WORKSHOP

ON

MODELLING OF HYDROLOGIC SYSTEMS

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Infiltration and Soil Water Movement

by

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1.0 Introduction

In water management and conservation studies, accurate information on the rate at which different soils will take in water under different field conditions is required. The rate of water entry into soil varies widely between different soil types and also within a single soil type, depending upon soil water content and management practices.

When water is applied to the soil surface through irrigation or rain, it subsequently enters the soil. If the supply rate is more than the rate of entry into the soil, the excess will accumulate over the surface and may flow as runoff. The time rate at which water will percolate into the soil through its soil-atmosphere interface is known as infiltration. Quantitatively, infiltration rate is the volume of water entering into the soil per unit area in unit time, when the soil is subjected to a shallow depth of ponding at the surface.

In hydrology, the term infiltration capacity is frequently used and is defined as the maximum rate at which rain can be absorbed by a soil in a given condition, i.e., it signifies soil infiltrability. The downward movement of water through a soil profile occurs due to tension and gravitational forces in the soil matrix. The soil matrix is generally heterogeneous and consists of a labyrinth of pores of varying shape and size connected by porous fissures and channels. Water is held in the soil matrix by tension forces at the air-water interfaces in pores. These tension forces are also frequently called suction or capillary forces.

Soil water tension varies from less than 1 inch of water head, for a soil near saturation, to as much as 10,000,000 inches of head for a very dry condition. The effect of surface tension in a soil matrix during drainage can be described by considering water held in a single pore within the soil profile connected to the groundwater table (fig.1). The figure describes an equilibrium state at the meniscus. The smaller the radius r_1 supports a column of water, h, then the gravitational forces will equal the surface tension forces and:

$$\pi r^2_1 h \rho_w g = 2\pi r_1 \tau \cos\alpha \qquad \text{(Eq 1)}$$

Where ρ_w is the density of water, g is the gravitational constant, $\rho\tau$ is the surface tension force, and α, the angle of contact of the meniscus with the soil.

Retention and movement of water during wetting or drainage is a function of the shape and size of the pores. When water is applied, soil air is displaced from the pores, the soil water content increases and the soil tension decreases. This results in decreased infiltration rates. Provided the amount of water applied is high enough, this process will continue until the soil is saturated. At saturation, the soil suction is zero. However, in most natural environments a

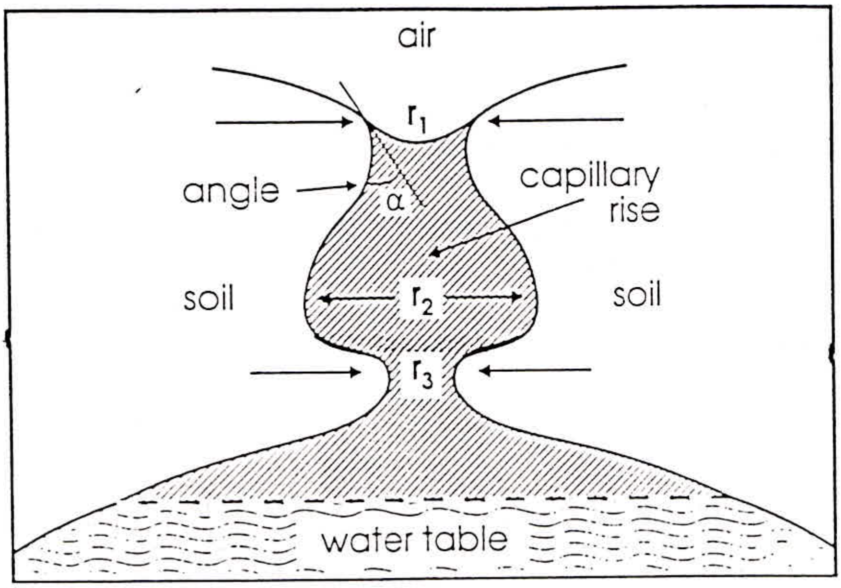

Figure 1. Water retention in a heterogenous porous media.

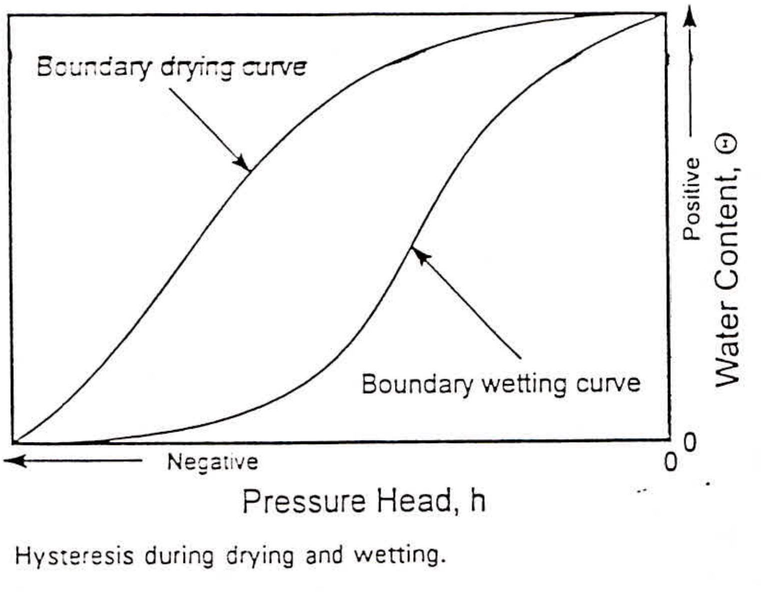

Figure 2. Hysteresis during drying and wetting.

small amount of air will be trapped in the pores and prevent complete saturation. The final degree of saturation will probably only is 80 - 90%.

Following wetting there will be a redistribution of soil water(fig 2 indicate hysterises curve) and pores will drain due to capillary and gravity flow. When gravity flow becomes negligible, the soil water content of the profile will be at field capacity. Field capacity typically occurs at soil suctions of 0.1 - 0.33 bars (fig 3). However, it will vary depending on the wetting and drying history of the soil, soil texture, porosity, and the sub-surface characteristics of the soil profile. The elusive nature of field capacity is illustrated in fig. 4.

Redistribution of water due to capillary flow will continue after gravity flow ceases. Usually, upward movement of soil water will occur due to evapotranspiration. The depth that plants can remove soil water will depend on the vegetation, profile characteristics, and climatic conditions. The soil water content in the root zone will eventually reach the wilting point unless there is near saturation. The soil suction at wilting point is typically 10 to 15 bars and is the point at which plants begin to wilt and die. For sand, the soil water content at wilting point is very low. For clay it is much higher (fig 2 to 5). Soil water in excess of the amount at the wilting point is available for evapotranspiration and the maximum amount is termed plant available soil water which is related to field capacity and wilting point as follows:

$$\theta_{paw} = \theta_{fc} - \theta_{wp} \qquad \text{(Eq 2)}$$

where, θ_{paw}, θ_{fc}, θ_{wp} are the volumetric or gravimetric plant available soil water content, and the soil water contents at field capacity and wilting point, respectively.

Knowledge of the wilting point, field capacity, and plant available soil water content is particularly important in agriculture. In arid areas, this type of information is needed to design irrigation systems and determine irrigation schedules. For most plants there is a reduction in yield when the soil waters content falls below a critical value, which is 40-70% of the plant available soil water content (fig 6).

Bodman and Coleman (1943) showed that soil water movement into uniform dry soil under conditions of surface ponding could be divided into four zones as shown in fig. 7. The saturated zone usually only extends from the surface to a depth of less than one inch. Below the saturated zone is the transition zone, which represents a zone of rapid decrease in soil water content. Below the transition zone is a zone of nearly constant soil water called the transmission zone. This zone increases in length as the infiltration process continues. Below the transmission zone is the wetting zone. The wetting zone maintains a nearly constant shape and moves downward as the infiltration proceeds. The wetting zone culminates at the wetting front, which is the boundary between the advancing water and the relatively dry soil. The wetting front represents a plane of discontinuity across which a high suction gradient occurs.

1.2 Factors Affecting Water Movement Through Soils

Infiltration may involve soil water movement in one, two, or three dimensions, although is often approximated as one-dimensional vertical flow. Water movement into and through a soil profile is dependent on many interrelated factors. Factors of importance include

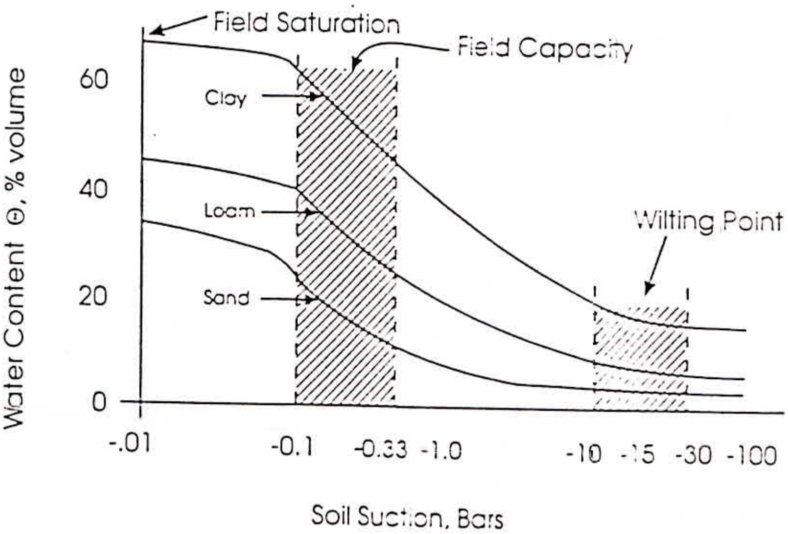

Figure 3. Soil water and soil suction relationships to field capacity wilting point and plant available water (Courtesy of S.W. Trimble, UCLA).

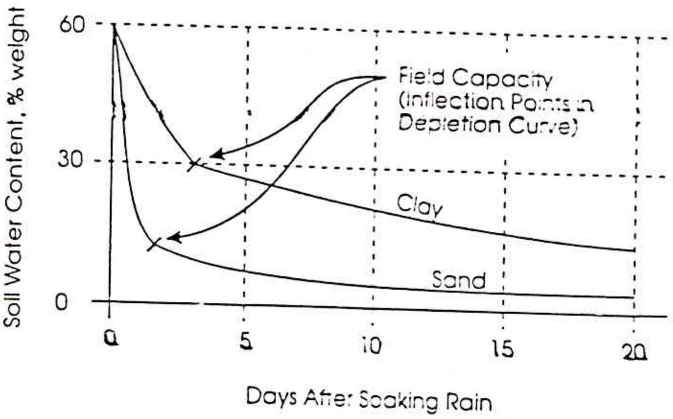

Figure 4. The elusive nature of field capacity (Courtesy of S.W. Trimble, UCLA).

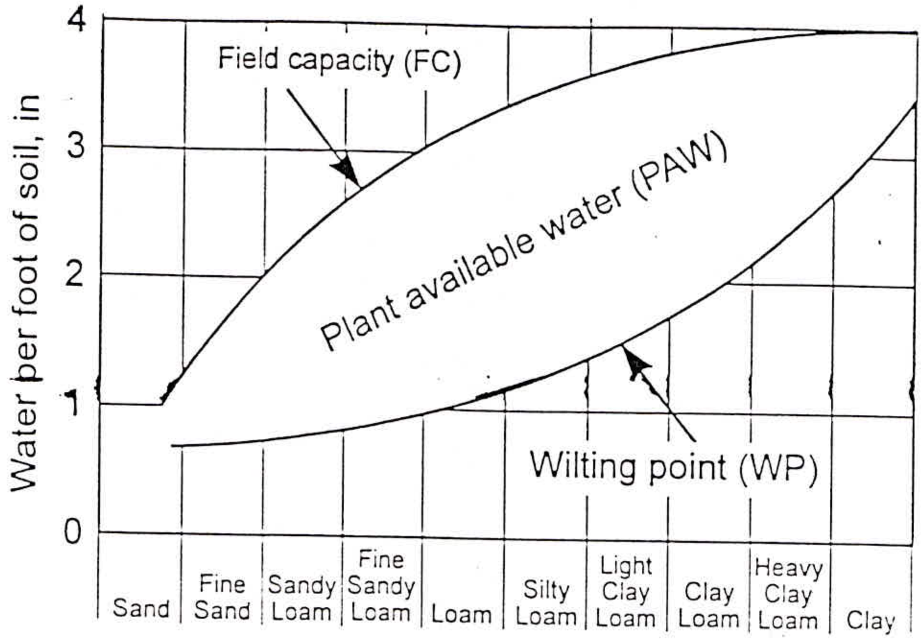

Figure 5. Water holding characteristics of soils with different texture.

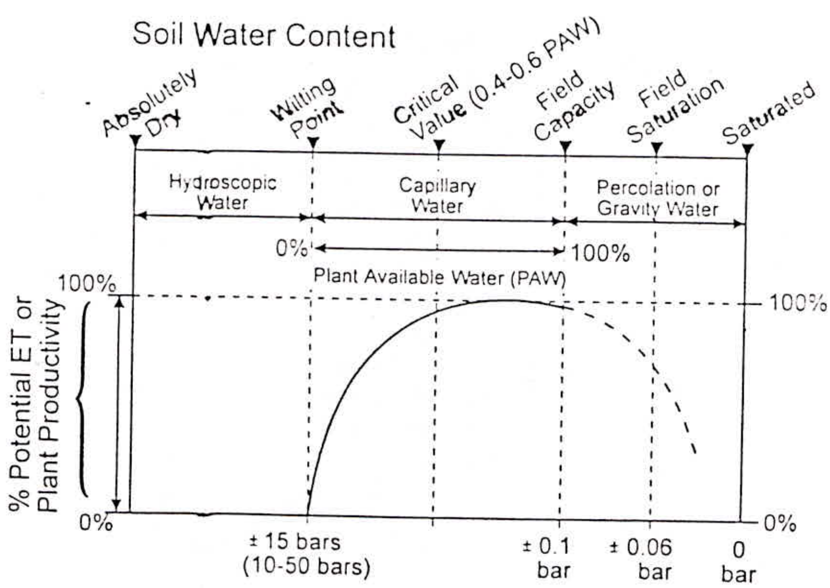

Figure 6. Yield responses to soil water content (courtesy of S.W. Trimble, UCLA).

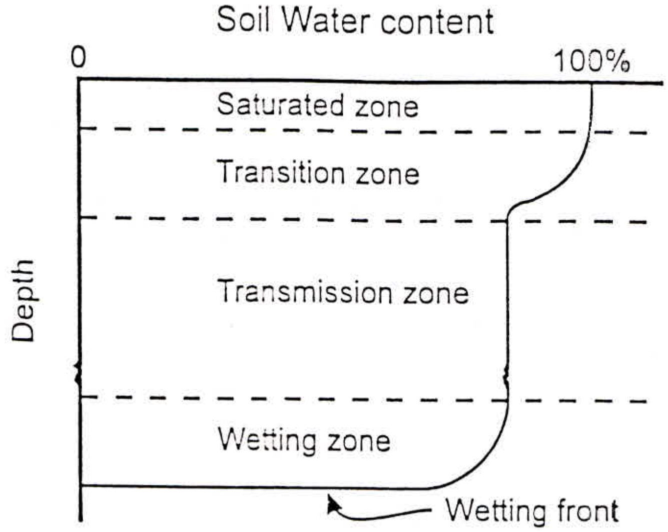

Figure 7. The infiltration zones of Bodman and Coleman (1943).

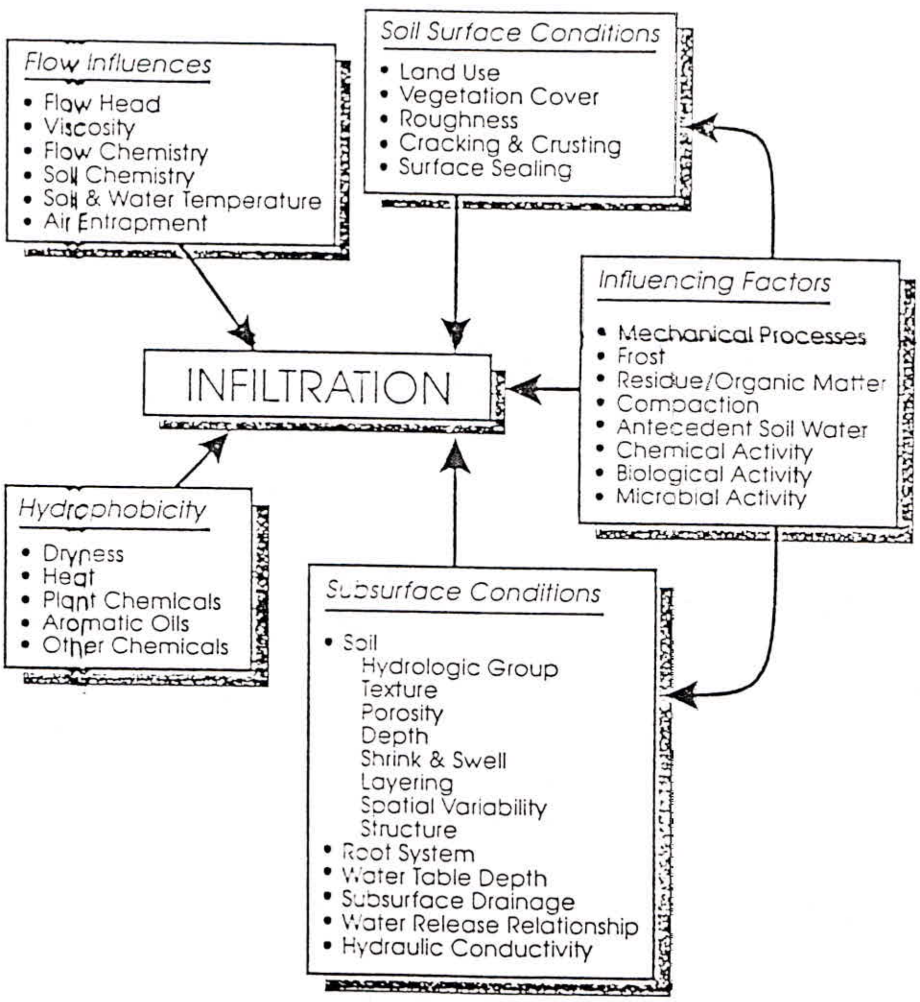

Figure 8. Factors affecting water movement through soils.

surface conditions, subsurface conditions, factors that influence surface and subsurface conditions, flow characteristics of the water or fluid, and Hydrophobicity (fig 8).

1.2.1 Surface Conditions

Vegetation and land use practices have a marked effect on infiltration. Surfaces such as paved roads, sidewalks, parking lots, and buildings allow negligible infiltration. In more natural environments, the type of vegetation, season of the year, and land management practices (such as tillage) that temporarily modify the near-surface soil conditions will greatly influence infiltration processes. Some practices, such as no-till (conservation tillage), are used to increase water movement into the soil. In other situations, such as, sites used for land disposal of toxic wastes, efforts are made to compact underlying soil in order to minimize water movement into the soil profile.

If the ability of the soil profile to transmit infiltrating water is not limiting; soil surface conditions usually govern infiltration. This is illustrated by curves of cumulative infiltration for hay cover and bare ground during the first 60 minutes of a rainstorm (fig.9). Hay provides a ground cover, which absorbs the energy of falling raindrops and prevents soil puddling or packing.

It is important to note that surface runoff occurs as soon as the infiltration rate is less than the water application rate and surface depressions have been filled with water. Once water is lost due to runoff, it is not available for infiltration. Tillage practices that leave the soil surface rough with many pockets for water storage are likely to have more infiltration than where the surface has been worked down and smoothed by tillage. In the latter case, infiltration occurs primarily while it is raining. In the first case with the rough soil surface, infiltration continues after the rain has stopped until all stored water is absorbed. This prolonged time for infiltration opportunity may be many minutes and, in some cases, significant amounts of increased infiltration result. A crust will form at the surface of some soils as they dry. Often cracks will form between crusted areas. The crusts will inhibit water movement into the soil matrix. Gravity flow will occur down the cracks provided a positive head of water is established at the top of the crack. This might be due to water ponding at the surface or surface runoff, which flows across a crusted area and into a crack. Cracks might also form in the absence of crusting and gravity flow might occur in the absence of cracks. If flow through a pre occurs primarily due to gravity flow, the pore is commonly called a macropore. Worm hole, coarse sands and gavels, and dry organic matter (residue) extending to the soil surface might all result in gravity flow. Water movement through macropores will result in more rapid wetting at deeper depths. In some cases wetting of the soil matrix might be due primarily to lateral movement of water, which has, ponded in cracks or upward movement of water from an impeding layer due to water, which reached the impeding layer by macropore flow.

Infiltration can also be impeded by surface sealing due to physical and chemical processes during a hydrologic event. As the soil surface becomes wet, a chemical reaction can cause a temporary bond between soil particles. For example, the author has conducted research on sandstone spoil materials from Kentucky surface mines. Infiltration through loosely packed air-dried soils was very rapid. However, after wetting and drying, it was noted that infiltration rates were extremely slow during any future applications of water. A very fine seal formed at the soil surface but water would flow readily through any breaks in the seal.

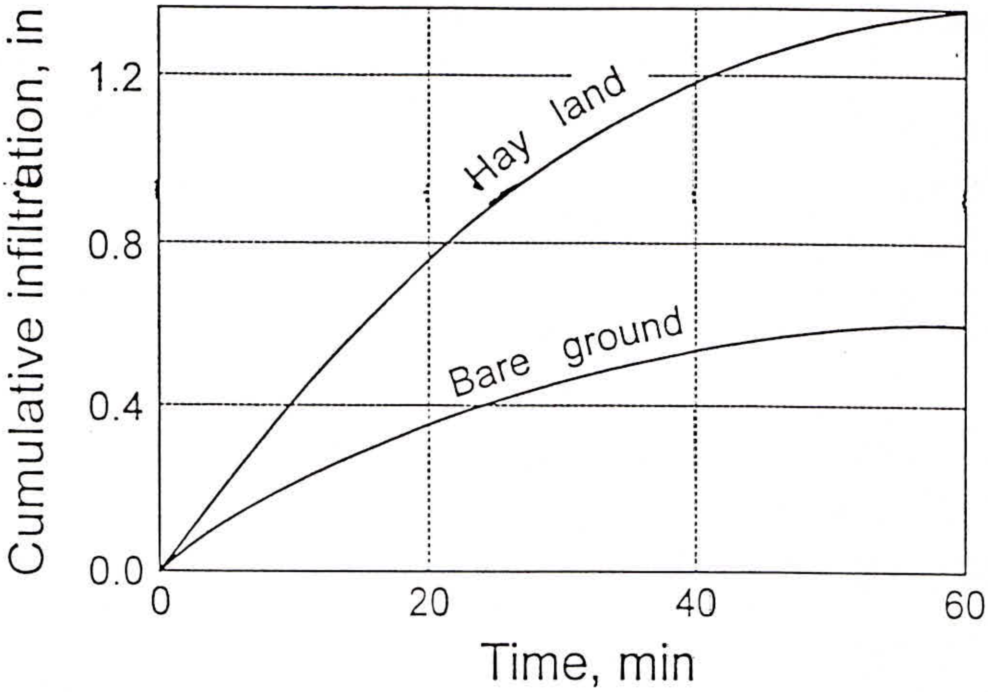

Figure 9. Infiltration into hayland and bare ground.

During high intensity rainfall or surface applications of water, soil might be detached and then moved across the soil surface. Some of the detached soil might then be redeposit in cracks and large pores. This deposition of soil in crack s and pores will tend to seal the surface, and infiltration or percolation will be reduced.

1.2.2 Subsurface Conditions

The soil texture, bulk density, soil heterogeneity, cracks and surface conditions will all influence water movement. Also, hydraulic conductivity is an important property of soils and is defined as the ability of a soil to transmit water under a unit hydraulic gradient. Hydraulic conductivity is often called permeability and is a function of oil suction and soil water content. Fine-grained soils tend to have lower hydraulic conductivity values than coarse-grained soils. However, fine-grained soils such as clays have smaller pores than coarse-grained soils such as sands. The smaller the pores the greater suction forces. Therefore water might move readily into a dry, fine-grained soil even though the hydraulic conductivity is low.

If there is no spatial variability in soil properties, the soil is described as homogeneous. Most soils are heterogeneous because they exhibit a considerable amount of variability in properties both laterally and vertically. Soils are also described as isotropic or anisotropic. Isotropic soils exhibit the same hydraulic conductivity in all directions, while anisotropic soils have different values in the vertical and lateral directions.

Land surfaces can be likened to a blotter or a sponge. Infiltration rates of the blotter are low. In order to visualize this situation; apply water onto a blotter, which is tilted. A plastic squeeze bottle can be used to apply water to the blotter. Note how little water goes into the blotter and how much runs off. The pores of the blotter are very fine and transmit water slowly. Now, apply water from the squeeze bottle onto a sponge. Note how little water runs off, even though the water application is high. Most of the sprinkled water infiltrates quickly into the sponge. Infiltration rates are high and might exceed 2 in. /hr., as found in sandy areas or in woodlands on well-drained soil. The blotter might represent slowly permeable clay with very few large pores or a well-drained silt lam where the surface had been compacted by excessive tillage and puddled and compacted by rainfall. Its infiltration capacity might be less than 0.2 in./hr. Now place the blotter on top of the sponge and apply water. Note the effect of the simulated thin, compacted layer of low infiltration capacity overlying a soil with high infiltration rates. Water will not enter the coarse textured subsurface layer until the suction forces at the interface between the two layers are equal.

The sponge on top of the blotter exemplifies a common field situation where a grass-covered, well drained soil overlays layer of soil of low permeability. This situation affects infiltration more than the sponge alone or the blotter on top of the sponge. In this case, at the start of the storm, infiltration rates into the sponge are high. However, the blotter restricts the percolation rate, causing infiltrated water to accumulate in the pores of the sponge; depleting its capacity for storing more water; and slowing down the infiltration rate until it corresponds with that of the blotter beneath.

Initial infiltration rates are higher for dry soil than for wet soil. This is illustrated in Figure 8. Infiltration rates decrease as the soil water content increases. This is because soil suction reduces with increases in soil water content. Some clay soils will swell during wetting and will then shrink during drying. Swelling will inhibit infiltration while shrinking will create cracks and increase macropore flow.

1.2.3. Hydrophobicity

Water falling on the surface of hydrophobic soil will bead up rather than enter the soil due to suction forces. This phenomenon occurs due to waxy organic materials on the soil surface, which create a negative contact angle between the waxes and any applied water. This effect normally occurs following brushland, range and forest fires. Organic matter at and above the soil surface vaporizes during the fire and then condenses on the bare burned soil as waxy materials.

1.2.4. Flow Characteristics

Viscosity of water can affect infiltration. The colder the temperature, the higher the viscosity and the slower the infiltration. In watershed studies the practicing hydrologist often neglects the viscosity of water flow. However in most laboratory research, viscosity of water is a significant factor. The effect of frost on infiltration will depend on the soil water content at the time of freezing. A very wet, frozen soil can be practically impervious, causing a condition commonly called concrete frost. A dry forest or hayland frozen soil is likely to be porous providing nearly normal infiltration rates and a condition called lattice frost.

As water infiltrates through the soil, not all air in the pores is freely displaced by the flow and air bubbles are trapped in the pores. Trapped air will block some pores and will retard infiltration as air bubbles try to move upwards against the direction of water flow. If you observe a puddle on bare soil immediately following a storm event, it is possible that you will see bubbles of air coming to the surface of the puddle. The soil water content following a rainfall event will depend on the rainfall depth, duration, and intensities. Intermittent light rain will produce higher soil water contents than if the same amount of rain occurred in a short period. This is due, in part, to the fact that during light intermittent rain, it is easier for air bubbles to escape upwards throughout the soil profile

Entrapment of air is also dependent on the physical properties of the soil and in particular soil pore sizes. Very small pores will have a larger amount of air entrapment. This is because the suction forces pulling water into the pores are larger than it would be for larger pores. These high suction forces oppose upward movement of air and, when combined with the small pore sizes, result in air being trapped or pushed downwards. The likelihood that air gets pushed downwards into the soil profile ahead of the wetting front increases and this causes a build up of air pressure which opposes the downward movement of water.

1.3 Soil Water Balance

Changes in soil water storage occur throughout the year. Increases occur due to precipitation, irrigation, and subsurface inflows. Percolation, gravitational drainage, and evapotranspiration cause depletion. A water balance equation describing these changes for any period of time is expressed as:

$$\Delta SM = P + IR - Q - G - ET \qquad \text{(Eq 3)}$$

where: ΔSM is the change in soil water storage in the soil profile, P is precipitation. IR is irrigation, G is percolation water, ET is evapotranspiration and Q is surface runoff. All quantities are expressed as a depth (inches or mm) of water over a study area for a specific period of time.

1.4 Estimating Infiltration Rates

1.4.1 Horton Equation

One of the most widely used infiltration models is the three-parameter equation developed by Horton (1939):

$$f = f_c + (f_o - f_c)\, e^{-\beta t} \qquad \text{(Eq 4)}$$

where, f is the infiltration rate at time t, f_o is the infiltration rate at time zero, f_c is the final constant infiltration capacity and β is a best-fit empirical parameter. Advantages of the method are that the equation is simple and usually gives a good fit to measured data because it is dependent on three parameters. Main disadvantages are that the method has no physical significance and field data are required to calibrate the equation. The equation does not describe infiltration prior to ponding.

Horton's equation has seen widespread application in storm watershed models. The most commonly used model which uses Horton's method is the Environmental Protection Agency Storm Water Management Model (Huber , 1981).

1.4.2 Green-Ampt Equation

In 1911, Green and Ampt developed an analytical solution of the flow equation for infiltration under a constant rainfall. The method is developed directly from Darcy's law and assumes a capillary -tube analogy for flow in a porous soil. The equation can be written as:

$$f = K (H_o + S_w + L) / L \ .... \text{(Eq 5)}$$

where, K is the hydraulic conductivity of the transmission zone, H_o is the depth of flow ponded at the surface, S_w is the effective suction at wetting from and L is the depth from the surface to the wetting f front. The method assumes piston or plug flow and distinct wetting front between the infiltration zone and soil at the initial water content (refer back to fig 5).

The Green Ampt method is often approximated by the equation:

$$f = \frac{A}{F} + B \qquad \text{(Eq 6)}$$

where f is the infiltration rate, F is the accumulative infiltration, and A and B are fitted parameters which depend on the initial soil water content, surface conditions, and soil properties.

1.5 The influence of spatial variation in biological and soil properties

On natural catchments, even in the unusual situation of completely uniform rainfall, there are spatial variations in the supply rate to the surface of the soil. They arise from not only spatial variations in the properties of the soil matrix, but also a number of other factors. By the process of interception, vegetation removes part of the rainfall and channels other parts to particular localities by the processes of stem flow and leaf drip. Vegetation also reduces the impact energy of droplets, so reducing surface slaking and crusting. Removal of water by transpiration results in a more uniform reduction of soil moisture with depth, enhancing infiltration, whilst the accumulation of organic matter provides a suitable substrate for soil microorganisms as well as conferring a degree of structural stability to the soil mantle and impeding overland flow.

Spatial variations in geology, weathering and erosion processes contribute to the creation of small depressions and of variations in surface slope and roughness, and soil matrix properties. At the microhydrological level, the supply of water for infiltration can be enhanced by local run-on or lateral redistribution of surface water, a process of particular significance in arid regions (Fleming 1976). On a catchment scale, these micro hydrological processes interact, so making spatial averages more realistic. Lateral redistribution at the micro-scale can be dealt with by mapping land capability classes and identifying their relative topographic sequence(Holtan and Lopez 1971). A fairly common sequence in Australia is the occurrence of skeletal and shallow soils in the higher parts of the landscape, deeper weathered and colluvial-alluvial soils on the midslope, and alluvial soils near the stream. Downslope movement of water in the saturated and unsaturated phases, particularly during periods of above average rainfall, provides areas of high water content and so low infiltration at breaks - in-slope and near the stream. Overland flow generated from areas of low soil moisture storage and / or high moisture content can run over areas with higher infiltration capacity. There is a close analogy with the irrigation advance problem widely explored in the irrigation literature. Irrigation bays are, however, usually assumed to have uniform slope without cross-fall and to be without direct infiltration supply from rainfall. It is considered that the irrigation-advance solution should be attempted only in cases where there are great differences in infiltration characteristics in a downslope direction.

More important is the manner in which known and unknown variations in soil properties influence the application of infiltration equations to nominally uniform areas. Zaslavsky (1970) has discussed this problem at length and implies that it is impossible to determine meaningful average values for substitution in infiltration equations. Rogowski (1972), on the other hand, has shown that many soil parameters are statistically normally distributed when sampled from a single mapped soil-type and that averages and standard errors can be estimated. If the standard errors are large, however, the range of error in estimated runoff is very large indeed (Zaslavsky 1970). It is therefore most important to take an adequate number of field samples to allow estimation of the reality of the algorithms used.

1.6 Infiltration and catchment modelling

For the purposes of this discussion we will examine only deterministic models, since in these models infiltration is usually modelled explicitly. In contrast, stochastic modelling usually treats infiltration only in a vague and implicit way. Deterministic models are usually subdivided into "lumped - parameter models" and distributed -parameter models", with the difference really lying in the area over which point water balance equations are presumed to be valid.

In a lumped-parameter model a water balance is struck for a single point on the catchment and input water is apportioned to appropriate stores and fluxes. In such a model the parameters concerned with infiltration must implicitly or explicitly attempt to average all catchment values. In the cases discussed by Boughton (1966) and Chapman (1970) the whole catchment is implicitly averaged. In the case of Stanford IV or HSP (Fleming and Black 1974) and the USGS Model (Dawdy, Lichty and Bergman 1972) an explicit attempt is made to model the interaction between a range of infiltration capacities and the supply rate of water to the infiltrating surface. At present, a simple linear variation from an estimated maximum infiltration capacity to zero issued and in the case of HSP an apportionment to interflow calculated on a similar basis.

None of these models can be considered to realistically tackle the problem of spatial variation in catchment infiltration parameters except that, hopefully, they apply Rogowski's averaging concept on a whole catchment basis. The Hydrocomp Simulation Programme does have the capacity to combine the outputs from multiple subcatchments by modelling the hydraulic processes of conveyance of flood-flows down the connecting stream network. Each subcatchment can have its own infiltration parameters.

In the distributed-parameter model, water balances are calculated at multiple locations in the catchment, either on a grid basis or in natural topographic and soil units. By the generation of components of the water balance at known locations in space and time it is possible to include surface and subsurface water movements into and out of the gird points or landscape elements for greater realism.

These concepts have been incorporated into practical hydrologic models by Huggins and Mongke (1968) and Holtan and Lopez (1971), although in both cases infiltration parameters are averaged over areas mapped as the same soil.

Freeze (1972a, 1972 b) using a distributed model, has carried out a simple two-dimensional analysis of a catchment slice with uniform soil properties K (θ) and $\Psi(\theta)$. He observed that extensive overland flow only occurs with very high intensity rainfall or unusually impermeable soils. Overland flow, however, does occur from elements low in the land scape and close stream channels, where initial water contents tend to be high as a result of downslope movement, within the profile, of water which infiltrated at an earlier time. Based on the study it is also suggested that permeability and water potential gradients in the soil mantle are too low to support the concept of interflow as a major component in the surface hydrograph, except as immediate drainage of saturated channel bordering areas. It is also apparent that most practical models do not adequately model the space and time variation of infiltration.

We therefore suggest that the most important factor in realistic catchment modelling of the infiltration process is the identification in space and time of areas of high initial moisture content, since these will be the areas which pass from pre-ponding to ponded infiltration, during storm events, and so generate surface run-off. The important, and measurable, soil physical relationships are therefore, the variation of sorptivity (S) with initial moisture content as well as the long-term infiltration rates.

Practically, it may be expected that baseflow level will be a good index of groundwater contributions to the high moisture content areas, and the actual contributing area and the values of S and θo will also be related to the level of baseflow discharge. These features could be incorporated in a quite simple model to predict surface runoff and redistribution of infiltrated water to a baseflow generating groundwater store.

1.7 Significance of Infiltration data

Infiltration in uniform porous media is well understood and the physics of the simple case of one-phase flow can be rigorously derived from measurable soil properties, K (θ) and $\Psi(\theta)$. In the case of non-cracking clay soils, the relative magnitudes of the diffusivity and conductivity functions are such that for normal rainfall events the gravitational effect on infiltration may be disregarded. The behavior of cracking soils depends very much on the nature of the rainfall, as well as the initial moisture content and the surface cover of organic matter.

In catchment hydrology, the spatial and temporal variations in rainfall intensity and the properties of the soil make simple application of soil physics difficult. With respect to temporal variation in rainfall intensity it is concluded that prediction of the time of occurrence of surface ponding can be made from an estimate of the infiltrated volume and the simple Philip two-parameter infiltration equation. The spatial infiltration parameters determine the pattern of surface and sub-surface flow. Ponding begins from preferred locations with low values of infiltration parameters, due either to intrinsic soil properties or to high moisture contents resulting from sub-surface flow. Most expressions used for infiltration calculation in the hydrologic literature do not explicitly account for either the pre-ponding/post-ponding situation, or the range of infiltration parameters present on the catchment. However, these features are implicit in the infiltration algorithms incorporated in the more successful hydrologic models, such as HSP. It is believed that detailed prediction of infiltration in catchment hydrology will depend on the prediction of the space and time locations of the occurrence of ponding and hence the location and magnitude of overland flow and interflow source areas. The algorithms, which incorporate these predictions, will be based on conventional soil-physical principles and use information from detailed field observations.

Bibliography

1. Chapman, T. G., and Dunin, F.X., 1975. Prediction in Catchment Hydrology, Australian Academy of Science.

2. Freeze, R.A., and J. A. Cherry, 1979. Groundwater. Prentice-Hall, Englewood Cliffs, NJ.

3. Hillel, D., 1982. Introduction to Soil Physics. Academic Press, Orlando, FL.

4. Ward, A. D. and Elliot, W.J., 1995. Environmental Hydrology. CRC Press, Inc.