concentrations and exhibit more rapid growth than trees grown on adjacent sites without irrigation.

Summary

1. The concept of land treatment of wastewater has a sound scientific and experience foundation which has proven that land can be used to renovate wastewater in a environmentally acceptable manner and that such land is not irreversibly withdrawn from any present or future societal use.
2. No human or animal health problems have been reported and studies have concluded that properly designed and operated wastewater irrigation systems are likely to pose less environmental health problems than most other wastewater treatment technologies.
3. Forests can be successfully used as the principal vegetative cover in a land treatment system. It has in fact a number of advantages over agronomic crops including greater flexibility to operate around climatic conditions, fewer interruptions to the irrigation schedule, and can be operated year-round.
4. The design of a forest system must be based on potential performance of the site to meet water quality performance criteria objectives including hydraulic capacity as well as nitrogen assimilative capacity. Both of these factors normally influence the total performance of the land treatment system.
5. Successful operation of the land treatment system is evaluated on the basis of performance standards established by water quality objectives.

See also: **Hydrology**: Impacts of Forest Conversion on Streamflow. **Silviculture**: Forest Rehabilitation. **Site-Specific Silviculture**: Silviculture in Polluted Areas. **Soil Development and Properties**: Water Storage and Movement. **Tree Breeding, Practices**: Nitrogen-fixing Tree Improvement and Culture.

Further Reading

Cole DW, Henry CL, and Nutter WL (eds) (1986) *The Forest Alternative for Treatment and Utilization of Municipal and Industrial Wastes*. Seattle, WA: University of Washington Press.

Hamsley WD (1979) *Estimation of Soil Phosphorus Fixation Capacity on a Forested Wastewater Land Treatment Site*. M.Sc. thesis, University of Georgia School of Forest Resources.

Henry CL, Harrison RB, and Bastian RK (eds) (2000) *The Forest Alternative: Principles and Practice of Residuals Use*, Proceedings of a Conference held at University of Washington, Seattle, July 14–16, 1997.

Iskander IK (1981) *Modeling Wastewater Renovation Land Treatment*. New York: John Wiley.

McKim HL, Sopper WE, Cole D, *et al*. (1982) *Wastewater Application in Forest Ecosystems*. CRREL Report no. 80-19. Hanover, NH: US Army Cold Regions Research and Engineering Laboratory.

Moffat AJ, Armstrong AT, and Ockleston J (2001) The optimization of sewage sludge and effluent disposal on energy crops of short rotation hybrid poplar. *Biomass and Bioenergy* 20(3): 161–169.

Nutter WL (1986) Forest land treatment of wastewater in Clayton County, Georgia: a case study. In: *The Forest Alternative for Treatment and Utilization of Municipal and Industrial Wastes*, pp. 393–405. Seattle, WA: University of Washington Press.

Nutter WL (2000) Implementation and operation of wastewater irrigation systems in forests. In: *The Forest Alternative: Principles and Practice of Residuals Use*, pp. 135–138. Proceedings of a Conference held at University of Washington, Seattle, July 14–16, 1997.

Nutter WL, Schultz RC, and Brister GH (1979) Renovation of municipal wastewater by spray irrigation on steep forest slopes in the Southern Appalachians. In: *Utilization of Municipal Sewage Effluent and Sludge on Forest and Disturbed Land*, pp. 77–85. Philadelphia, PA: Pennsylvania State University Press.

Overcash MR and Pal D (1979) *Design of Land Treatment Systems for Industrial Wastes: Theory and Practice*. Ann Arbor, MI: Ann Arbor Science.

Reed SC, Middlebrooks EJ, and Crites RW (1988) *Natural Systems for Waste Management and Treatment*. New York: McGraw-Hill.

US Environmental Protection Agency (1981) *Process Design Manual for Land Treatment of Municipal Wastewater*. EPA 625/1-81-013 (COE EM1110-1-501). Washington, DC: US Government Printing Office.

Water Storage and Movement

M Weiler, University of British Columbia, Vancouver, BC, Canada
J J McDonnell, Oregon State University, Corvallis, OR, USA

Introduction

Water storage and movement in forest soils is a key regulator for a variety of hydrological, physiological, and biogeochemical processes in a forest. The climate and geology controls on soils vary around the world; these can range from conditions of colluvial infilling of steep unstable hollows in and around the Pacific Rim, to till soils that develop on recently glaciated sites in Scandinavia, eastern Canada, and Russia, and deeper clay-rich soils in lower-latitude regions. While the physics of flow in porous media is the same regardless of land-use type, forest soils often have rather different depth-integrated and spatially variable properties relative to soils in agricultural or suburban areas. This entry considers a number of properties of forest soils as it relates to some of the basic definitions and physical processes governing water storage and movement. It then considers in detail the main processes of how water moves vertically and laterally through forest soils – from the plot scale, to hillslope scale, and catchment scale. Finally, influences of forest management and forest fires on water storage and movement are discussed. This entry focuses on some of the first-order controls common to most landscapes. The reader should consult material provided in the Further Reading section for a comprehensive review of water storage and movement in all climatic and physiographic regions of the world.

Soil Physics Terminology and Measurement

Soils, especially forest soils, are a complex mixture of organic and inorganic, living and dead, or solid, liquid, and gaseous materials. This complexity appears at first glance difficult to characterize; however, classifying soils into three phases – air, water, and solids – provides a convenient means to define the basic physical soil properties (**Figure 1**).

The particle density is equal to the ratio of mass to volume of solids. In contrast, the bulk density is the ratio of mass of solids to the total volume. The porosity is the proportion of pore space (air and water) in a given volume of soil. The water content of the soil is described in two ways, as the ratio of water to soil volume, if it is volumetrically defined, or mass, if it is gravitationally defined. Given the water content, the degree of saturation can be calculated as the ratio between volumetric water content and porosity. The water content is an important hydrological property of soils. It can be determined in the laboratory or using field methods. In the laboratory, one weighs a field-extracted intact core of known volume, dries it at 105°C for 24 h, and then weighs it again. The difference is used to compute the volumetric water content. In the field, water content is most often measured by time–domain reflectometry (TDR), although many investigators still use neutron probes, gypsum blocks, and capacitance techniques. TDR instruments operate by measuring the propagation velocity in the soil of an electric pulse that is related to the dielectric permittivity or dielectric constant, which is closely related to water content.

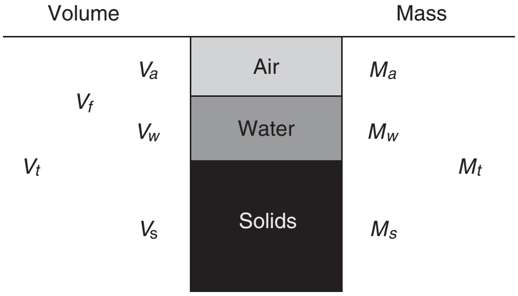

Figure 1 Soil and its three phases: air, water, and solids. V_t and M_t are totals.

Energy State of Water in Soil

Knowing the temporal and spatial variation of water content in the soil is sufficient for determining the total soil water storage (usually expressed as a depth or volume per unit area). For measuring and defining the direction of water movement in soils, the energy state of soil water must be defined since differences in the energy state (potential) drive the direction of water movement. The total soil water potential is the sum of various forces acting on the soil water: gravitational potential, pressure potential, matric potential, and osmotic potential. The gravitational potential depends on the position in the gravitational force field relative to some reference level. Pressure potential is the hydrostatic pressure of the water column under saturation. The matric potential (also referred to as matric suction or capillary potential) is defined under unsaturated conditions, where capillary and adsorptive forces act to create a negative pressure (often called tension or suction). This is measured in soils relative to the external gas pressure. The osmotic potential is attributed to the presence of solutes in the soil water and only in arid environments significantly affects the water movement compared to pressure and matric potential. The primary effect of osmotic potential relates to the uptake of water by vegetation. In this case, the roots act as a membrane which regulates the movement of water as a function of osmotic potential, since water vapor pressure is lowered by the presence of solutes. The pressure potential of soil water is often measured with piezometers. Piezometers are tubes augered into the soil below the water table and only open at the bottom of the tube. Thus, the pressure potential at the bottom of the tube is reflected by the height of water rise in the standing water column within the tube. The soil matric potential is measured with porous cup devices called tensiometers that have a practical range from 0 to 800 cm (0–78 kPa). This range covers most of the naturally observed tensions in forest soils. For higher tensions, thermocouples, psychrometers, or gypsum electrical-resistance blocks are available.

Soil Moisture Characteristic Curve

The relationship between soil moisture and tension is called soil moisture characteristic curve (also soil moisture release curve, retention curve) and is an important property of unsaturated soil. This relationship strongly depends on the soil texture, but also on other soil properties like soil structure, organic matter, and bulk density. No universal theory yet exists to describe or predict the soil moisture characteristic curve (SMCC) from soil properties. In addition, the value of tension at a given water content is not unique, but depends on the soil history of wetting and drying. This hysteresis can have a significant influence on water movement, but is often not considered in describing water movement in hydrological models, since no unique functional relationship can be easily assumed.

The SMCC of forest soils is often highly nonlinear. **Figure 2** shows drying curves for forest soils in old-growth Douglas–fir at three soil depths. The water content decreases by 10–30% between saturation and 20–40 cm of tension. This typical 'drop' for many forest soils is related to soil structure and the macroporosity (large pore space) related to the effect of roots, especially in the topsoil. These macropores in the upper soil horizons drain water at a very low tension (low capillary). This nonlinearity often declines with soil depth (**Figure 2**) since the pore space often declines rapidly into the profile, with concomitant increases in bulk density.

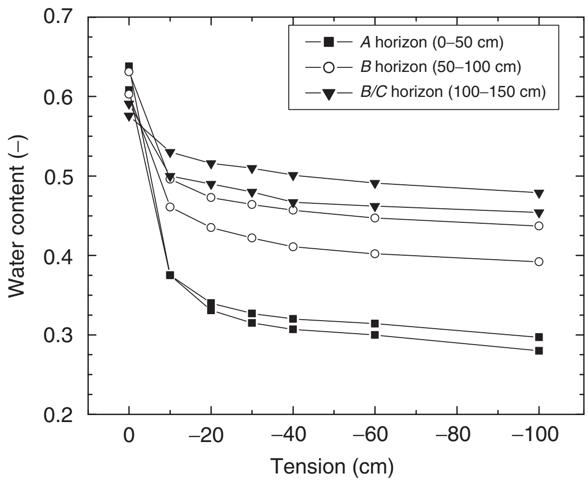

Figure 2 Soil moisture characteristic curves at three different depths of a forest soil (data from HJ Andrews Experimental Forest, Oregon, USA). Note the flattening of the curves deeper in the profile. These curves represent the trajectory of water content and water suction upon drying. Often, soils show differences in these curves during wetting. This difference, called hysteresis, can be significant for some forest soils.

Field Capacity and Permanent Wilting Point

Two points of the SMCC are particularly important: the field capacity and the permanent wilting point. Field capacity is the water content of a soil after gravitational drainage over approximately a day. The suction that defines this value varies from soil to soil, but is generally in the range of 10–33 kPa. Drainable porosity of a soil is defined as the water content between field capacity and saturation. The drainable porosity controls the transient water-table dynamic that often develops at the soil–bedrock interface or some zone of low permeability at depth (hardpan, duricrust, or other layer). The permanent wilting point is the water content at which plants start to wilt during daytime – indicating that they are no longer able to extract water from the soil. The suction at this point is very high, about 1470 kPa. The difference of water content between field capacity and permanent wilting point is often called the available water content.

Water Storage

Water storage in soils depends on the water balance of a soil pedon. The water balance represents one of the most basic equations in hydrology. The change in water storage is equal to the changes in input and output (**Figure 3**). The principal input flux in forest

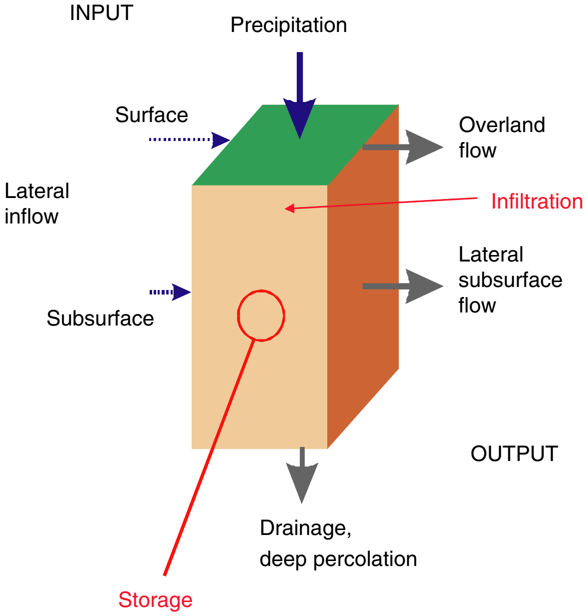

Figure 3 Water storage, input, and output fluxes for a soil column.

soils is precipitation. Only a proportion of total precipitation, termed throughfall, reaches the forest floor. The difference between total amount of precipitation and throughfall results from interception and storage by the forest canopy. Throughfall or snowmelt from snow accumulated above the forest floor infiltrates into the soil, flows over the soil surface, or is stored within surface depressions. While forest soil infiltration rates almost always exceed the precipitation input to the soil surface, ‘excess water’ may flow on the soil or decayed-leaf surfaces and become stored in small microtopographic depressions in the forest soil surface. Vertical inflow to the soil pedon may therefore originate as ponded overland flow or by lateral subsurface flow.

Three mechanisms act to deplete water in the soil pedon. The first includes direct evaporation from the soil surface and plant uptake in the rooting zone and then transpiration by plants. This loss is collectively termed evapotranspiration. The second is the vertical drainage into the underlying bedrock and possible recharge into underlying groundwater. The third mechanism is lateral subsurface flow within the soil. The dynamics of these fluxes in forest soils are unique as compared to soils under other land use. The modulation of the incoming precipitation by the canopy results in a reduction of both the rainfall intensity and the total amount of precipitation. In addition to trees, smaller plants (e.g., shrubs and bushes) as well as moss and the litter layer affect the disposition of incoming precipitation. Direct evaporation rates from the soil surface are generally lower than the transpiration rates by plants in forested settings. Since trees extract water from soil and have a deep root zone, the reduction of the water content with depth during dry periods is much more sustained in forest soils compared to soils under other land use. Alternatively, direct evaporation of water at the soil surface is reduced under the forest canopy. Water redistribution in forest soils by tree roots, especially by water uptake from the saturated zone (groundwater), and redistribution into the unsaturated zone is an important process, particularly in semi-arid climates.

Spatial Variability

The heterogeneous structure of forests, in combination with complex topography and soil heterogeneity, results in significant spatial variability of water storage within forest soils. The plant canopy modifies precipitation input and also produces a persistent spatial pattern of throughfall to the soil surface. Studies under coniferous canopies have shown up to 100% differences in throughfall application to the soil surface over distances of less than 1 m. In addition, flow down the tree stem (stem flow) further increases this variability of precipitation input to the soil. Snowmelt may produce a similar spatial variability through factors affecting the energy budget of the snow (e.g., slope aspect, inclination, and cover type). Lateral flow of water within the soil or upwards movement of groundwater into the soil profile is often described by topography. Thus, topographic position, local slope angle, and upslope contributing area are key variables to explain larger-scale spatial variability of soil water storage. These influences are most pronounced in areas with significant topographic relief, shallow lateral flow pathways, and under humid conditions. Finally, the variability of transpiration by trees and other plants may also affect a spatial variability of soil water storage within forests.

Water Movement

Hydrological Concepts – Runoff Generation Processes

Runoff generation processes during rainfall or snowmelt events are often separated into two classes: those processes which were relevant for generating overland and those processes that are relevant for generating lateral subsurface flow. Overland flow can be generated by infiltration excess (rainfall intensity is larger than the soil’s infiltration capacity) or by saturation excess (where soils become saturated by a rising water table). In forest soils, overland flow is usually generated by the saturation excess mechanism. One exception is where infiltration excess overland flow may be produced on logging roads and other low-permeability areas (e.g., compacted soils) or on soils with seasonal water repellency due to fire (*see* **Health and Protection**: Forest Fires (Prediction, Prevention, Preparedness and Suppression)). Saturation overland flow is most common in areas where soils are often waterlogged (topographic confluence zones, near springs, and in riparian zones).

On steeper hillslopes, water infiltrating into the soil will either be stored in the soil or will continue moving vertically to recharge local groundwater or flow as lateral subsurface flow. This lateral subsurface flow (also called subsurface stormflow, interflow, and throughflow) are very common in forest soils since the lateral hydraulic conductivity and the gravitational gradients (in areas with a steep relief) are often high, and additional preferential flow pathways are present to enhance the downslope flow. Knowing the dominant runoff generation processes at a site is an important first step to understand runoff generation in a catchment, as well as flood generation, nutrient transport, and prediction of forest management practices on water quantity and quality.

Vertical Movement

Water movement in porous media like soils is often described based on the Darcy equation that flow is proportional to the hydraulic potential times the hydraulic conductivity:

$$q = -K\frac{\Delta H}{\Delta z}$$

where q is the water flow (length per time), K is the saturated hydraulic conductivity (length per time), ΔH is the hydraulic head difference, and Δz the distance. The saturated hydraulic conductivity (or permeability) is a spatially variable property of soils (over several orders of magnitude over short distances). Hence, it varies with the scale of measurement. It depends on the soil texture, but also on the soil structural features. Due to these structures, the saturated hydraulic conductivity of forest soils is usually much smaller in the vertical direction than in the horizontal direction (known as anisotropy). In addition, the hydraulic conductivity depends strongly on the degree of saturation and thus on the soil water tension. Based on this functional relationship, the Richards equation was developed by combining the Darcy equation with the continuity equation to describe flow in unsaturated porous media:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[K(h)\left(\frac{\partial h}{\partial z}+1\right)\right]$$

where $K(h)$ is the unsaturated hydraulic conductivity and θ the water content. Both equations are based on the capillary concept and work well in relatively homogeneous soils. However, in forest soils, the influence of plant roots, soil structure, burrowing animals, and worm casts creates a variety of larger pores (macropores) that water may follow preferentially (**Figure 4**). Due to these macropores, the unsaturated hydraulic conductivity relationship often shows a significant reduction (one to some orders of magnitude) at values near saturation. Thus, while macropores may comprise only a small part of the total soil porosity, they may control almost all the water flow at or near saturation within the profile. The water flow in these macropores is often turbulent and mostly driven by gravity. The resulting water movement in the soils is very heterogeneous and dry areas within the soil may be bypassed. These processes run counter to the Darcy and Richards formulations that rely exclusively on capillary-driven laws of fluid flow.

Water movement that may be influenced by preferential flow can be visualized by adding a dye tracer to the infiltrating water. **Figure 5** shows some experimental results where a food dye (Brilliant Blue FCF) was added to a simulated rainfall event. Soil profiles were excavated and pictures were taken from these vertical soil sections. **Figure 5** shows two examples of these dye patterns for forest soils in Oregon, USA. The patterns show that the soil surface itself may affect the disposition of infiltrating rainfall. The litter layer may be a significant generator of flowpath heterogeneity near the soil surface. Deeper in the soil profile, water flow may occur only in macropores, bypassing large dry areas of matrix in the soil profile. In contrast to homogeneous infiltration, the process of water flow into the macropores (initiation) and water flow from macropores into the surrounding soil matrix (interaction) mainly controls the vertical water movement of water in forest soils. In general, macropore flow results in a much faster flow and increased transport of solutes and nutrients.

Lateral Movement

Lateral water movement in the soil is an important process for redistribution of water, nutrients, and solutes in the environment. This process also controls the generation of storm runoff in many upland forested environments. Detailed process studies in forest soils in the last half-century have revealed a variety of flow pathways in forest watersheds. **Figure 4** illustrates the most important of these pathways. If the bedrock is relatively impermeable compared to the soil, infiltrating water perches on the soil–bedrock interface and flows laterally downslope along this interface. Since this interface is generally topographically 'rough' due to weathering and mass movement, water concentrates in hollows and depressions. The resulting channalized flow acts similarly to macropore flow whereas the average flow velocity increases and areas with a relatively higher soil–bedrock relief interface are bypassed. The lateral flow is less preferential if the soil–bedrock interface is more gradual in texture and where the hydraulic conductivity decrease with depth is more gradual. This gradual decline can be observed in soils developed from glaciated deposits (e.g. till).

Macropore flow is also a major control on lateral flow on forest hillslopes. Similar to the processes governing vertical flow in macropores, laterally oriented macropore flow may dominate in many forest environments where macropores are generated by plant roots and burrowing animals. These

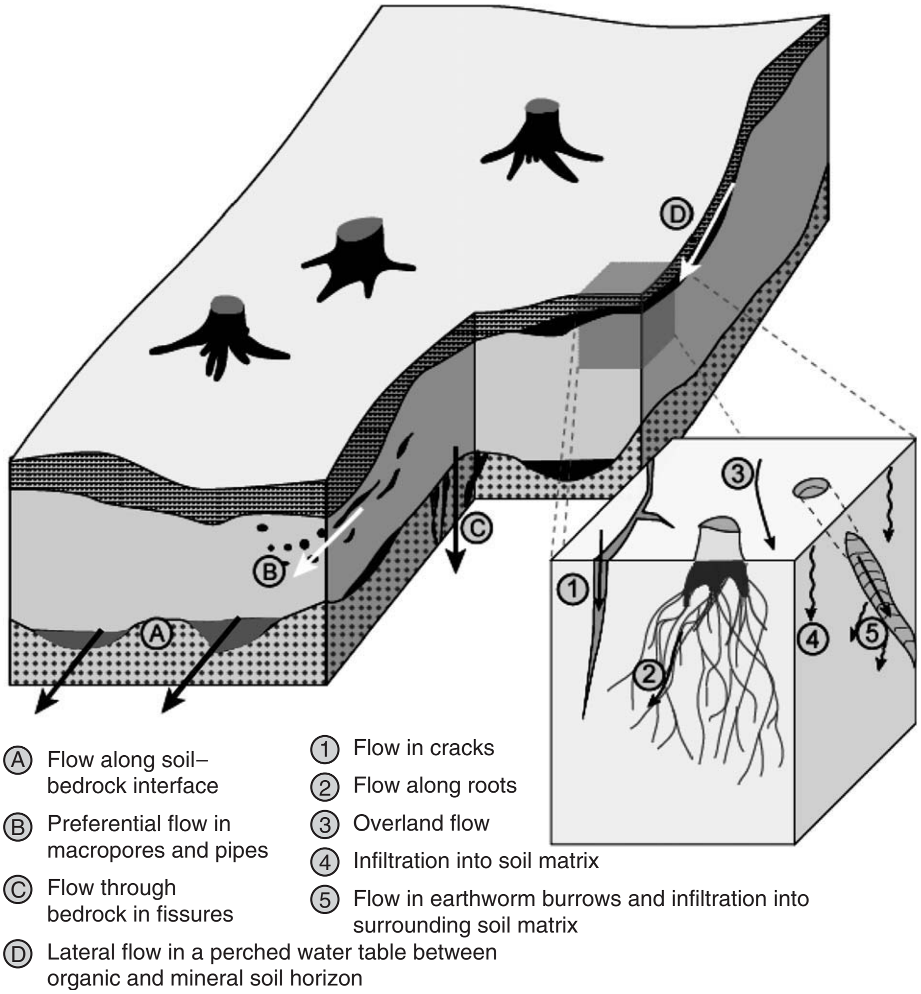

Figure 4 Water movement in forest soils: lateral flow pathways (A–D) and vertical flow pathways (1–5). Note the complexity of water flow pathways due to physical and biological agents acting on soil.

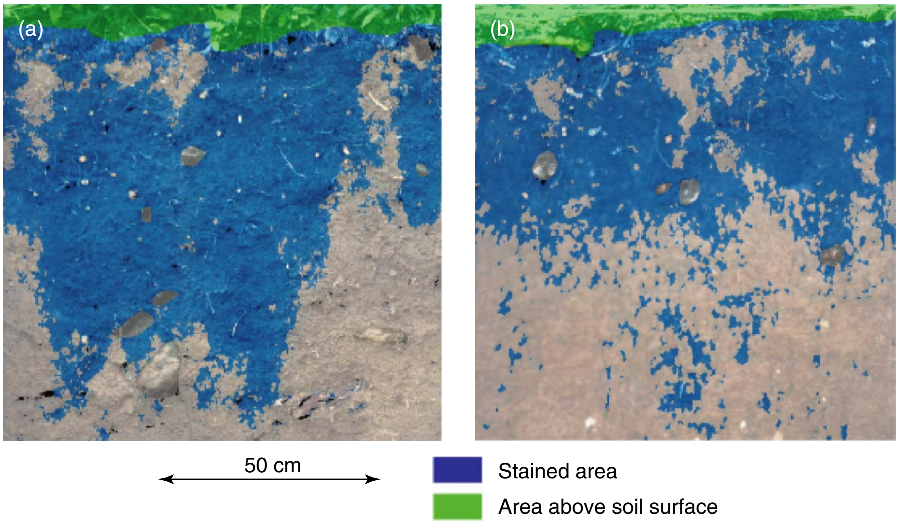

Figure 5 Dye patterns from two different forested sites: (a) HJ Andrews Experimental Forest, Oregon, USA; (b) Low Pass Area, Coastal Range, Oregon, USA. Note the spatial heterogeneity of dyed water and preferred nature of water flow vertically within the forest soil profile.

macropores are often termed soil pipes. If a connected network is developed due to internal erosion and connection of the macropores, piping can provide effective drainage augmentation to hillslopes. If the underlying bedrock is more permeable, water can infiltrate into the bedrock and then percolate vertically in fissures and cracks. The top soil layer (litter or organic layer) in forest soils includes a high proportion of organic material and roots that enhance the hydraulic conductivity. If the underlying mineral soil has a low hydraulic conductivity, a perched water table can sometimes develop during rainfall and snowmelt events within the organic horizon. Under these conditions, water may flow laterally within this layer.

Important Issues

Harvesting and Forest Management

Harvesting and forest management operations can strongly influence water storage and movement in soil. An important issue is soil compaction by using heavy machinery for timber harvesting. Depending on the soil texture, soil structure, and the soil moisture content during the operations, soil compaction can occur. Soil compaction is defined as an increase in bulk density and a decrease in soil porosity resulting from applied loads, vibration, or pressure. Compaction can reduce infiltration, leading to the development of areas that can produce infiltration excess and overland flow. Water storage capacity of the soil can also be reduced since the proportion of larger pore space may be reduced through compression. A specific feature of forest soils is the organic layer on top of the mineral soil. This organic layer modulates infiltration, reduces evaporation, and increases water storage. Removing this organic layer by forest management (mechanical removing and burning) often negatively changes the hydrological behavior of a site. While much progress has been made in forest operations in the developed world in reducing site disturbance associated with logging, poor logging practices can have deleterious effects on soil water storage and movement.

Fire

Forest fires may consume a large part of the forest floor, eliminating beneficial effects of the organic layer on soil properties. However, the effects of low-temperature fires (e.g., from controlled burning) are generally less than the effects from hot wildfires. The most pronounced impact of forest fires on soil properties is the reduction of infiltration rates due to water repellency. This reduced infiltration not only increases the amount of overland flow and thus soil erosion, but may further reduce the availability of soil water for plants, especially in semiarid regions. Water-repellent soils can develop from hydrophobic substances vaporized during burning of the surface litter layers. The degree and persistence of these hydrophobic substances depend not only on the temperatures during the fire but also on the timing, number, and magnitude of subsequent rainfall events on the burned site (as water repellency can decrease with time). In addition, drier soils show higher water repellency than wetter soils, which should be kept in mind considering these effects in different climate zones.

Measuring the effects of fire on water movement and storage in soils is difficult, and results from experimental studies concerning the effects of fire on changes in infiltration rate are equivocal. The major problem is to determine if small-scale effects due to hydrophobic substances translate into larger-scale (hillslope and watershed) behavioral changes. One possibility to explore the effects of forest fire on infiltration is to perform infiltration experiments in combination with dye patterns. **Figure 6** shows the impact of the water-repellent surface layer. Water was only able to infiltrate at few locations within the profile, coincident with local depressions or plant roots. Thus, the resulting dye pattern shows only a thin staining near the soil surface and some isolated 'spots' within the soil. Nevertheless, comprehensive analysis of the effects of fire on water storage and movement in forest soils on larger scales remain the topic of future research.

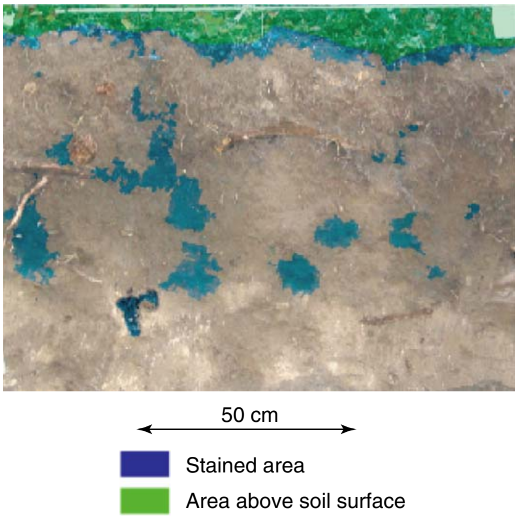

Figure 6 Dye pattern from a recently burned forest soil in the Western Cascades, Oregon, USA.

Summary

Water movement and storage in soils are regulated by a variety of temporally and spatially variable processes. This article presents an overview of the most important properties and processes influencing water movement and storage in forest soils. Soil water storage and movement are controlled by the size and spatial distribution of macropores, through which water can move rapidly but which drain under gravity, and micropores, through which water moves more slowly but can retain water against gravity. The relationship between water content and soil water tension is described by the soil moisture characteristic curve. Two points on this curve, field capacity and permanent wilting point, are particularly important as they describe storage of plant available water. Runoff from forests can be generated by overland flow and lateral subsurface flow. Overland flow usually only occurs on sites compacted by harvesting or which have water-repellent surfaces resulting from fire. Lateral subsurface flow is very common in forests since the lateral hydraulic conductivity and the gravitational gradients (in areas with a steep relief) are often high, and additional preferential flow pathways are present to enhance the downslope flow. The described processes of water storage and movement are applicable in various climates and geographical locations. However, certain processes dominate in certain locations – predicting and understanding water storage and movement in soils require one to use critical thinking to define the first-order controls at a particular site.

See also: **Ecology**: Forest Canopies; Natural Disturbance in Forest Environments. **Health and Protection**: Forest Fires (Prediction, Prevention, Preparedness and Suppression). **Hydrology**: Hydrological Cycle; Impacts of Forest Conversion on Streamflow; Impacts of Forest Management on Streamflow; Impacts of Forest Management on Water Quality. **Soil Biology and Tree Growth**: Soil Organic Matter Forms and Functions; Tree Roots and their Interaction with Soil. **Soil Development and Properties**: Forests and Soil Development; Landscape and Soil Classification for Forest Management; Nutrient Cycling; The Forest Floor.

Further Reading

Beven K (1989) Interflow. In: Morel-Seytoux HJ (ed.) *Unsaturated Flow in Hydrologic Modeling Theory and Practice*, pp. 191–219. Dordrecht, Germany: Kluwer Academic Publishers.

Beven K and Germann P (1982) Macropores and water flow in soils. *Water Resources Research* 18(5): 1311–1325.

Bonell M (1993) Progress in the understanding of runoff generation dynamics in forests. *Journal of Hydrology* 150: 217–275.

DeBano LF (2000) The role of fire and soil heating on water repellency in wildland environments: a review. *Journal of Hydrology* 231: 1–4.

Feddes RA, Kabat P, van Bakel PJT, Bronswijk JJB, and Halbertsma J (1988) Modelling soil water dynamics in the unsaturated zone – state of the art. *Journal of Hydrology* 100: 69–111.

Grayson R and Blöschl G (eds) (2000) *Spatial Patterns in Catchment Hydrology: observations and modeling*. Cambridge, UK: Cambridge University Press.

Hillel D (1998) *Environmental Soil Physics*. San Diego, CA: Academic Press.

Jury WA, Gardner R, and Gardner WH (1991) *Soil Physics*. New York: John Wiley.

Klute A (ed.) (1986) *Methods of Soil Analysis*, Part I, *Physical and Mineralogical Methods*. Madison, WI: American Society of Agronomy.

Kutílek M and Nielsen DR (1994) *Soil Hydrology*. Cremlingen-Destedt, Germany: Catena-Verlag.

McDonnell JJ (1990) A rationale for old water discharge through macropores in a steep, humid catchment. *Water Resources Research* 26(11): 2821–2832.

Pritchett WL and Fisher RF (1987) *Properties and Management of Forest Soils*. New York: John Wiley.

Weiler M and Naef F (2003) An experimental tracer study of the role of macropores in infiltration in grassland soils. *Hydrological Processes* 17(2): 477–493.

Weyman DR (1973) Measurements of the downslope flow of water in a soil. *Journal of Hydrology* 20: 267–288.

Youngs EG (1995) Developments in the physics of infiltration. *Soil Science Society of America Journal* 59: 307–313.