. Address spatial variations in infiltration and runoff processes by measuring runoff measurements not only at the end of the plot, but also at points within the plot.
H(1),L(7)
6. Examine all components of measured hydrographs. Progress in infiltration research is dependent on being able to define the change in depression and surface storage with time as well as being able to compute runoff.
L(7)
7. Quantify temporal responses: Temporal responses may be larger than spatial responses, but little plot and watershed data exists coupled with soil and vegetation information to explain such responses. [Temporal responses are a response to process in nature and we must understand the processes.]
M(1),L(2),H(5),L(7)
8. Design an ARS rainfall simulation experimental procedure which would maximize the information collected and provided for consistency in data between experiments.
H(5)
9. Develop consolidated database of existing plot and watershed infiltration data. Similarities and differences in data sets need to be identified and documented. [Need one FTP site for all data, in standardized format and information such as georeference, size, soil, methods, etc.]
H(2),H(5)

+ A new book on standardized methods of measurement and analysis; revised Hydrology Handbook.
H(2),H(8)

+ Develop a data collection protocol - minimum data set appropriate methodology, and data format.
H(5),H(8)

+ Develop a data collection network - instrumented natural rainfall sites.
M(5)

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[*]Research need added by Cross-Goal Groups.

[**]The letter represents the level of priority assigned - L:low, M:medium,H:high; and the number in the parenthesis denotes the Cross-Goal Group number.

Specific Goal 2: Spatial Variability

1. Examine relationships between correlation scales and measurement scales in order to minimize the affect of the method (or size of the measurement) on the resulting analysis.
   H(1),H(6),M(7),M(8)
2. Expand methods for large area analysis.

3. Incorporate terrain attributes into relationships for surface soil attributes, soil hydraulic properties and soil textures to obtain regional-scale prediction equations.
   H(5),H(6),M(7),M(8)
4. Test and improve multiple geostatistical analysis (e.g. Co-kriging) using slope, vegetation and cover characteristics, and soil structural properties [because we need a stat's method of relating soils and other attributes; sampling is a major issue, will need new stat's].
   L(2),H(3),M(6),M(7)
5. Consider multi-variable geostatistical methods as a framework for measuring the spatial variability of infiltration on rangelands.
   L(2),H(4),M(6)
6. Relate hydraulic properties on rangelands to vegetation type and other methods of indirect parameter estimation.
   M(2),H(3),M(5)
7. Establish relationships between soil hydraulic/infiltration characteristics and remotely sensed data using theoretical and experimental analysis. [Sampling is a problem, we must do it cheaply, co-krig with something, that is expensive to collect for ground truth.]
   M(2),M(5),H(6),M(7)
8. Test multifractal techniques for modeling spatial dependence of properties/processes.
   H(1),H(4),M(6)

\+ Move toward mechanistic (process)-based parameter estimation (long term).
  H(5)

\+ Develop methods to incorporate parameter variability into parameter estimation.
  H(5)

\+ Develop "indices" from new data sources (e.g., remote sensing) for parameter estimation.
  M(5)

Specific Goal 3: Temporal Variability

1. Develop a soil structure parameter which is related to water flow which incorporates the effects of tillage and reconsolidation, freeze-thaw, biological activity; roots, depth, compaction [,and worm holes]. [Rated low due to wording - important topic but not as stated.]
   L(2),M(3),L(4),M(7),H(8)
2. Develop a data base on the effects of farming implements on soil structure parameter.
   L(1)
3. Incorporate into present models of bulk density the process soil compaction models.
   M(7)
4. Develop methods for integrating the spatial variability caused by tillage into a water related soil structure parameter.

5. Develop methods for characterizing the temporal variation of macropores, caused by tillage, reconsolidation.

6. Test and further develop functions for the changes in soil bulk density due to tillage, reconsolidation, and compaction.
   M(1),M(7)

7. Test and further develop methods to describe effects of tillage, reconsolidation, and compaction on soil water retention and hydraulic conductivity, and their spatial variabilities.
   M(1),M(7)
8. Develop simplified infiltration methods to measure macropore flow and its continuity with depth.
   M/H(1),M(2),H(8)
9. Collect sufficient data over space and time to characterize macropores in major soils and soil-cropping systems . [Characterize subsoil pedology in respect to macropores. We need to understand the development of macropores and their persistence under different management methods and how they influence infiltration and the movement of nutrients and pesticides. Macropores will be different between no-till and tillage management, and different again below the area of influence of the crop zone.]
   M/H(1),M(3),M(7)
10. Develop improved predictive models by including:
    a) short- term temporally varying infiltration rates due to reconsolidation and surface sealing;
    b) seasonally varying infiltration rates due to plant growth and (fall) worm activity, shrink/swell;
    c) systematic small- scale spatial variability such as crop- row position effects.
    H(3),M(6),M(8)
11. Study the physics of macropore flow, in complex macropore geometries.
    M/H(1)
12. A much more comprehensive review and evaluation of existing data and creation of new data to quantify/estimate the temporal/spatial character of seal/crust as influenced by residue cover, soil type, and crops.
    H(3),H(4),H(5)
13. The field work on decomposition of crop residues [and animal waste] needs to be revisited. [Idea of evaluating residue production, how it's incorporated, and how it becomes functional infiltration variation and is under the control of management.]
    M(1),H(3),M(7)
14. Characterization of the aggregating compounds should be associated process information.

15. An area of great need is the measure of water stable aggregation and dispersed clay as related to decomposition and supply of the aggregating agents.
    M(1),H(4)
16. The whole area of seal formation and amelioration as a sequence during secession of rains and dry periods needs attention. [Temporal changes, cover, and chemistry effects on crusts/seals; Process of flow related to crust morphology.]
    H(1),H(4)
17. Develop methods to measure/estimate and quantify the seal/crust conductivity and distribution on a field scale.
    H(4),H(5)
18. Determine shrink-swell characteristics of important soil types.
    M(1),M(4)
19. Relate shrink-swell characteristics to soil properties as well as management practices.
    M(4),M(7)
20. Test and refine methods for prediction of soil water content with time in swell-shrink soils.
    M(4)
21. Characterize relationships between soil water content, ice content, hydraulic conductivity, and infiltration in frozen soils.
    M(4)
22. Better measurement and characterization of ice content and structure (i.e., ice lenses).
    M(4)
23. Measure and quantify spatial variability of infiltration parameters between plant bases and interspaces, as a function of soil type, grazing intensity, and other factors.
    H(3),M(4)
24. Characterize the occurrence of water repellency in rangelands as a function of soil and grazing.
    M(4)

25. Study the occurrence and effects of microbial crusts from different soil textures, climates, and grazing management on infiltration.
H(4)

+ Evaluate the statistical properties of effective size, geometry, and distribution of macropores toward developing a deterministic/stochastic field model.
M(6)

+ Amend and manage soils to increase infiltration.
M(8)

+ A quick, simple, and affordable method of estimating infiltration across a field. Set up a working group.
M(8)

Specific Goal 4: Computer Modeling

1. Experimental measurements be made to develop a general model for the statistical character of soil area heterogeneity across [and within] major soil types.
H(2)
2. Probably the largest area of uncertainty in infiltration modeling is the changes that the infiltration behavior or real soil undergoes as a result of mechanical modifications.
H(1)
3. There is progress being made in modeling the aggregate behavior of an area containing internal infiltration variability, but there remain significant challenges in their application in "management" modeling. [We need to combine GIS data bases, remote sensing, and available predictive simulation/management models for producers to use as a management tool. Management techniques to raise/lower (tillage, residue, chemical, "surge," amendments).]
M(6),H(7),M(8)
4. Given the preponderance of daily rainfall data, much needs to be done to improve our knowledge of disaggregation statistics and rainfall intensity distributions. [Disaggregate weather service data to shorter time frames].
H(2),M(5),H(6)
5. At larger scales (e.g. 10 ha +), modeling a really variable infiltration should not be done independently of the surface runoff, itself with considerable organized and random heterogeneity, nor should it be modeled without consideration of small-scale rainfall rate heterogeneities.
H(2),H(6)
6. Macropore flow models need considerable testing to determine appropriate conditions for their application and parameter estimates to help users in their application.
H(1),L(2)
7. More knowledge is needed on the physics of the freezing and thawing of soils to determine the relation of partial freezing conditions to the reduction of hydraulic conductivity.
M(4)

+ Decision support system for precision farming.
H(8),H(7)

+ Physically-based, simple infiltration model - convertible to time basis (2D G&A).
H(8)

Additional Recommendations/Comments

+ Management impacts, macropores, seals, and root effects are national problem areas; grazing, freezing-thawing, swelling-shrinking, compaction, and stone effects are regional problems.

+ Measuring infiltration of natural rains is the highest national priority; rain simulation comes second.

+ The end product of above research should be a model; look at the above needs form this view point.

+ Bring in other disciplines as needed - microbiologist, plant physiologist, chemist, etc.

+ Set up a team to develop a coordinated plan.

+ Develop a network mechanism to communicate innovative methodologies and equipment.
M(8)