Appendix A: Assumed Water Movement Hydraulics for Modeling BMPs

At minimum, each BMP consists of a reservoir for surface water storage, an overflow outlet and a soil medium. In general, runoff flows into the surface storage reservoir and either infiltrates into the soil or flows through the overflow outlet structure.

Water that does not overflow the surface-storage reservoir infiltrates into the top soil medium and is stored as soil water. Once in the soil, water percolates downward at a rate that is dependent on the soil moisture content, the hydraulic properties of the soil and the boundary conditions of the soil layer.

Many BMPs also include a gravel or aggregate layer below the upper soil layer. Similarly, the rate at which water percolates downward through the gravel/aggregate layer is dependent on the soil moisture content, the hydraulic properties of the soil and the boundary conditions. The lower boundary is often controlled using an underdrain with an orifice outlet.

The following sections describe the theoretical relationships used to develop the FTABLEs for HSPF modeling of the BMPs. The first four sections of this appendix describe the discharge equations used for each of three overflow outlet types and the underdrain orifice:

- Circular Overflow Outlet,
- Straight, Sharp-crested Weir,
- V-notch Weir,
- Underdrain Orifice.

The last three sections describe infiltration, soil water storage and soil water movement.

Circular Overflow Outlet

A circular overflow outlet is basically a vertical pipe with a horizontal opening set to a specific height. This type of outlet is used for bioretention and the flow-through planter BMPs. Hydraulically, this is sufficiently similar to the overflow gate and weir designs shown in the Countywide SUSMP.

Outflow control conditions vary as head over the pipe opening increases. As the water level begins to rise above the opening the pipe acts as a circular weir and flow is crest-controlled. As the head over the opening increases the flow condition transitions to become orifice-controlled and eventually pipe-controlled (the pipe flows full).

Under crest-controlled conditions outflow is calculated using a modified weir equation:

$$Q = C_d \left(2\pi R\right) H^{3/2} \qquad \textbf{Equation 1}$$

Where Q = outflow in cfs, C_d = discharge coefficient, R = pipe radius in ft, and H = the head over the crest in ft.

The discharge coefficient for crest-controlled flow is highly variable depending on the head over the crest, the radius of the circular weir, and the ratio of the inlet height to radius. USBR (1987) published a series of curves that are used to determine the appropriate discharge coefficient for each water surface level.

Straight Sharp-crested Weir

A second type of overflow outlet is a straight sharp-crested weir. A sharp-crested weir is used to control overflow in a vegetated/grassy swale. The following weir equation is used to calculate overflow discharge:

$$Q = C_d L H^{3/2} \quad \textbf{Equation 2}$$

Where Q = outflow in cfs, C_d = discharge coefficient, L = weir length in ft and H = head over the weir crest in ft. The weir coefficient is assumed to be 3.10 for straight sharp-crested weirs.

V-notch Weir

In some cases a v-notch is added to the overflow weir. A v-notch weir is incorporated into the overflow weir of the vegetated bioswale with check dams. The flow through the v-notch is calculated using the following equation.

$$Q = C_d \tan\left(\frac{\phi}{2}\right) H^{5/2} \quad \textbf{Equation 3}$$

Where Q = outflow in cfs, C_d = discharge coefficient, ϕ = angle of the v-notch, and H = head over the weir crest in ft. The v-notch is assumed to be 90 degrees and the weir coefficient was assumed to be 2.55.

Underdrain Outlet

The perforated pipe of lateral underdrains is assumed to be sufficiently large as to not limit the flow into the drain. Drain outflow is limited by single orifice at the end of the drain pipe. Outflow through this orifice was calculated using the orifice equation:

$$Q = C_d A \sqrt{2gH} \quad \textbf{Equation 4}$$

Where Q = outflow, C_d = discharge coefficient, A = area of the orifice, g = gravitational constant, H = head over the centerline of the orifice. The discharge coefficient is assumed to be 0.6 in all cases.

Infiltration

Infiltration is the process of water penetrating from the ground surface into the soil (Chow et al. 1988). Many factors influence the rate of infiltration including ground cover, soil hydraulic properties and soil moisture. As water infiltrates into the soil the soil moisture and hydraulic gradient change. As a result the infiltration rate itself changes over time. This non-linear relation is given by Richard's equation, which is the governing equation for unsteady unsaturated flow in a porous medium. Eagleson (1970) presents Richard's equation in its one-dimensional form:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[D \frac{\partial \theta}{\partial z} + K \right]. \qquad \textbf{Equation 5}$$

Where D = diffusivity, K = hydraulic conductivity, q = soil moisture content, z = elevation and t = time.

Numerous equations have been developed as approximate solutions to Richard's equation. Eagleson (1970) shows that Horton's equation is derived from Richard's equation by assuming D and K are constants independent of soil moisture:

$$f(t) = f_c + (f_0 - f_c)e^{-kt}. \qquad \textbf{Equation 6}$$

Where, f_0 = initial infiltration rate, k = decay constant and f_c = final constant infiltration rate. Using Horton's approximate solution we can see how infiltration rate changes over time.

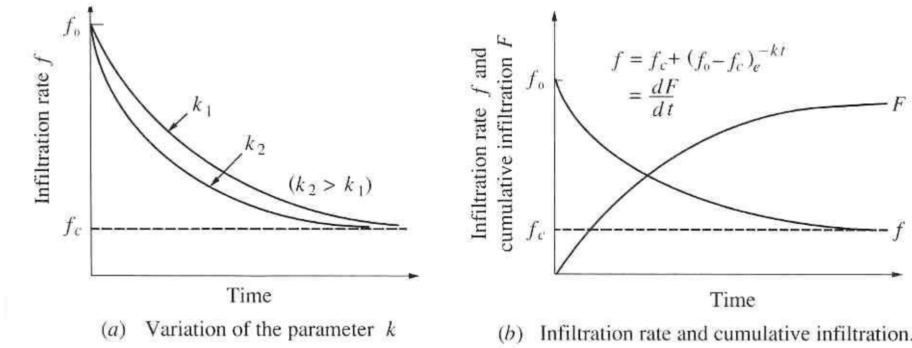


Figure B1– Horton's Equation for Infiltration (graphs from Chow et al. 1988)

We can see from Figure B1 that infiltration begins at a very high rate due to the high matric potential in a dry soil and decreases exponentially as the soil becomes saturated, matric potential becomes insignificant and gravity governs the hydraulic gradient. Thus the infiltration rate approaches a steady-state final rate that approximately corresponds to the saturated hydraulic conductivity of the soil.

After water has been infiltrated into the soil the movement of water through the soil is termed percolation. The rate of percolation can be calculated using Darcy's Law (see Soil Water Movement Section).

Horton's equation showed that the potential infiltration rate of water into the soil always exceeds the saturated hydraulic conductivity of the soil. Conversely, the percolation rate of soil water is limited by the saturated hydraulic conductivity of the soil. Therefore, it is reasonable to assume that the potential infiltration rate is always greater than the percolation rate, and that the percolation rate will limit the flow rate through the soil layer.

Water Storage

The amount of water stored in soils (soil moisture) is expressed as a dimensionless ratio called the volumetric water content, θ. For any given water content the total volume of water stored in the soil, V_{water}, is equal to the volumetric water content (θ) times the total volume of soil, V_{total}.

$$\theta = \frac{V_{water}}{V_{total}} \qquad \textbf{Equation 7}$$

The total void space within a soil is the porosity, η. Soil is saturated when the volumetric water content is equal to the porosity.

Some voids do not actively store and convey water. The void space within the soil that is hydrodynamically effective is called the effective porosity, θ_e. The difference between the total porosity and the effective porosity is known as the residual water content, θ_r. Maidment (1993) provides typical porosity, effective porosity and residual water content values by soil texture (see Table B1).

Table B1– Soil Porosity, Effective Porosity and Residual Water Content by Soil Texture (Maidment, 1993)

Soil Type	Porosity η	Effective Porosity θ_e	Residual Water Content θ_r
GRAVEL[1]	0.420	0.415	0.005
SAND	0.437	0.417	0.020
LOAMY SAND	0.437	0.401	0.035
SANDY LOAM	0.453	0.412	0.041
LOAM	0.463	0.434	0.027
SILT LOAM	0.501	0.486	0.015
SANDY CLAY LOAM	0.398	0.330	0.068
CLAY LOAM	0.464	0.390	0.075
SILTY CLAY LOAM	0.471	0.432	0.040
SANDY CLAY	0.430	0.321	0.109
SILTY CLAY	0.479	0.423	0.056
CLAY	0.475	0.385	0.090

1 – Values for gravel were obtained from Fayer (1992) as presented in INEEL (2002).

Porosity, effective porosity and residual water content values by hydrologic soil group were obtained for this project by assuming each group corresponds with a specific soil texture.

- Group A → Sand
- Group B → Loam
- Group C → Sandy Clay Loam

- Group D → Clay

These assumptions were based on the hydrologic soil group descriptions provided by NRCS (2001). Table B2 provides the assumed porosity, effective porosity and residual water content values by hydrologic soil group.

Table B2 – Soil Porosity, Effective Porosity and Residual Water Content by Hydrologic Soil Group

Soil Type	Porosity η	Effective Porosity θ_e	Residual Water Content θ_r
HYDROLOGIC SOIL GROUP: A	0.437	0.417	0.020
HYDROLOGIC SOIL GROUP: B	0.463	0.434	0.027
HYDROLOGIC SOIL GROUP: C	0.398	0.330	0.068
HYDROLOGIC SOIL GROUP: D	0.475	0.385	0.090

Soil Water Movement

Darcy's Law is used to calculate the rate of water movement through a porous medium:

$$q = -K\frac{\partial h}{\partial z}$$ **Equation 8**

Where q = Darcy flux, K = hydraulic conductivity of the porous medium, h = total hydraulic head, and z = elevation. The total head, h, is the sum of the matric head, ψ, and the gravity head, z (velocity head is negligible):

$$h = \psi + z\ .$$ **Equation 9**

Assuming flow only in the vertical direction and substituting for h, Equation 1 becomes:

$$q = -K\frac{d(\psi + z)}{dz}\ .$$ **Equation 10**

The matric potential within a soil varies greatly with soil moisture. The relation between matric potential and soil moisture for a specific soil is known as the water-retention characteristic of that soil. Figure B2 shows some examples of typical water-retention curves for soils of various textures.

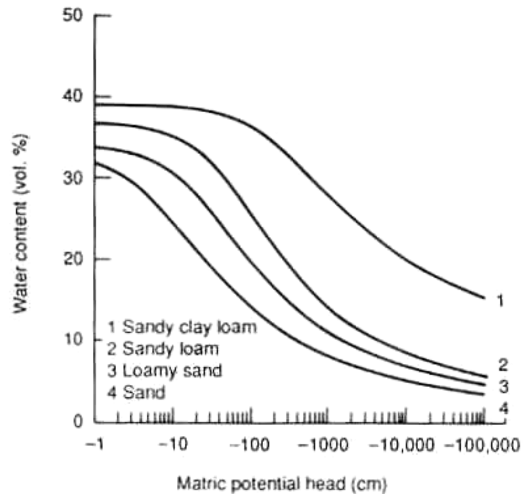


Figure B2 – Typical water retention curves (graph from Maidment, 1993)

Several equations have been developed to approximate water-retention relationships based on the physical characteristics of the soil. One such equation was developed by van Genuchten (1980):

$$\frac{\theta-\theta_r}{\eta-\theta_r}=\left[\frac{1}{1+(\alpha\psi)^n}\right]^m \qquad \textbf{Equation 11}$$

Where the constants α, *n* and *m* are given by:

$$\alpha=(h_b)^{-1} \qquad \textbf{Equation 12}$$

$$n=\lambda+1 \qquad \textbf{Equation 13}$$

$$m=\frac{\lambda}{\lambda+1}\ . \qquad \textbf{Equation 14}$$

The bubbling pressure head, h_b, and pore-size index, λ, are soil-specific parameters. Maidment (1993) provides typical bubbling pressures and pore-size index values by soil texture (see

Table B3).

Table B3 – Bubbling Pressure and
Pore-size Index by Soil Texture (Maidment, 1993)

Soil Type	Bubbling Pressure (cm) h_b	Pore-size Distribution λ
GRAVEL[1]	0.20	1.190
SAND	7.26	0.694
LOAMY SAND	8.69	0.553
SANDY LOAM	14.66	0.378
LOAM	11.15	0.252
SILT LOAM	20.76	0.234
SANDY CLAY LOAM	28.08	0.319
CLAY LOAM	25.89	0.242
SILTY CLAY LOAM	32.56	0.177
SANDY CLAY	29.17	0.223
SILTY CLAY	34.19	0.150
CLAY	37.30	0.165

1 – Values for gravel were obtained from Fayer (1992) as presented in INEEL (2002).

As discussed previously, soil properties were assigned to hydrologic soil groups based on soil textures. Table B4 provides the bubbling pressure and pore-size index values by hydrologic soil group.

Table B4 – **Bubbling Pressure and Pore-size Index by Hydrologic Soil Group**

Soil Type	Bubbling Pressure (cm) h_b	Pore-size Distribution λ
HYDROLOGIC SOIL GROUP: A	7.26	0.694
HYDROLOGIC SOIL GROUP: B	11.15	0.252
HYDROLOGIC SOIL GROUP: C	25.89	0.242
HYDROLOGIC SOIL GROUP: D	37.30	0.165

Hydraulic Conductivity, *K*, is also dependent on soil moisture. Van Genuchten (1980) also developed a relationship to approximate the hydraulic conductivity of soils based on soil properties:

$$\frac{K(\theta)}{K_s} = \left(\frac{\theta-\theta_r}{\eta-\theta_r}\right)^{1/2} \left\{1-\left[1-\left(\frac{\theta-\theta_r}{\eta-\theta_r}\right)^{1/m}\right]^m\right\}^2 . \qquad \textbf{Equation 15}$$

Saturated hydraulic conductivity, K_s, is a measure of a saturated soil's ability to transmit water along a hydraulic gradient. This value is highly variable in field conditions; however, Maidment (1993) does provide estimates of saturated hydraulic conductivity by soil texture (see

Table B5).

Table B5 – Saturated Hydraulic Conductivity by Soil Texture (Maidment, 1993)

Soil Type	Saturated Hydraulic Conductivity (cm/hr) K_s
GRAVEL[1]	1260
SAND	23.56
LOAMY SAND	5.98
SANDY LOAM	2.18
LOAM	1.32
SILT LOAM	0.68
SANDY CLAY LOAM	0.3
CLAY LOAM	0.2
SILTY CLAY LOAM	0.2
SANDY CLAY	0.12
SILTY CLAY	0.1
CLAY	0.06

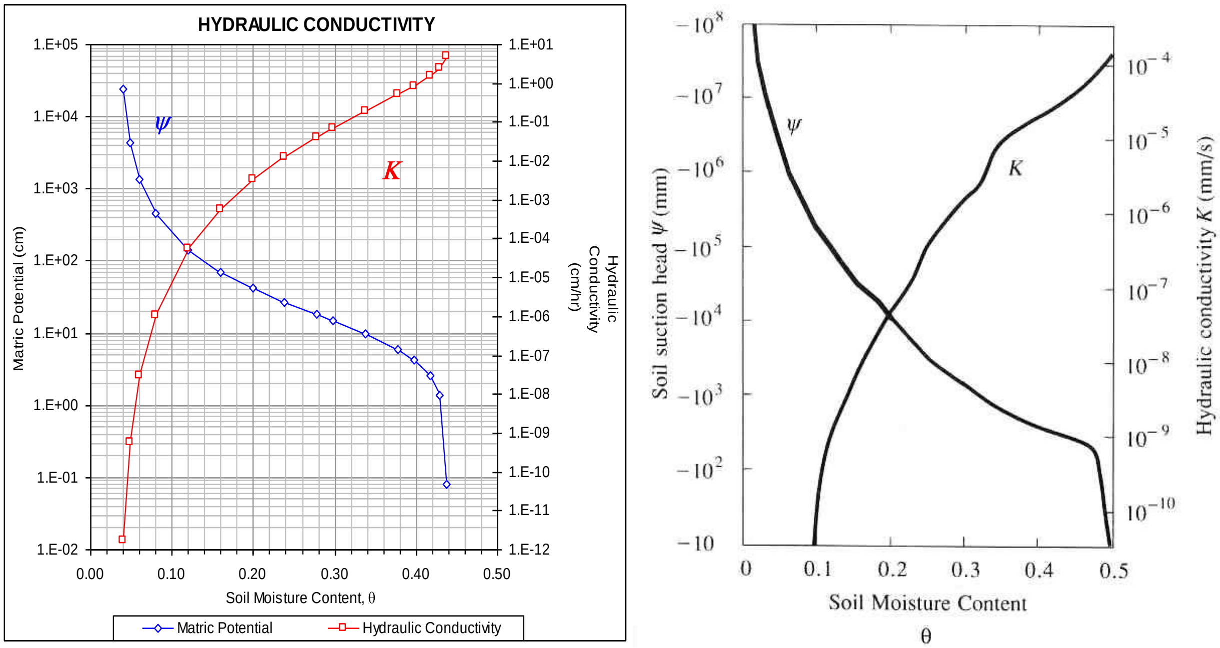
1 – Values for gravel were obtained from Fayer (1992) as presented in INEEL (2002).

As discussed previously, soil properties were assigned to hydrologic soil groups based on soil textures. Table B6 provides the saturated hydraulic conductivity by hydrologic soil group.

Table B6 – Saturated Hydraulic Conductivity by Hydrologic Soil Group

Soil Type	Saturated Hydraulic Conductivity (cm/hr) K_s
HYDROLOGIC SOIL GROUP: A	23.56
HYDROLOGIC SOIL GROUP: B	1.32
HYDROLOGIC SOIL GROUP: C	0.20
HYDROLOGIC SOIL GROUP: D	0.06

Figure B3(a) shows a plot of the van Genuchten relationships using the soil properties assumed for a loamy sand soil. Figure B3(b) is a graph from Chow et al. (1988) that shows the typical variation of matric head and hydraulic conductivity based on experimental data for an example soil.



(a) (b)

Figure B3 - (a) variation of matric head and hydraulic conductivity for a loamy sand using van Genuchten relations, (b) example provided in Chow et al. (1988)

The van Genuchten relations were used to calculate the matric head and hydraulic conductivity for a given soil moisture content. These results were then used in the Darcy equation to compute the flow through the soil. Calculated over a range of soil moisture contents, a table can be created relating soil water storage and flow through the soil layer.

References

1. Chow, V.T.; D.R. Maidment; L.W. Mays. 1988. Applied Hydrology. McGraw-Hill Series in Water Resources and Environmental Engineering. McGraw-Hill, Inc. San Francisco. pp 103.
2. Eagleson, P.S. 1970. Dynamic Hydrology. McGraw-Hill, Inc. pp 291-305.
3. Fayer, M.J.; M.L. Rockhold; and M.D. Campbell. 1992. "Hydrologic modeling of protective barriers: Comparison of field data and simulation results," Soil Science Society of America Journal, Vol 56 pp 690-700.
4. Idaho National Engineering and Environmental Laboratory (INEEL). March 2002. "Engineering Design File: Central Facilities Area Sewage Treatment Drainfield (CFA-08) Protective Cover Infiltration Study." Project File number 021048, Prepared for the U.S. Department of Energy, Idaho Operations Office, Idaho Falls, Idaho. Document Number EDF-2696.

5. Maidment, D.R. 1993. Handbook of Hydrology. McGraw-Hill, Inc. San Francisco. pp 5.6, 5.14, 5.34.
6. National Soil Conservation Service (NRCS). 2001. *National Engineering Handbook, Part 630 Hydrology, Appendix A: Hydrologic Soil Groups*. United States Depertment of Agriculture.
7. United States Bureau of Reclamation (USBR). 1987. Design of Small Dams, Third Edition. United States Department of the Interior, Water Resources Technical Publication. U.S. Government Printing Office, Washington D.C.
8. van Genuchten, M. Th. 1980. "A Closed-form Equation for Predicting the Hydraulic Conductivity of Unsaturated Soils." Soil Science Society of America Journal. Vol. 44, pp 892-898.