Water in soils: Infiltration and redistribution
Overview

• *Infiltration:* The movement of water from the soil surface into the soil

• *Redistribution:* The subsequent movement of infiltrated water in the unsaturated zone of a soil. This can involve exfiltration (evaporation from the upper layer of the soil), capillary rise (movement upward from the saturated zone to the unsaturated zone due to surface tension), recharge (movement of water from the unsaturated zone to the saturated zone) and interflow (flow that moves downslope). Percolation is a general terms for the downward flow in the unsaturated zone.

Infiltration and redistribution depend critically on the material and hydraulic properties of soils

Dingman 2002, Figure 6-1
A soil is a matrix of individual solid grains (mineral or organic) between which are interconnected pore spaces that contain varying fractions of water and air. Most soils have a mixture of grain sizes; the above example is for a soil formed on glacial till (Dingman 2002, Figure 6-2); it is plotted as a cumulative frequency plot of grain diameter. The size of pores through which water flow occurs increases with the size of the grains.
Soil texture triangle

Particle size distribution can be characterized in terms of soil texture, determined by the proportions be weight of clay, silt and sand. The figure at right is the scheme developed by USDA. The texture is determined by the proportions of sand, silt and clay after particles larger than sand (>2 mm or so) are removed. If a significant proportion (>15%) of the sample is of larger size, qualifiers such as “gravelly” or “stony” are added to the soil texture term.

Dingman 2002, Figure 6-3
Soil types: spatial variations

Key determinants of spatial variations in soil types:

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- Climate (P, ET, T)
- Organic material
- Relief
- Parent material (bedrock)
- Time
Material properties of soils

Soils are mixtures of solids, air and water. Key material properties are particle density, bulk density and porosity

1) Particle density (average density of the mineral grains of the soil) \( \rho_m = \frac{M_m}{V_m} \)

\( M_m \) is the mass of mineral grains of the soil sample and \( V_m \) is the volume of the mineral grains of the soil

2) Bulk density: \( \rho_b = \frac{M_m}{V_s} = \frac{M_m}{(V_a + V_w + V_m)} \)

Where \( V_s \) is the total volume of the soil sample, and \( V_a, V_w \) and \( V_m \) are the volumes are the air, water and mineral components of the soil. Typical values are 1000-2000 kg m\(^{-3}\)

3) Porosity (proportion of pore space in a volume of soil): \( \Phi = \frac{(V_a + V_w)}{V_s} = (1- \frac{\rho_b}{\rho_m}) \).

Porosity is typically around 0.5. In many soils porosity decreases with depth due to compaction.
Soil water storage

Volumetric water content (ratio of water volume to total volume):

\[ \theta = \frac{V_w}{V_s} \]

Equivalently

\[ \theta = \frac{(M_{swet} - M_{sdry})}{(\rho_w V_s)} \]

Where \( M_{swet} \) and \( M_{sdry} \) are the wet and dry masses of a soil sample and \( \rho_w \) is the density of water.

Degree of saturation (or wetness), the proportion of pores that contain water, is:

\[ S = \frac{V_w}{(V_a + V_w)} = \frac{\theta}{\Phi} \]

If the soil is saturated, then \( S = 1 \)

Finally, the total amount of water is a layer of soil can be expressed as a depth,

\[ D = \theta T_L \] where \( T_L \) is the thickness of the soil layer
Soil water storage (cont.)

Water is attracted to soil particles. If there is enough water available, water will be held between grains. This is known as capillary water. As the soil dries, the capillary tension increases, the water is held more tightly to grains (the menisci have a smaller radius of curvature – we’ll look at this more a bit later). With further drying, there, capillary water disappears, leaving only a thin film of water held very tightly to grains known as hygroscopic water. The maximum amount of water that a soil can hold after gravitational drainage (which leaves pore spaces) is termed field capacity (θ$_{fc}$). The wilting point (θ$_{pwp}$) is the water content at which plants can no longer extract water from the soil.

http://www.tutorvista.com/topic/soil-pore-space
Soil water storage (cont.)

At shown at left, field capacity and wilting point vary depending on the soil.

Fine grained soils (e.g., clay, made of “platy” minerals) have a high porosity (many small pores) leading to a high field capacity and high wilting point.

By contrast, coarse gained soils such as sand have large pores (but a lower porosity compared to fine grained soils), which allows for lots of gravity drainage and hence a low field capacity. The grains have a low surface area/volume giving a low wilting point.

Silty soils have intermediate wilting points and field capacities.
Soil water flow: Darcy’s Law

Water infiltration can be described by Darcy’s law. The flow rate (m.s\(^{-1}\)) across a unit cross section (m\(^{-2}\)) of soil is:

\[ q_x = -K_h \cdot d(z + p/\gamma_w)dx = -K_h \cdot [dz/dx + d (p/\gamma_w)/dx] \]

\[ q_z = -K_h \cdot d(z + p/\gamma_w)dz = -K_h \cdot [dz/dz + d (p/\gamma_w)/dz] = -K_h \cdot [1 + d (p/\gamma_w)/dz] \]

Where \( q_x \) and \( q_z \) are the flow rates (m.s\(^{-1}\)) in the horizontal (x) and vertical (z) directions (z is taken here to increase upwards), respectively, \(-K_h\) is the hydraulic conductivity (m.s\(^{-1}\)), \( p \) is the soil water pressure (F.m\(^{-2}\), where F is force) and \( \gamma_w \) is the weight density of liquid water (F.m\(^{-3}\)). The weight density of water (kg.a\(_g\).m\(^{-3}\), where a\(_g\) is gravitational acceleration) is approximately a constant.

Darcy’s Law hence states that the flow rate in the horizontal depends on (1) the ability of the soil to “conduct” water; (2) the magnitude and direction of the slope, which is the gravitational potential energy gradient per unit weight of flowing water (soil water wants to flow downhill, towards lower gravitational potential energy), and (3) the magnitude and direction of the horizontal gradient in soil water pressure (if soil water pressure decreases in the x direction, the water wants to flow in that direction). For a given hydraulic conductivity, the vertical flow rate depends on the sum of the gravitational potential energy gradient per unit weight of water (dz/dz =1) and the vertical gradient in soil water pressure.

From here on, we focus on the vertical component of flow.
Soil water flow: Darcy’s Law

From the previous slide, we have:

\[ q_z = -K_h \cdot [1 + d (p/\gamma_w)/dz] \]

Given that \( \gamma_w \) is a constant (we are assuming hydrologic problems with no temperature of salinity gradients that affect water density), one can define a pressure head as

\[ \Psi = p/\gamma_w \]

Which has units of length. Noting that the hydraulic conductivity and the pressure head both depend on the soil water content \( \theta \) (again, be patient) we get:

\[ q_z = -K_h(\theta) \cdot [1 + d \Psi(\theta)/dz] \]

In a steady state (\( q_z \) equals zero), there is a balance between the effect of gravity (which always wants to pull the water “down the gradient” of gravitational potential energy, towards lower \( z \)), and an upward decrease in soil water pressure (“down the gradient” is in the positive \( z \) direction, \( d \Psi(\theta)/dz = -1 \)). If the two effects don’t balance, there is a vertical water flow (up or down, depending on which term “wins”), the magnitude of which depends on the size of the imbalance and the value of the hydraulic conductivity coefficient.
Soil water pressure

Liquid water has a *surface tension* $\sigma$ with dimensions of force per unit length caused by cohesion of molecules. If one immerses a thin (a few mm or less) tube (open at both ends) in a body of water with a free surface at atmospheric pressure, there will be a *capillary rise* due to attraction of the water molecules to the side of the tube by hydrogen bonds. This degree of attraction is reflected in the contact angle $\theta_c$ between the water surface or *meniscus* and the tube. The height of capillary rise is inversely proportional to the radius of the tube, and directly proportional to the surface tension and contact angle:

$$h_{cr} = \frac{2.\sigma \cdot \cos(\theta_c)}{\gamma \cdot r}$$

where $r$ is the radius of the tube and $\gamma$ is the weight density of water.

The water column of height $h_{cr}$ is *suspended from the meniscus*, which is in turn attached to the walls of the tube by hydrogen bonds. The water is said to be under *tension* (related to but not the same as $\sigma$). We can think of the tension in terms of the “desire” of the water to stick to the walls of the tube.

Dingmn 2002, Figure B-9
Soil water pressure

Note that we could also get a rise in the water column in the tube if we were to somehow reduce the pressure above the tube to a value less than the atmospheric pressure at the free surface (there would be less force pushing down on the water in the tube). As such, the rise in the tube due to the capillary effect could also be expressed in terms of a negative pressure.

Going back to the previous equation:

\[ h_{cr} = \frac{2 \sigma \cos(\theta_c)}{\gamma r} \]

Recall that the capillary rise, hence the magnitude of negative pressure, and hence the water tension are inversely proportional to the radius of the tube. Furthermore, as the radius of the tube decrease, so does the radius of curvature \( r_{\text{curve}} \) of the meniscus.
Soil water pressure

Soil water pressure $P$ and hence pressure head $\Psi$ are defined as $<0$ in unsaturated soil, and $>0$ (greater than the atmospheric pressure) in saturated soil. At the water table, $p=0$. Negative pressure $P$ is also called tension or suction. Negative pressure head $\psi$ is also called tension head, matric potential, or matric suction.

The regions between soil grains can viewed as capillary tubes; the water there has tension inversely proportional to the curvature of the menisci. As water content decreases, the water sits deeper in the regions between the soil grains (the capillary tube is effectively narrower) and hence the radius of curvature of the menisci decreases, meaning higher water tension, and the water is held more tightly to the soil grains. A vertical soil water pressure gradient hence promotes a migration of water from higher pressure (where the water is less tightly held to the grains. i.e., lower tension) to lower pressure, where it is held more tightly (higher tension). Hence we can get water flow from suction in an unsaturated soil.

http://www.tutorvista.com/topic/soil-pore-space
The pressure head, hydraulic conductivity and water content

The relationship between the pressure head and volumetric water content $\theta$ for a given soil (the moisture characteristic curve) is nonlinear. This example is for a soil with porosity $\Phi$ of 0.5. At a water content of 0.5 the soil is saturated and the pressure head is hence zero (the pressure is atmospheric). The air tension line point is where significant volumes of air appear in the soil pores. At very high soil tension (strongly negative pressure head or tension head) the curve become nearly vertical, reflecting residual water content held tightly to the soil grains. The hydraulic conductivity in an unsaturated soil is determined largely by the size (cross sectional area) of pathways for water transmission. It increase non-linearly with water content to its saturated value as the water content increases to saturation.
Soil water pressure head by soil type

The pressure head depends on soil type. For a given degree of saturation, the pressure head $\psi$ is much more negative (the tension $|\psi|$ in units used in the figure at left) is much higher in fine-grained soils (e.g., clay) than in coarser-grained soils (e.g., sandy loam). However, the value of tension for a given water content also depends on the history of wetting and drying (hysteresis).
Hydraulic conductivity for different soils

For a given degree of soil saturation, the hydraulic conductivity increases by several orders of magnitude going from clay to silty clay loam to sand. With small pores, water takes a sinuous path through grains (high resistance to flow), with large pores, the path is less sinuous (less resistance to flow) The hysteresis effect for hydraulic conductivity is small and can usually be neglected.

Dingman 2002, Figure 6-10
Hydraulic diffusivity

Relationships between matric suction, volumetric water content and hydraulic conductivity can also be expressed in terms of the hydraulic diffusivity, $D_h(\theta)$:

$$D_h(\theta) = -K_h(\theta).[\partial \Psi(\theta)/\partial \theta]$$

where $D_h(\theta)$ has diffusivity units of m$^2$ s$^{-1}$. To refresh, Darcy’s Law is:

$$q_z = -K_h(\theta).[1 + d \Psi(\theta)/dz]$$

where $q_z$ is the flow rate.

With substitution, Darcy’s Law can hence be written as

$$q_z = -K_h(\theta) - D_h(\theta).[d(\theta)/dz]$$

Stating that the flow due to the pressure gradient is equal to the product of the hydraulic diffusivity and the water content gradient.
If a soil is saturated then the allowed to drain, and assuming no evaporation, plant uptake or capillary rise, the water content will decrease by gravitational drainage indefinitely in a quasi-exponential manner. However, within a few days, the gravitational drainage rate is negligible; at this point, the soil is at field capacity. The figure at left shows the change in water content with time for loamy sand and clay loam; the arrows indicate the value for field capacity. Field capacity values $\theta_{fc}$ may range from 0.1 for sands (which has big pores and drains readily) to more than 0.3 for clays.
Soil water status in natural soils can range from a flooded condition to drainable gravitational water to plant available water down to hydrosopic water. Tensions in natural soils do not exceed -31,000; in this range, water is absorbed from the air.
Soil profiles: Pedalogic horizons

Soil Profile

Most soils have three major horizons -- the surface horizon (A) the subsoil (B), and the substratum (C)

Some soils have an organic horizon (O) on the surface, but this horizon can also be buried.

The master horizon, E, is used for horizons that have a significant loss of minerals (eluviation).

Hard bedrock, which is not soil, uses the letter R.

http://qwickstep.com/search/soil-profile-horizons.html
A soil profile can also be described in terms of hydrologic horizons. The *ground-water zone* (also called the *phreatic zone*) is saturated. The static pressure will increase linearly with depth below the water table as

\[ P(z) = -\rho_w g (z-z_0) \]

Where \( z_0 \) is the value of \( z \) at the water table. There will be hydrostatic equilibrium:

\[ \frac{\partial P}{\partial z} = -\rho_w g \]

Above the water table is a tension-saturated zone (*vadose zone*) where the soil is saturated due to capillary rise. Water enters the *intermediate zone* as infiltration from above (from a precipitation event) and leaves by gravity drainage. Water content may temporarily rise above field capacity. The intermediate zone may extend over many tens of meters (or may be absent in other soil regimes).
Summary 1: Direction of soil water flow

The soil water flow will be upward when:

a) There is an upward decrease in the pressure head that exceeds the downward gravitational potential energy gradient. For a given sum of these two terms of opposing sign, the upward volumetric flow rate per unit cross sectional area (m s\(^{-1}\)) depends on the hydraulic conductivity.

The flow will be downwards when:

b) There is an upward decrease in the pressure head smaller than the downward gravitational potential energy gradient. For a given sum of these two terms of opposing sign, the flow rate (m s\(^{-1}\)) depends on the hydraulic conductivity.

c) There is an upward increase in the pressure head, which will always work in concert with the downward gravitational potential energy gradient. For a given sum of these two terms of the same sign, the flow rate (m s\(^{-1}\)) depends on the hydraulic conductivity.
Summary 2: Hydraulic conductivity and pressure head

The hydraulic conductivity $K_h(\theta)$ depends on:

a) The type of soil, tending to be larger for coarse-grained soils (water path is less sinuous);

b) The volumetric soil water content $\theta$; it increases with increasing soil moisture to its maximum value at saturation (the saturation hydraulic conductivity $K^*_h$).

The pressure head $\psi(\theta)$:

a) Is negative for unsaturated soils, and increasingly negative the drier the soil. It is positive for saturated soils.

b) It depends on soil type. For a given degree of saturation, it is more negative in fine-grained soils than in coarser-grained soils.
Evolution of soil moisture profiles

The figure at left shows changes in the soil moisture profile in clay with a porosity $\Phi = 0.50$ during steady water input rate with no ponding (no water standing on the surface). The numbers on the curve are the number of hours since infiltration began. The initial water content was $\theta = 0.23$. The wetting front, defined as the region with a rapid downward decrease in water content (shown by the dashed lines), is at about 32 cm after 0.83 hours and about 33 cm after 83 hours of steady infiltration. In clay soils (as at left) the wetting front is typically diffuse; it can be sharp for sand. Note in the case at left, the depth of saturated soil increases with time.

- **Infiltration rate**: Rate (m s$^{-1}$) at which water enters the soil from the surface
- **Water input rate**: Rate at which water is added to the surface ($\omega$)
- **Infiltration capacity**: Maximum rate at which infiltration can occur (also called infiltrability)
- **Depth of ponding**: Depth of water standing on the surface.

In the case of no ponding (as at left) the infiltration rate equals the water input rate and is less than or equal to the infiltration capacity. With saturation from above, ponding is present and the water input rate exceeds the infiltration capacity. With saturation from below, ponding is present because the water table has risen to or above the surface; the entire spoil profile is saturated.
Studies of infiltration during water infiltration events from above show high infiltration rates at the beginning of the event, followed by a relatively rapid decline that transitions towards near-constant values. The figure at left shows infiltration rates from laboratory studies into grassed loam plots using artificial rainstorm events of 15 minutes duration and of varying intensity.
Factors affecting the infiltration rate

- Water input rate from rainfall, snowfall or irrigation
- The saturated hydraulic conductivity of the soil profile
- The degree to which pores are already filled with water when the infiltration process begins
- Variations in hydraulic conductivity through the soil profile
- The inclination and roughness of the soil surface
- Chemical characteristics of the soil surface
- Physical and chemical properties of water

- Organic surface layers - leaf litter, etc, has large openings
- Frost – a frozen surface can be nearly impermeable
- Rain compaction of the soil
- In-washing of fine sediments carries into larger pores
- Human modification of the soil surface
- Swelling and drying - some soils like vertisols, have clay minerals that swell land shrink with wetting and drying. Swelling can reduce the effective surface porosity. During dry periods, surface cracks that develop can accept high infiltration rates.
The Richards Equation

The Richards equation combines Darcy’s Law for vertical unsaturated flow with the conservation of mass. It is widely used as a basis for numerical modeling soil water flow by specifying appropriate boundary conditions, dividing the soil profile into very thin layers, and applying the equation to each later sequentially over small increments of time.

\[
\frac{\partial \theta}{\partial t} = -\frac{\partial K_h(\theta)}{\partial z} + \frac{\partial}{\partial z} [K_h(\theta) \frac{\partial \psi(\theta)}{\partial z}]
\]

Expressed verbally, the time rate of change in volumetric soil moisture for a given thin layer of soil depends on the vertical rate of change of the hydraulic conductivity (itself a function of \( \theta \)) and the vertical rate of change of the product of (a) the hydraulic conductivity, and (b) the vertical rate of change of the pressure head \( \psi \), (the matric suction gradient) the pressure head also being a function of \( \theta \). In this expression, \( z \) is taken to increase downwards.
The Green-and-Ampt Model

Like the Richards equation, the Green-and-Ampt model (named for its originators, way back in 1911) applies Darcy’s Law and the principle of conservation of mass.

Consider two experiments with this model with highly idealized initial conditions. In each experiment, we start with a homogeneous block of soil of indefinite depth (porosity $\Phi$ and saturated hydraulic conductivity $K_h^*$ are constants and there is no water table, capillary fringe or impermeable layer), with no surface slope and no evapotranspiration. The soil water content at the time $t=0$ ($\theta_0$) is constant through the block at an initial value $\theta < \Phi$.

Beginning at time $t=0$, water (rain or snowmelt) is poured on top of the block at a specified input rate $w$ and continues for a specified time $t_w$. 
Experiment 1: Water input rate is greater than $K_h(\theta_0)$ and less than the saturated hydraulic conductivity $K_h^*$.

Water enters a thin layer of soil (call it layer “A”) faster than it is leaving; increasing the water content as well as the hydraulic conductivity and the soil water content gradient (the matric suction gradient). Hence the flux out of layer “A” into the layer below (layer “B”) also increases. However, as long as the downward flux to layer “B” is less than the input water rate, the water content of layer “A” continues to increase. When the water content of layer “A” reaches $\theta_w$ the rate of outflow from the layer equals the rate in inflow to the layer ($q_z = w$) and there is no change in water content until water input ceases. This process happens successively at each layer (layer “B”, “C”, “D”, etc) giving a sharp, descending wetting front. As the wetting front descends, the suction gradient decreases, so that the rate of downward flow approaches $K_h(\theta_w)$. The pattern of successive water content profiles is shown in the top panel of the figure at left; the bottom panel is the graph of infiltration rate versus time.

Dingman 2002, Figure 6-24
Experiment 2: Water input rate greater than saturated hydraulic conductivity $K_h^*$

When $w > K_h^*$, the initial processes is as described previously - water will arrive at layer “A” faster than it can be transmitted downward, and goes into raising the water content of the layer. However, the water content cannot exceed its value at saturation and the hydraulic conductivity cannot exceed its saturation value. Once layer “A” reaches saturation, the wetting front begins to descend downwards. The matric suction force decreases as the wetting front descends, but the downward water flux in this case approaches $K_h^*$.

The pattern of successive water content profiles is shown in the top panel of the figure at left; the bottom panel is the graph of infiltration rate versus time. Since $w > K_h^*$, some rain continues to infiltrate after the surface reaches saturation, but the excess accumulates at the surface as ponding.
Summary 3: Evolution (time change) of volumetric soil moisture at given point in a soil profile

The soil moisture ($\theta$) will increase with time if (for example):

a) The water flow is *downward* and the hydraulic conductivity $K_h(\theta)$ decreases *downwards*.

b) The water flow is *upwards* and the hydraulic conductivity $K_h(\theta)$ decreases *upwards*.

c) The water flow is *upward*, there is an upward decrease (gradient) in the pressure head $\psi(\theta)$, but the upward decrease in the pressure head becomes less steep the higher up you go (the upward suction above the point is less than the upward suction below the point, hence water accumulates).

d) The water flow is *downward*, there is an upward decrease in the pressure head $\psi(\theta)$, but the upward decrease in the pressure head becomes steeper the higher up you go (water moves downwards but the opposing force on the downward flow due to the upward pressure gradient gets weaker, so water accumulates).

One can develop similar relationships for a decrease in soil moisture with time.
Ponding and overland flow

Following from the previous slides, if the water input rate exceeds the hydraulic conductivity, some rain continues to infiltrate after the surface reaches saturation, but the excess accumulates at the surface as ponding. The instant when the surface layer $z_f$ becomes saturated is called the time of ponding $t_p$:

$$t_p = \frac{[K_h^*|\psi_f|.(\Phi - \theta_0)]/(w.(w-K_h^*))}{w}$$

Where $\psi_f$ is the effective tension at the wetting front. This makes sense; $t_p$ increases with increasing $K_h^*|\psi_f|$ and the initial soil water deficit $\Phi - \theta_0$ and decreases with an increasing water input rate $w$. If there is a slope, the excess water moves downslope as overland flow, also termed surface runoff.

http://www.ceg.ncl.ac.uk/thefarm/
The figures at right show the infiltration rate by time using the Green-and-Ampt model (symbols) and the Richards Equation (lines) showing the effects of (top) different initial water content $\theta_0$ for an identical water input rate; (bottom) different water input rates for the identical initial soil moisture content $\theta_0$. The time of ponding is the time when the infiltration rate starts to decline.

Dingman 2002, Figure 6-27
There can be a high degree of spatial variability in even a small watershed in infiltrability and the factors determining infiltration. This reflects many things, including variations in soil properties, vegetation cover, land use (parking lots versus forest) and animal activity (e.g., burrowing rodents). Values for porosity, saturation hydraulic conductivity and other parameters determined at a point may not transfer well to larger scales.
Fire and floods

The Hayman Fire was a forest fire that started 95 miles southwest of Denver on June 8, 2002, and became the largest fire in the state's recorded history. It has had lasting effects on the regional hydrology.

http://www.kirkhanes.com/other/haymanfire.htm
Risks associated with wildfire

• Increased flooding
• Increased erosion

Why?

1) Lack of vegetation: This seems to be most important (direct raindrop impacts as opposed to interception by vegetation, lots of loose sediment).
2) Hydrophobic soils: The fire vaporizes compounds in organic material which condensed on soil particles. Infiltration is reduced.