

4. Water relations and hydrologic cycles

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How do ecosystems control the flow and storage of water, and how does water control the functioning of ecosystems? As a result of the International Biological Programme, we are in a better position to answer these questions than ever before. A systematic attempt to quantify water budgets for a variety of vegetations and climates has resulted in a data base for testing both old and new hypotheses. The ecosystem objective of linking the hydrologic system to carbon and mineral cycling has forced a reevaluation of the processes controlling water movement. This perspective extends to understanding how water flows from one terrestrial unit to another and into aquatic systems.

This chapter has two major objectives. The first is to describe the processes or groups of processes affecting water movement and storage, e.g. the details of structure and function. The second objective is to demonstrate the general application of the hydrologic processes by assembling them into a detailed computer simulation model and applying this model to three extremely different kinds of forested watersheds where streamflow data were available.

Throughout this presentation we have eliminated all formal mathematics. We rely heavily on figures to summarize mathematical relations graphically, and on some box and arrow diagrams to illustrate relations between various processes. Pertinent literature is cited for those who want the mathematical details. The second objective requires the use of the master accountant, the computer, to keep track of water within the ecosystem. Computer simulations facilitate interpretation of behavior but also subsume many details.

Fig. 4.1 provides an overview of the structure of a watershed hydrologic system. It summarizes the main components and processes controlling

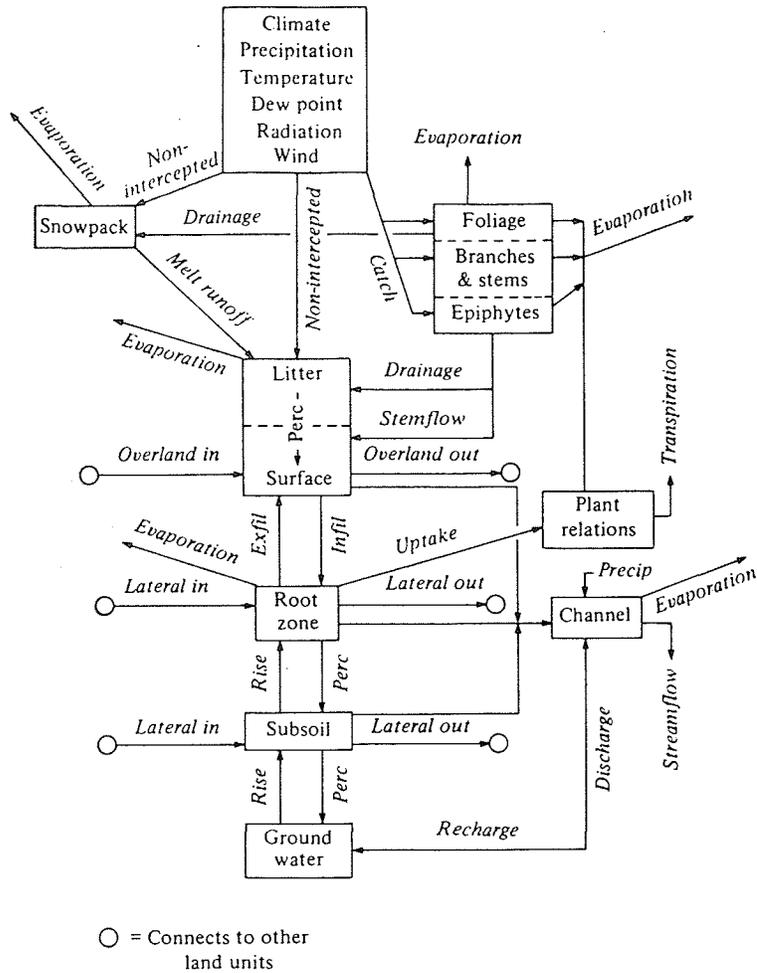


Fig. 4.1. The structure of a watershed hydrologic system showing the main components and processes controlling water movement.

water movement. These processes have a major effect on the structure and functioning of forest and aquatic ecosystems. Each component or process is the subject of individual sections.

The climatic variables indicated in Fig. 4.1 drive the entire system. Of particular importance are humidity and radiation, variables too often ignored but essential to estimate adequately transpiration, evaporation, and the energy content of snow, litter, and soil.

The hydrologist may see refinements in Fig. 4.1 unnecessary for a simple

hydrologic budget. For example, the biological system is subdivided into several compartments above and below ground. For ecosystem purposes an essential distinction is made between the green canopy, important in fixing carbohydrates and the intercepting non-foliar surfaces. The process of evaporation operates from both. The canopy, however, derives water also from the root zone and stem wood of the tree; the amount supplied affects the actual rate of transpiration.

Some compartments, such as litter, hold very little water. Nevertheless, the moisture content of the litter is important because it affects decomposition through its control on the activity of micro-organisms. The snow-pack is hydrologically important in many areas yet completely absent in others. When present it moderates the temperature of the soil, in turn affecting water, nutrient uptake, plant growth, and canopy development.

The below-ground portion of the hydrologic system involves at least two subdivisions of the soil: (a) the root zone which can be subdivided depending upon rooting depth of different plants; and (b) the subsoil below the rooting zone. Processes controlling overland flow and lateral movement through the soil will be discussed in detail.

It should be noted that the size of a compartment which affects the water stored is often in part defined by biological criteria, such as rooting depth, leaf area, or sapwood volume. For example, water stored in the sapwood compartment plays a part in determining stomatal control and growth as will be treated in some detail in the section on plant water relations. These are critical distinctions for ecosystem analyses.

The diagram represents only a single terrestrial ecosystem with simple linkages shown to streams, lakes, and to the ground water. These linkages, characterized by properties of the vegetation and soil, control the amount of water and its routing through the system. Routing forms the basis for examining the interrelation of one terrestrial unit to another.

Finally, to evaluate whether the essential properties of the hydrologic system have been described, we will contrast a variety of watersheds with differing climates, soils and vegetation using simulation models developed independently by the US Deciduous and Coniferous Forest Biomes.

Structure and function of a watershed hydrologic system

The structure we develop is somewhat different from the conventional way of depicting the hydrologic system. The eventual objective of linking to erosion and carbon-mineral cycling has caused us to view the hydrologic system from an ecosystem perspective. Thus, we emphasize the importance of interactions between biological components and the hydrologic system, and spatial variation of these processes within a watershed.

Dynamic properties of forest ecosystems

Stratification into terrestrial and aquatic units

Spatial variation of hydrologic processes within a watershed may be caused by differences in soils, vegetation, geology, physiography and climate. Differences may exist in both terrestrial and aquatic portions of the watershed.

Where significant spatial variation does exist we can stratify a watershed into a set of homogeneous terrestrial units and a set of homogeneous aquatic (or channel) units. The following considerations are important and should be considered in addition to those commonly used in stratification of watersheds such as soil type, slope, and aspect.

Environmental stimuli of soil moisture, soil and air temperature, light, soil fertility, and mechanical stress have been shown to be significant factors in the distribution and growth of forest flora and vegetation (Cleary & Waring, 1969; Waring, 1969; Waring & Youngberg, 1972; Emmingham & Waring, 1973). These stimuli produce quantifiable plant responses including phenology, carbon dioxide exchange, plant moisture stress, stomatal resistance, and foliar nutrition (Waring, Reed & Emmingham, 1972; Reed & Waring, 1974). Stimuli and responses then form ecological indices which locate ecosystems in an environmental grid and permit prediction of such characteristics as productivity and species composition (Waring *et al.* 1972). An example of this grid is Fig. 4.2, which shows the distribution of selected conifers in the Siskiyou Mountains in southern Oregon in relation to a temperature-growth index and plant moisture stress (Waring *et al.* 1972). Such an ecological assessment of environment served to stratify experimental watersheds in the Coniferous Forest Biome.

Given stratification of a watershed into terrestrial and aquatic units on an ecological basis, we are in a position to study the behavior of the watershed from an ecosystem viewpoint. That is, we will be able to interpret watershed behavior from the behavior of the smaller interacting units. We can in turn interpret the behavior of each unit in terms of the vegetation physical characteristics and inputs to that unit.

Climate variation over a watershed

Before discussing the hydrologic processes on each unit we should first discuss the climatic environment and how this varies over a watershed. We are generally interested in: (1) amount and kind of precipitation, (2) air temperature, (3) vapor pressure, (4) short and long wave radiation, and (5) wind movement. We will limit our discussion to some methods of estimating important variables, and problems associated with extrapolating point measurements to other areas. We will not discuss the microclimate within a stand, and refer the reader to Geiger (1965) and to Chapter 3 of this volume for information.

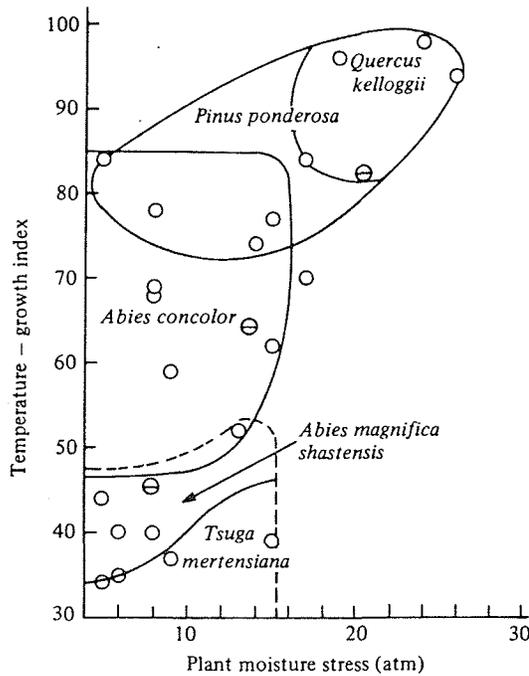


Fig. 4.2. Distribution of natural regeneration in relation to gradients of moisture and temperature defined by plant response indices (Waring *et al.*, 1972). Indices relate to response by 1-2 m tall Douglas fir. The temperature index sums the fractions of potential growth for each day during the growing season, assuming maximum potential with 25°C air and 20°C soil temperature. The moisture stress index represents predawn values measured in September, the peak of the summer drought.

Amount and kind of precipitation

The processes controlling precipitation are understood and the reader is referred to Eagleson (1970) for a careful review which includes cloud physics. It is important to have an appreciation of the phenomena and to be aware of the spatial and temporal variability of precipitation which may occur over a watershed.

The occurrence, total duration and total amount of storm precipitation seems to be largely stochastic and treatable by probabilistic methods (Eagleson, 1970). Stochastic models for point precipitation from summertime convective storms have been developed by Grace & Eagleson (1966), Sariahmed & Kisiel (1968), Duckstein, Fogel & Kisiel (1972), Duckstein, Fogel & Thames (1973), and Smith (1974). It may be possible to treat synoptic scale storms in a similar manner.

The internal structure of a given storm may be largely deterministic but is dependent on the type of storm (Fogel & Duckstein, 1969; Eagleson, 1970).

Dynamic properties of forest ecosystems

Smith, 1974). Eagleson (1970) states that in general the difference between mean rainfall over an area about the storm center and what is measured at the storm center: (1) increases with decrease in the total rainfall, (2) decreases with increasing duration, (3) is greater for convective and orographic precipitation than for cyclonic, and (4) increases with increasing storm area.

Topography can have significant effects on precipitation. Precipitation usually increases with elevation because of orographic cooling and a precipitation–elevation relation is frequently needed when using point rainfall data. Eagleson (1970) indicates this effect may only extend to about 1500 above the general terrain, but local topography can have significant influence (Burns, 1953; US Army Corps of Engineers, 1956; Linsley, 1958). A World Meteorological Organization (1972) symposium was dedicated to discussion of precipitation in mountain areas.

The above considerations enter into the design of rainfall monitoring networks (Rodriguez-Iturbe & Mejia, 1974*a,b*). Methods of extending point data to areas are discussed by Rodriguez-Iturbe & Mejia (1974*b*), Rodda (1970), Mandeville & Rodda (1970), Hutchinson & Walley (1972), Shaw & Lynn (1972), and Wei & McGuinness (1973).

Separation into type of precipitation, snow and rain, is done on the basis of observation or by temperature. Eagleson (1970) suggested using screen air temperature to separate snow from rain with a dividing line of 1.1–1.7°C. Anderson (1968) suggested using wet bulb temperature, which can be estimated from air and dew point temperature, with a dividing line of 0.5°C

Temperature and vapor pressure

Ambient temperature typically decreases with increasing elevation at a rate between 5 and 8°C km⁻¹. This gradient is referred to as the ambient lapse rate of temperature (Eagleson, 1970), and can be safely used only if the mountain localities being compared have approximately similar environments and differences between them in character and profile of the surface are small. Local topography and climatic conditions may influence temperatures within a watershed. For example, the presence of night-time cold air drainage or formation of inversion layers may result in a positive ambient lapse rate near the ground. This could cause warmer temperature at higher elevations than at lower elevations and the formation of thermal belts. Large bodies of water may also influence local temperatures. An example of how the lapse rate can vary is shown in Fig. 4.3. The difference shown are due to transition from plains to the foothills and to influences of glaciers in the upper part of the valley. Variations within the day are due to different rates of cooling of various parts of the valley. The open areas of the plains and foothills cool more rapidly than the higher elevations (from Kuzmin, 1961).

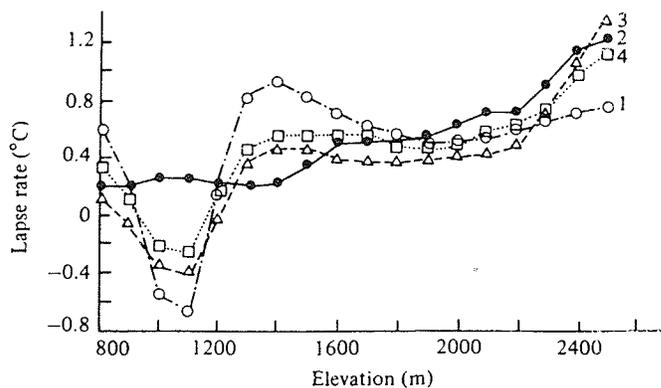


Fig. 4.3. Changes in mean annual lapse rate at (1) 7 a.m., (2) 1 p.m., (3) 9 p.m., and in (4) the mean daily air temperatures with elevation along the Zeraushan River Valley, USSR, for 1933. From Kuzmin, 1961.

The water holding capacity of the air is generally dependent on the temperature and pressure. We can expect vapor pressure to vary in ways similar to temperature. Since this is strongly influenced by elevation, we can expect vapor pressure to decrease with altitude. Kuzmin (1961) reports several relations of which one is shown in Fig. 4.4.

Radiation

The total radiation incident on any surface is the sum of (1) direct short wave radiation from the sun, (2) diffuse short wave radiation from the sky, (3) reflected short wave from nearby surfaces, (4) long wave radiation from atmospheric emission, and (5) long wave emitted from nearby surfaces.

For the northern hemisphere, Houghton (1954) reports mean values for direct short wave of 0.12 ly min^{-1} , diffuse short wave of 0.11 ly min^{-1} and atmospheric long wave of 0.52 ly min^{-1} . For ecological purposes the short wave component is generally the most important. The long wave component becomes important in some water balance studies such as snow melt.

The short wave radiation on a plane surface at any instant is dependent on the slope, aspect, and latitude of the surface, the declination of the sun, time of day, and transmissivity of the atmosphere. Given this information, the direct short wave radiation on any slope may be estimated for any time interval using theoretical methods (Garnier & Ohmura, 1968, 70; Buffo, Fritschen & Murphy, 1972); Short wave radiation can vary significantly with topography (Fig. 4.5).

A problem arises in determining the atmospheric transmissivity. Garnier & Ohmura (1970) reported a method for estimating atmospheric transmissivity.

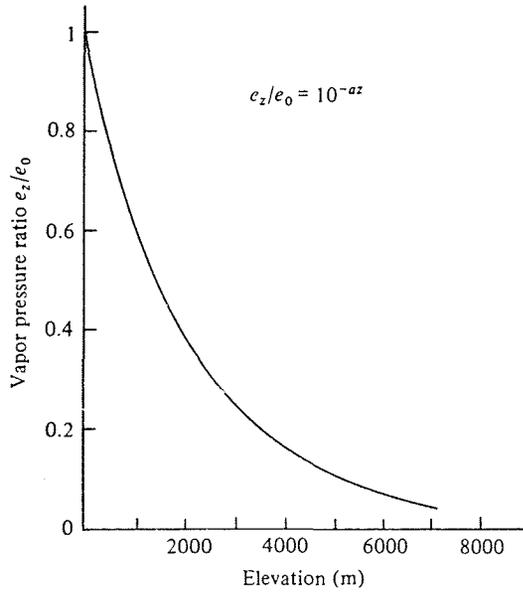


Fig. 4.4. Ratio of vapor pressure, e_z , at elevation, z , to that at sea level, e_0 , for $a=0.2$.

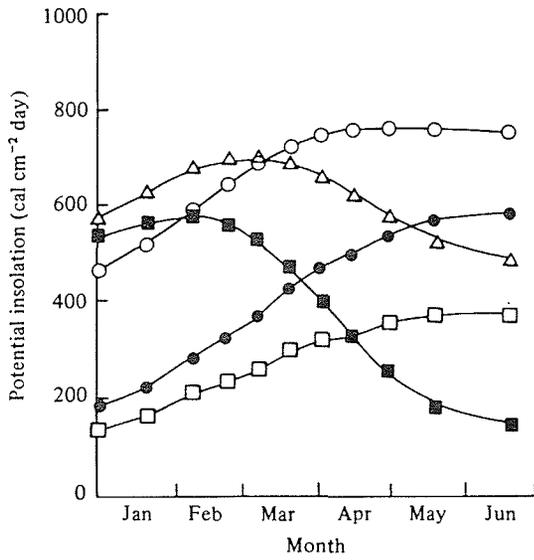


Fig. 4.5. Isograms of daily values of direct solar radiation at latitude 40°N and slopes of 60°E (\bullet — \bullet), 90°E (\square — \square), 30°S (\circ — \circ), 60°S (\triangle — \triangle), and 90°S (\blacksquare — \blacksquare) for atmospheric transmissivity of 0.90. From Buffo *et al.*, 1972.

sivity from observed global and diffuse short wave radiation on a horizontal surface. They also report a method for estimating diffuse short wave radiation on slopes. On clear days, transmissivity generally increases with altitude because the optical path length through the atmosphere is reduced while the moisture content and dust pollution are decreased. This increases clear sky short wave radiation at a rate of about 0.012 ly/min per 100 m (Kuzmin, 1961). Since cloudiness usually increases with altitude in mountain areas, the gain under clear sky conditions may be more than offset and total short wave radiation on horizontal surfaces may remain constant or decrease with altitude (Kuzmin, 1961).

Evaluation of long wave radiation income is more difficult. Various methods proposed have been summarized by Sellers (1965) and Eagleson (1970). Some methods are applicable only to clear sky conditions while others consider effects of clouds. Anderson & Baker (1967) proposed a method for use under all atmospheric conditions. Long wave radiation from atmospheric emission is generally uniformly distributed over a watershed. It is not significantly affected by topography except in extremely steep terrain, when back radiation from adjacent slopes may become important.

Wind

Wind speed increases with altitude only in the free atmosphere. The determination of wind speed and direction in mountain areas requires data from direct observation since it is influenced by local conditions of topography, exposure, heating and other factors. These effects are extensively discussed by Geiger (1965).

Hydrology of terrestrial strata

Here we will discuss the hydrology of individual terrestrial units from an ecosystem viewpoint. In the next sections, we will link these units to each other and to aquatic units.

In words, the water content of a terrestrial unit is the sum of the water stored (1) on the foliage, branches and stems, (2) in the snowpack, (3) in the litter, (4) on the soil surface, (5) in the vegetation, (6) in the soil root zone, and (7) in the subsoil. The structure and relations were shown in Fig. 4.1. We will discuss each part of the water balance in the following sections.

Interception by foliage, branches and stems

In vegetated regions, plant surfaces are the first obstacles encountered by precipitation. As a storm begins, rain strikes the foliage, branches, and stems or falls directly through canopy openings to the forest floor. The

latter route is minor in fully closed forest stands. During the initial stages of a storm, much of the precipitation is stored on the canopy or upon the stems. As a storm continues and these surfaces reach their capacities, excess water drains to the forest floor. Evaporation of intercepted precipitation may take place throughout a storm but is of primary importance after precipitation ceases. Thus, interception is the combination of processes determining water storage on tree foliage, branches, and stems. The change in water storage on foliage, branches and stems is the sum of (1) intercepted precipitation, minus (2) drainage from canopy and stem drip, (3) stem flow to the ground, and (4) evaporation from vegetation surfaces. Intercepted precipitation may be rain, snow, rime or dew.

Interception has been studied for nearly a century (Hoppe, 1896) and has been shown to account for losses ranging from 10 to 35% of annual precipitation (Kittredge, 1948; Zinke, 1967). The principle process causing interception loss is evaporation from water stored on plant surfaces. Leonard (1967) discussed and depicted the theoretical relationships between interception loss processes during a storm (Fig. 4.6), and pointed out the need for improvement in measurement techniques for separating processes. Other descriptions of the rainfall interception process have been provided by Horton (1919), Grah & Wilson (1944), Rowe & Hendrix (1951), and Rutter, Kershaw, Robins & Morton (1971). The process of snow interception by trees is described by Hoover & Leaf (1967) and Miller (1967). We do not consider the interception processes for other forms of water deposited on plant surfaces although they may be of local importance. For example, rime has been recognized as a contribution to the hydrological balance of conifers (Berndt & Fowler, 1969) and hardwoods (Gary, 1972); the importance of dew in the water balance of conifers has also been demonstrated for some localities (Kittredge, 1948; Fritschen & Doraiswamy, 1973).

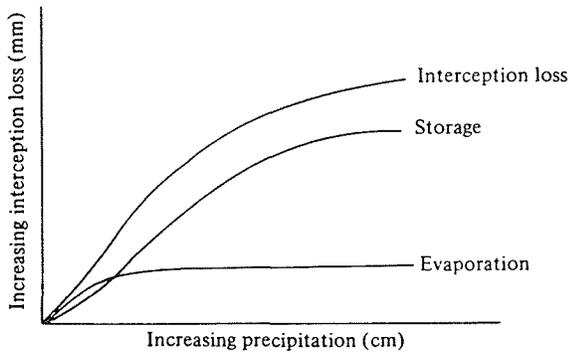


Fig. 4.6. Theoretical relationship between interception loss and precipitation by components during the initial stages of a storm. After Leonard, 1967.

In the following sections, we will address the problems of understanding and estimating interception components.

Storage and evaporation processes. The surface water storage capacity of a forest is related to the surface area represented by foliage, branches, and stems. The configuration, orientation, and texture of the surfaces also influence storage capacities. Since the surface water holding characteristics of each tree component vary, it is conceptually desirable to treat each component separately.

The storage capacity of forest vegetation has been estimated by extrapolating regressions of interception loss on precipitation back to the intercept on the y-axis and also by experimental techniques (Leyton, Reynolds & Thompson, 1967). For a plantation of Scots pine (*Pinus sylvestris* L.), Rutter (1963) found that leafy shoots retained an amount of water approximately equal to the dry weight of foliage. Storage values of foliage, branches, and stems were estimated to be 0.8, 0.3, and 0.25 mm, respectively, Voigt & Zwolinski (1964) estimated that storage capacities for red pine (*Pinus resinosa* Ait.) and white pine (*Pinus strobus* L.) were 0.8 and 0.5 mm, respectively, with about 40% of the water retained on the stems. Swank (1972) estimated 1.8 mm storage capacity for a Douglas fir (*Pseudotsuga menziesii* (Mirb.) Franco) stand with about 36% retained by the stem. In a predictive model of rainfall interception for Douglas fir, Rutter, Morton & Robins (1975) used 2.1 mm storage capacity for the stand with 43% contributed by the stems. Canopy storage for Scots pine, Norway spruce (*Picea abies* (L.) Karst.), grand fir (*Abies grandis* (Dougl.) Lindl.) and a deciduous stand of *Fagus* and *Carpinus* was estimated as 3.0, 3.1, 3.8, and 1.9 mm, respectively (Aussenac, 1968) and a value of 1.5 mm was estimated for Norway spruce (Leyton *et al.* 1967). Zinke (1967) summarized interception storage values for conifers and hardwoods; when taken collectively with other values in the literature, average rainfall storage capacities for conifers are about 1.9 mm and for hardwoods about 1 mm. Snow storage for conifers averages about 3.8 mm (Zinke, 1967).

Large deviations from these average values occur by season and stage of stand development. The distribution of storage by tree components also changes with stand development. Surface area of branches and stems continues to increase with stand age and comprises a substantial proportion of the total intercepting surface in mature stands (Fig. 4.7). Values of foliar surface (both sides) per unit of ground area for closed deciduous forests of the eastern United States range between 6 and 12, branch area 1.2–2.2, and stem area 0.3–0.6 (Whittaker & Woodwell, 1967). Values reported for a young eastern white pine plantation were 17.8, 2.3, and 0.4 for foliage (all surfaces), branch, and stems areas, respectively (Swank & Schreuder, 1973).

Evaporation is a process operating from all surfaces which hold water and are exposed to the atmosphere. Evaporation from the litter and soil

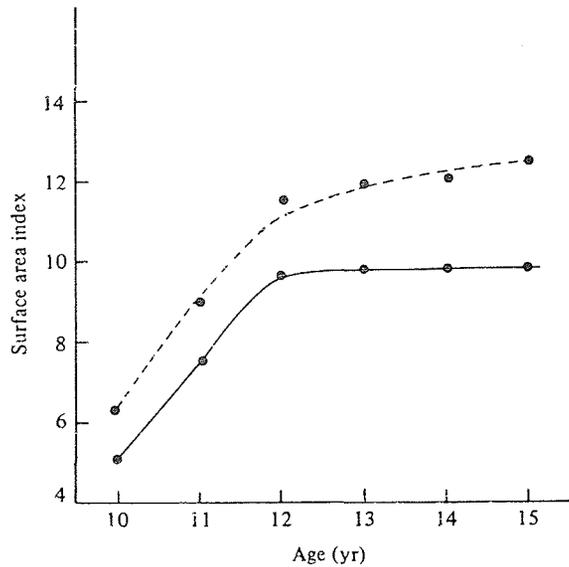


Fig. 4.7. Time trends of leaf area index (—) and total above ground surface area index (---) for a plantation of eastern white pine (*Pinus strobus*) from age 10 to 15 yr. From Swank Schreuder, 1973.

surfaces are discussed in a separate section. Evaporation of intercept water is an energy-dependent process and is predicted from measurement of net radiation, temperature, humidity, and wind speed (Penman, 1956; Monteith, 1965).

Currently, interception models are based on a combination of physical processes and empirical relationships. A model of the evaporation of intercepted rainfall (Rutter *et al.*, 1971) was modified and applied to both deciduous broad-leaved and evergreen coniferous stands (Rutter *et al.*, 1975). Good agreement was found when interception loss values calculated by model simulation were compared with observed values. Evaporation of water retained on tree stems was estimated to be 20–30% of the total loss from trees in leaf, and 30–40% for leafless trees (Rutter *et al.*, 1975). A simulation model of evaporation of intercepted rainfall has also been used to show that precipitation intercepted by vegetation evaporates at a greater rate than transpiration from the same type of vegetation in the same environment (Murphy & Knoerr, 1975).

Empirical relationships for drainage and stemflow. Summaries of the results from many traditional interception studies were made by Helvey & Patton (1965) for mature hardwood forests of the eastern United States and

Table 4.1. Summary of equations for computing throughfall and stemflow for coniferous^a and hardwood^b forests from measurements of gross rainfall

Species	Average equations (gross rainfall in cm)		Interception loss (cm)
	Throughfall	Stemflow	
Red pine	$0.87p - 0.04$	$0.02p$	0.15
Loblolly pine	$0.80p - 0.01$	$0.08p - 0.02$	0.15
Shortleaf pine	$0.88p - 0.05$	$0.03p$	0.14
Ponderosa pine	$0.89p - 0.05$	$0.04p - 0.01$	0.13
Eastern white pine	$0.85p - 0.04$	$0.06p - 0.01$	0.14
Average (pines)	$0.86p - 0.04$	$0.05p - 0.01$	0.14
Spruce-Fir-Hemlock	$0.77p - 0.05$	0.02	0.26
Mature mixed hardwoods			
Growing season	$0.90p - 0.03$	$0.041p - 0.005$	0.10
Dormant season	$0.91p - 0.015$	$0.062p - 0.005$	0.05

^a Conifer equations from Helvey (1971).

^b Hardwood equations from Helvey & Patric (1965).

Helvey (1971) for conifers in North America. More than 50 different forest stands are included in these summaries and the stands represent closed conditions rather than thinned, pruned, or other manipulated conditions. In these standard interception studies, interception loss is derived from the difference in measured values of gross precipitation and throughfall plus stemflow. The general form of prediction for throughfall, stemflow, and interception loss for individual storms is a linear regression with gross precipitation as the independent variable. A composite of the summaries is given in Table 4.1 with estimated interception loss assuming 1 cm of precipitation. Annual or periodic throughfall and stemflow can be computed by solving the equations for annual (or periodic) gross rainfall and multiplying the constant term by the number of storms in which gross rainfall equals or exceeds the constant term. Total rainfall delivered in storms which are smaller than the constant terms in the equations, is added to the difference between precipitation - (throughfall + stemflow) to obtain interception loss. These equations represent average values from a variety of forest stands and a range of constants exist. Interception loss is greatest in the spruce-fir-hemlock type, intermediate in pine, and least in broad-leaved deciduous forests. Because surface area index also is greatest in the spruce-fir-hemlock type (Burger, 1925), intermediate in pine species (Swank & Schreuder, 1974), and least in deciduous forests (Whittaker & Woodwell,

1967), the surface area of forests provides an important denominator in modeling interception processes. Other extensive summaries of interception studies are provided by Kittredge (1948), Molchanov (1960), and Zinke (1967).

There have been few reports of the quantitative relationship between interception processes and specific vegetation characteristics. Attempts to relate interception loss quantitatively to stand basal area have been inconclusive (Rogerson, 1967; Kittredge, 1968), but some results suggest a relationship with canopy density (Rothacher, 1963; Clegg, 1963). Grah & Wilson (1944) showed a close correlation between interception loss and foliage weight for Monterey pine, and Swank (1972) described models for estimating interception components from live crown weight for Douglas fir. Several studies have demonstrated a strong relationship between stemflow and coniferous tree measurements (Wicht, 1941; Leonard, 1961; Rutter, 1963; Swank, 1972).

From this examination of the interception process, it is clear that periodic interception loss of rainfall can be regarded primarily as a function of periodic precipitation, evaporation during the storm, and the number of times storage capacity is filled within a specific period. Equally clear is the fact that a wide range of storage capacities must exist for forest communities since the quantity and proportion of evaporating surfaces are highly variable.

Snowpack

Snowpack dynamics are complex but fairly well understood. The major reference volume is the report of the Cooperative Snow Investigations Program by the US Army Corps of Engineers (1956). Eagleson (1970) summarized this and more recent work and Kuzmin (1961) summarized Russian work.

An understanding of snowpack dynamics requires familiarity with the physical nature of the snowpack as well as with the energy and water balance processes. We will briefly review each of these.

Snowpack structure. A new snowpack has low density because of snowflake structure. However, it undergoes metamorphosis with time resulting from (1) exchange of heat at snow surface by radiation convection and condensation and at ground surface by conduction, (2) compaction under its own weight, (3) rain or melt water percolation through the snowpack, (4) wind, and (5) variations within the snowpack of temperature and water vapor (US Army Corps of Engineers, 1956; Eagleson, 1970).

The density of new snow varies from about 0.06 to 0.34 g cm^{-3} (US Army Corps of Engineers, 1956; Eagleson, 1970) increasing with both wind and temperature. Variation of density with temperature is shown in Fig. 4.8.

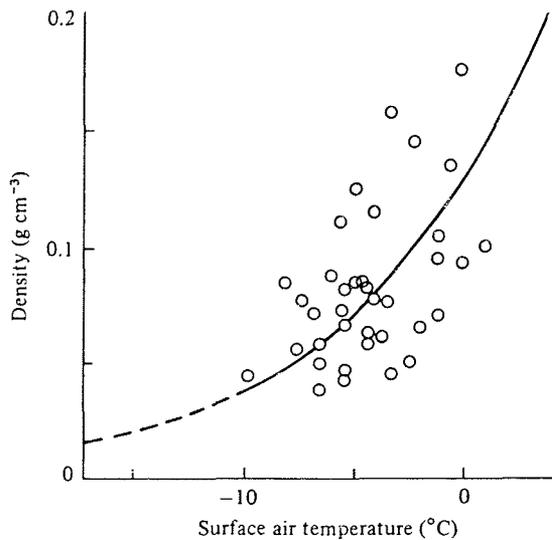


Fig. 4.8. The density of new fallen snow as related to air temperature. From US Army Corps of Engineers, 1956.

Average density of the snowpack generally increases with time, decreasing after each new snowfall. It generally increases with depth with local variations due to buried snow crusts, ice lenses and planes. Snow crystals gradually assume a granular form due to percolating water and convection of heat and water along temperature gradients in the pack.

Physical characteristics of the snowpack which are generally related to density include thermal conductivity, liquid water holding capacity, and thermal diffusivity. Relations for these are reported by the US Army Corps of Engineers (1956), Eagleson (1970), Anderson (1968), and Eggleston, Israelson & Riley (1971). The relation of liquid water holding capacity to density is shown in Fig. 4.9.

As metamorphosis proceeds, the snowpack tends to become homogeneous with respect to density, temperature, liquid water content and grain size. When runoff from the snowpack occurs, the greater part of the snow has been brought to the melting point and the latent heat of fusion has been added. When the pack is at the melting point it is called isothermal. The pack is said to be ripe, or ready to transmit and discharge any water which enters at the surface, when it is isothermal and contains all the water it can hold against gravity. Only the upper surface changes when this state is reached. These changes are day-time thaw which produces melt water and night-time freezing which forms a surface crust.

Another physical characteristic affecting snowpack energy balance and melt rate is the albedo or reflectivity of the snow surface. Albedo is generally

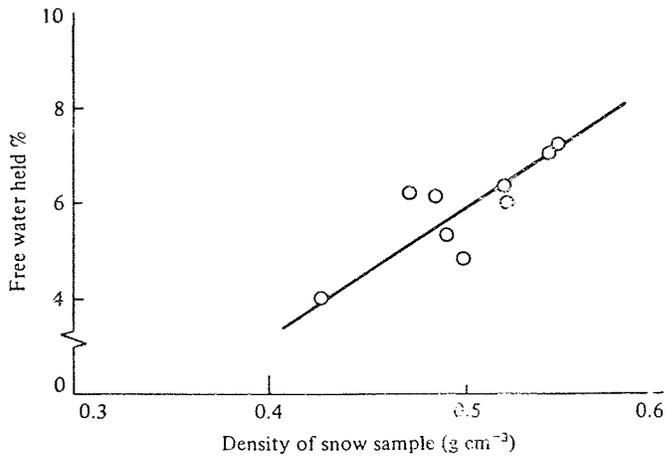


Fig. 4.9. Water holding capacity of ripe snow as related to snow density. From US Army Corps of Engineers, 1956.

related to the state of metamorphosis of the snow surface. This is reflected in Fig. 4.10 in which albedo is shown to decrease with accumulated temperature after snowfall.

Energy and water balance of the snowpack. In studying heat and water balance processes of snowpacks it is most useful to distinguish between the

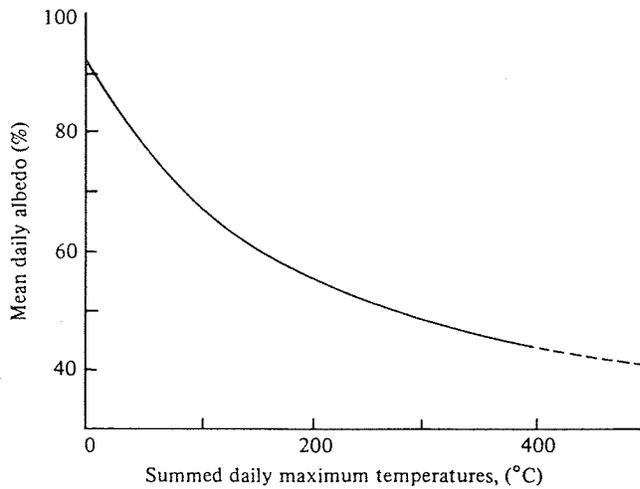


Fig. 4.10. Variation of albedo of snowpack with accumulated temperature index. From US Army Corps of Engineers, 1956.

ice portion and the liquid water portion of the pack. We generally assume that if liquid water exists, it is at 0°C.

The change in water equivalent of the ice portion of the snowpack is the sum of (1) snowfall, and (2) condensation from vapor to solid (sublimation, if negative) minus (3) melt. The change in liquid water content is the sum of (1) rainfall, (2) melt, and (3) condensation from vapor to liquid (evaporation, if negative), minus (4) runoff from the pack. If the melt is negative it represents the freezing of liquid water within the pack.

A number of heat-transfer processes are involved in snowpack dynamics and melting of snow. The importance of each depends on geography, season and climate. The energy balance assumes that the change in heat storage of the snowpack is the sum of (1) net radiation heat transfer, (2) release of latent heat of vaporization by condensation or, if negative, removal by sublimation or evaporation, (3) convective transfer of sensible heat from air, (4) gain of latent heat by freezing of liquid water within pack (or melt if negative), (5) conduction of heat from underlying ground, (6) advection of heat by a gain (precipitation) or loss (runoff) of water (Anderson, 1968). Fig. 4.11 summarizes the energy and water balances of a snowpack. We will briefly discuss each component of the energy balance.

With the exceptions of freezing of liquid water within the pack, and conduction of heat from the ground all components of the energy balance can be assumed to take place in a thin surface layer of snow (Anderson, 1968). This is important in determining the surface temperature of the snowpack since the net radiation and transfer of sensible and latent heat are dependent on surface temperature.

The transfer of latent heat to the surface is a turbulent exchange process. The direction of transfer is controlled by the vapor pressure gradient. If the air vapor pressure is greater than the surface vapor pressure, moisture is transferred to the surface. This results in condensation and release of latent heat to the surface. If the gradient is reversed, the direction of moisture and heat transfer is reversed. If the snow surface is at 0°C it is generally assumed the phase change is from vapor to liquid or vice versa and uses the latent heat of vaporization (597.3 cal g^{-1}). If it is below 0°C the phase change is from vapor to solid and uses the latent heat of sublimation (677 cal g^{-1}). This process can be important in areas where warm, moist, turbulent air occurs over snowpacks, since 1 cm of condensate can produce about 7.5 cm of melt, or 8.5 cm of runoff (Anderson, 1968; Eagleson, 1970).

The transfer of sensible heat to the surface is also a turbulent exchange process. The direction of transfer is controlled by the temperature gradient. If the air temperature is warmer than the snow surface there is a direct heat transfer from the air to the snow and vice versa. The quantities of melt by convection from warm turbulent air can be similar to those from condensation melt.

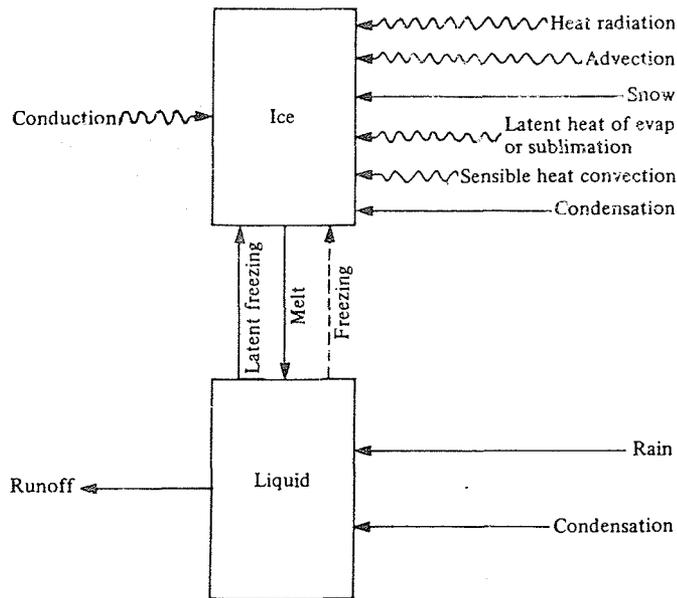


Fig. 4.11. The energy and water balance components of snow showing major processes affecting them. Solid lines denote water transfer and wavy lines energy transfer.

Rain is warmer than the snowpack and hence there is a transfer, by advection, of heat from rain to the snowpack. Since the specific heat of water is 1 cal g^{-1} , the amount of heat added is easily calculated from the amount (cm) and temperature ($^{\circ}\text{C}$) of the rain. The temperature of rain can be assumed to be air temperature. Since the latent heat of fusion is 80 cal g^{-1} one can see that the melt-producing capacity of rain is small. For example, 4 cm of rain at 20°C must be added to a ripe snowpack to produce 1 cm of melt. However, when the snowpack is below freezing the rain would add its small heat content and also give up the latent heat of fusion (80 cal g^{-1}) when it froze within the pack. This is a quite significant source of heat and will quickly bring a cold snowpack to an isothermal or ripe condition.

Net radiation heat transfer includes both short wave and long wave radiation balances. In forested areas this is complicated by the presence of forest cover. The solar radiation incident above the canopy is attenuated by the foliage. This must be considered in determining what is incident at the snow surface (see p. 231). The albedo of the snow surface then determines the portion reflected, and the balance is absorbed.

Part of the long wave radiation from the atmosphere, discussed in an earlier section, is absorbed by the forest cover. However, the forest cover

also emits long wave radiation, and is usually assumed to behave as a black body radiating at air temperature. Hence, the portion of the snowpack covered by forest is assumed to receive radiation from the canopy, while that in the open receives it from the atmosphere. The snowpack is usually assumed to absorb all incident long wave radiation. It in turn emits long wave radiation as a black body at the temperature of the snow surface.

Conduction of heat from the ground generally is the result of release of heat stored in the soil during the summer months. This may produce melt from the bottom of the snowpack. The process is controlled by the temperature gradient from the soil-snow interface. The amounts are generally small and usually neglected.

Melt which occurs in the surface of a cold snowpack percolates down where it refreezes and releases the latent heat of fusion. This warms the pack until it eventually becomes isothermal. Additional melt then is held as liquid water until the water holding capacity of the pack is reached and runoff from the bottom of the pack occurs.

Simulation models of the snowmelt process have been reported by Anderson (1968), Eggleston *et al.* (1971), and Leaf & Brink (1973). The model by Leaf & Brink (1973) is generally restricted to sub-alpine conditions. The latter two have broader applicability, include more processes and require more climatic data than Leaf & Brink's model.

Litter and soil surface

Water balance processes in the litter layer are important in litter decomposition and nutrient movement. Hence it is necessary that this be dealt with in hydrologic models of an ecosystem.

The change in water content of the litter and soil surface is the sum of (1) snowpack runoff, (2) drip from the foliage, stems and branches, (3) direct precipitation throughfall, (4) lateral flow into the layer, and (5) exfiltration from the soil up into the litter; minus (6) evaporation, (7) lateral flow out of the layer, and (8) percolation from the bottom of the litter layer into the soil as infiltration. For water balance purposes we assume that any stemflow moves directly into the soil. Any roots in the litter are included in the root zone.

Infiltration is the process of entry into the soil of water made available at its surface (Philip, 1969). Percolation is the process by which water moves downward under gravitational forces through the litter, soil, or other porous media. Lateral flow is the movement of water under gravitational forces generally parallel to the slope. Although infiltration is primarily a soil related phenomenon, we discuss it here because of its importance in determining what happens in the litter layer.

There have been several studies of litter water balance. Many of these are reviewed by Helvey & Patric (1965), Zinke (1967), and Helvey (1971). A detailed study is reported by Plamondon, Black & Goodell (1972), but such detailed studies are rare literature.

During a storm, water falling on the litter will gradually increase the litter moisture level to field capacity. Additional water will then percolate through to the bottom of the litter and infiltrate into the soil unless the capacity of the soil to absorb water is exceeded. Then additional rainfall could raise the litter moisture to saturation. Lateral flow can occur through the litter under the force of gravity after field capacity is reached. This may appear as overland flow in areas where litter is sparse.

The forest floor may be divided into several layers, based on stages of decomposition. There are several methods of nomenclature for this. For purposes of discussion we will divide the forest floor into two main layers: (1) an upper horizon in which the origin of most material is identifiable to the naked eye, and (2) a lower horizon in which the origin of material is not recognizable to the naked eye. The hydrology of the two layers is different.

The upper layer is generally composed of leaves, branches, and logs. These have the capacity to hold water on their surface and absorb a certain amount of water. The lower layer on the other hand behaves more like the soil.

Since the moisture content of fine fuels (<8 cm diameter) in the upper layer is significant in determining flammability, these have been studied extensively in developing fire danger ratings (Schroeder & Buck, 1970; Deeming *et al.*, 1974). In general, litter tends to approach a moisture content in equilibrium with its surroundings. The rate at which it comes to equilibrium depends on temperature, humidity, structure of the material, surface weathering, density and so on (Deeming *et al.*, 1974).

The water-holding capacity of the surface horizon essentially depends on the surface area of the material, analogous to storage on foliage, stems and branches. Data reported by Clary & Ffolliott (1969) indicate that upper horizons of Ponderosa pine litter can hold about 175% of dry weight at field capacity. In the lower layers water is held by capillary force and the capacities increase. Clary & Ffolliott (1969) report moisture contents of 210% for the lower layer of Ponderosa pine litter at field capacity.

For the entire litter layer Helvey (1971) reports an average capacity for conifer litter of 215% at field capacity and about 330% at saturation. For hardwood litter Helvey & Patric (1965) report about 160% at field capacity and 220% at saturation. The minimum moisture content for both hardwood and conifer litter was about 30% (Schroeder & Buck, 1970).

Evaporation from litter storage is another area in which literature is limited. Most work in the area has been done in development of the fire danger rating technology (Deeming *et al.*, 1974).

Moore & Swank (1975) developed a process model for litter water content and evaporation. They used data reported by Helvey (1964) to test the model and results are shown in Fig. 4.12.

Plamondon *et al.* (1972) found that excess water percolates very rapidly through litter. However, movement into litter and percolation may be restricted if the litter is very dry initially and hydrophobic conditions exist (Meeuwig, 1971; Deban & Rice, 1973) or if matting occurs (Pierce, 1967). We will neglect hydrophobic problems here. Hence the rate at which percolating water can infiltrate into the soil determines whether an excess is available for lateral flow through the litter (Pierce, 1967).

In examining infiltration under natural conditions it is helpful to divide soil porosity into two parts – a channel system and a capillary system (R. M. Dixon, 1971; Dixon & Peterson, 1971). A channel system exists in both the surface and subsurface and includes a network of large soil pores which drain and fill by gravity. The channel system also includes micro-topographical characteristics of the soil surface and pores created by clay shrinkage, cracking, roots, earthworms and other soil organisms. The

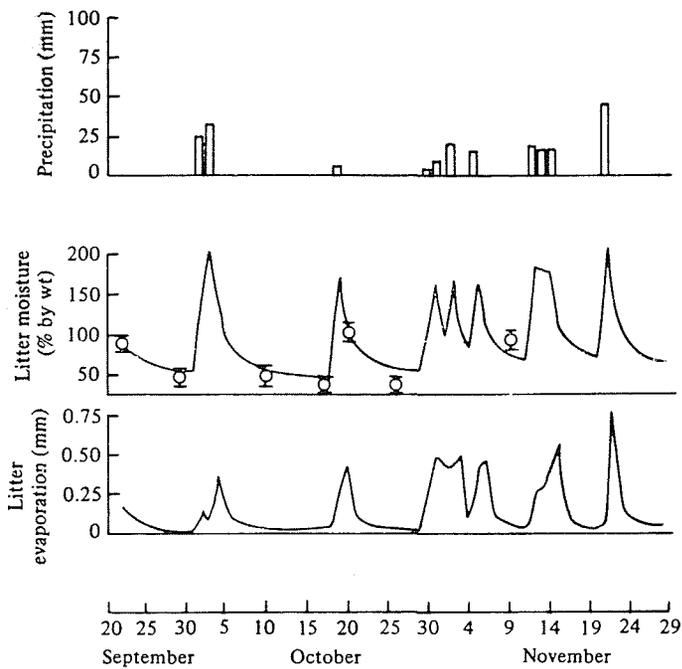


Fig. 4.12. Measured precipitation and simulated litter moisture content and litter evaporation for an 80-day period in 1961. Experimental estimates of water content (from Helvey, 1964) with 67% confidence intervals shown. From Moore & Swank, 1975.

capillary system includes textural and structural pores within the soil mass which drain and fill mainly as a result of capillary forces. The channel system is embedded in the capillary system and shares a common boundary at the soil surface and along the walls of the large pores. R. M. Dixon (1971) reports that a good channel system can increase intake rates by an order of magnitude or more over that observed on similar soils lacking such. This is substantiated by other studies where root development and vegetative cover combined to produce high infiltration rates (Rauzi, Fly & Dyksterhuis, 1968; Verma & Toogood, 1969; Meeuwig, 1970; Rauzi & Smith, 1973). Vegetation removal or soil compaction leads to significantly reduced infiltration capacities (Dyrness, 1972).

There are two general approaches to the treatment of infiltration. The first and most common is to neglect the channel system and consider soil as an isotropic, homogeneous porous medium. The second attempts to consider the channel system. The first approach led to the development of a mathematical theory for soil water movement and infiltration through homogeneous soils or soils of two or more homogeneous, horizontal layers (Bear, Zaslavsky & Irmay, 1968; Philip, 1969; Childs, 1969). The infiltration equations developed from this theory can be used to predict infiltration rates in laboratory tests where the channel system has been destroyed, but are rarely accurate under natural field conditions. The second approach is, at present, beyond deterministic mathematical analysis. Hence one must rely on empirically or physically based conceptual models for estimating infiltration, such as those developed by Green & Ampt (1911), Gardner & Widstoe (1921), Kostiaikov (1932), Horton (1933), Philip (1954, 69), and Holtan (1961, 1969, 1971). These have been reviewed by Childs (1969), and Whisler & Bouwer (1970).

The infiltration process can also be affected by water temperature (Klock, 1972), frozen soils (Haupt, 1967; Sartz, 1973; Dingman, 1975), and shrinkage cracks (Blake, Schlichting & Zimmerman, 1973).

When the infiltration capacity of the soil is exceeded, the excess water in the litter may begin to move laterally downslope under the influence of gravity. Conditions which may cause this to happen in forests are reviewed by Pierce (1967).

Some water may be held in small depressions as detention or static storage. The excess water moves overland as sheet flow or in small rivulets. This eventually converges into the lowest order channels formed by microtopography provided that the flow does not infiltrate before reaching the channel.

The physical nature and mechanics of the overland flow processes are described elsewhere by Chow (1964), and Eagleson (1970). The equations describing the process are also discussed by Wooding (1965), Grace & Eagleson (1966), Woolhiser, Hansen & Kuhlman (1970), and Overton (1971).

The application of the equations to natural watersheds is complicated by the presence of the litter layer and other low ground cover such as rocks, and low vegetation. However, Simons, Li & Stevens (1975) have recently developed a method for modeling lateral flow through or over litter which takes into account the resistance to flow from the litter and other low cover.

Plant water relations

Plants exert a unique influence upon the hydrologic cycle by extracting water from the soil below the zone affected by evaporation. This in turn influences the amount remaining for seepage and streamflow. Water moves through plants as a liquid in the conducting system and by diffusion across cell membranes. From the mesophyll tissue within the leaf, water is transformed into the vapor phase passing out through the stomata or the leaf cuticle to the atmosphere. The driving force for the movement of water through plants is a fall in water potential initiated in the leaves as a result of the loss of water by evaporation.

Specifically, this section describes (1) how roots extract water from the soil, (2) how the internal water balance of vegetation changes diurnally and seasonally and (3) how plant water relations interact with the environment to control transpiration. From this discussion, important structural and physiological characteristics of the vegetation are identified that aid in assessing the water balance of vegetation and its influence upon the hydrologic cycle.

Root zone water. In mature forests, plant roots usually have reached their maximum depth. Disturbing the vegetation results in untapped reservoirs of water until roots reoccupy the area. In assessing water uptake by vegetation it is thus necessary to have some knowledge of root distribution.

The distribution of tree roots is affected by the depth of soil and by the soil water regime. In areas where precipitation is adequate, trees often develop rather shallow root systems. In such ecosystems a brief period without precipitation may stress the vegetation more than in areas where drought is more common and roots occupy a greater volume of soil.

The pattern of water withdrawal from the root zone is similar for many plants. If the soil is fully charged, plants first remove water from the upper layers until it becomes difficult to extract (Woods, 1965). Thereafter lower layers are sequentially drawn upon. The pattern is similar for agricultural crops (Hsiao *et al.*, 1976), conifers and hardwoods (Krygier, 1971), and desert shrubs (Caldwell & Fernandez, 1975). For forest plants, the total volume of water available in the rooting zone is approximately that held between 0.1 and 15 bars soil water potential.

Measurements of plant water potential made before dawn usually represent an equilibrium between the water available in the soil and the internal

balance of the plant. This assumes that the stem is not frozen and that transpiration is not occurring. Under such conditions, curves similar to that presented in Fig. 4.13 result as soil water is withdrawn (Sucoff, 1972; Hinckley & Ritchie, 1973). Typically, a large fraction of the available water can be removed without causing a change in plant water potential. Assuming the rooting depth to represent the zone occupied by at least 90% of the root biomass, Running, Waring & Rydell (1975), defined the inflection point for an abrupt change in plant water potential to correspond with a depletion of 80% of the available water. Nocturnal uptake was restricted exponentially as the remaining water was extracted until at about -30 bars essentially no uptake was possible. Accordingly, only a fraction of the total soil profile need be recharged to permit plant water potentials to recover to values of -3 to -5 bars at night.

If plants vary greatly in their rooting depth they should be grouped into appropriate classes to better estimate root water extraction and its consequences upon the community water budget.

Internal storage of water. There is often a lag between transpirational losses from the foliage and uptake by the roots as indicated by sap flow at night

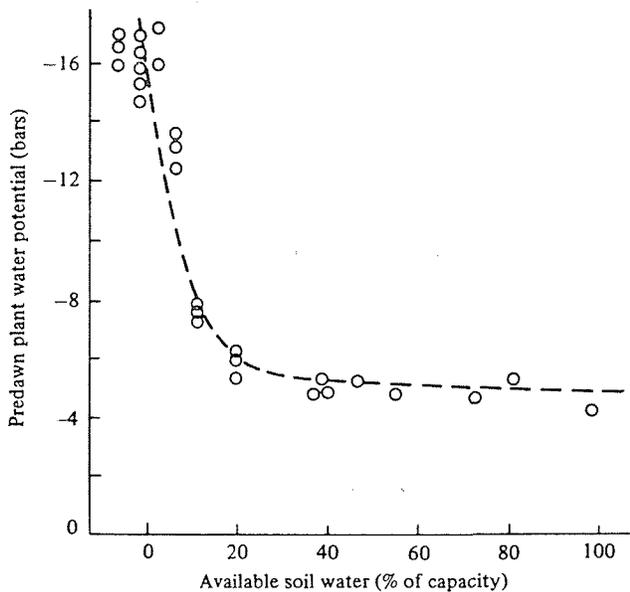


Fig. 4.13. Until more than three-quarters of the available water is depleted, red pine (*Pinus resinosa*) trees recover to the same predawn water potential. Thereafter, an abrupt decrease begins until all water is depleted. After Sucoff, 1972.

when there is no transpiration (Lassoie, 1973). A lag indicates water is not merely passing through a rigid conducting system but is being withdrawn from somewhere within the tree. Two kinds of tissue are involved: one shows volumetric change and is made up of extensible living cells; the other is predominantly non-living sapwood which shows no dimensional fluctuation as its water content changes.

The extensible tissue reserve, represented by foliage, phloem, cambium and new xylem cells has significance diurnally because it can provide supplemental water to maintain normal transpiration for one to several hours (Jarvis, 1975). Sapwood, on the other hand, contains a much larger volume of extractable water which is less readily available. In conifers where the sapwood exhibits up to a 50% change in volume of water, a mature forest may store 6 cm per hectare (Waring & Running, 1976). Different species of trees have varying proportions of sapwood to leaf area (A. F. G. Dixon, 1971; Grier & Waring, 1974) and therefore different amounts of sapwood. Although the amount of water withdrawn from the sapwood on a given day may be relatively small, under sustained demand the reserve can be emptied in a few weeks (Fig. 4.14) and recharge may be even more rapid (Fries, 1943). The sapwood of conifers can withstand repeated breakage of water capillaries because the vascular elements have valves (border pits) which seal off one cell from another when steep water potential gradient

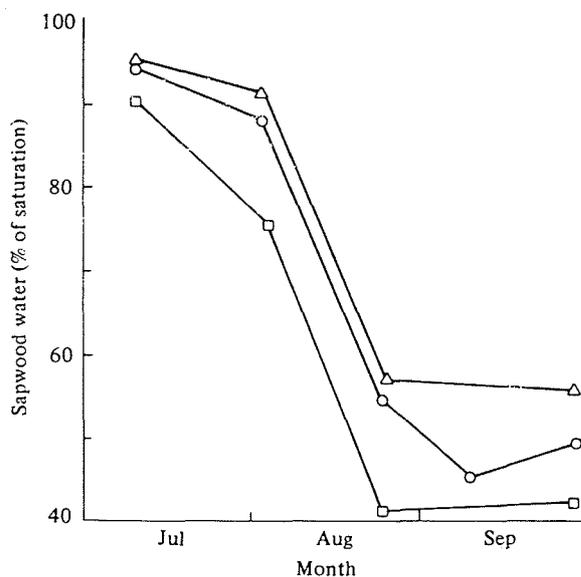


Fig. 4.14. Seasonal course of water extraction from different sapwood zones of Sitka spruce (*Picea sitchensis*) grown in Scotland. \triangle — \triangle , outer 2 cm; \circ — \circ , mid 2 cm; \square — \square , inner 2 cm. After Chalk & Bigg, 1956.

occur (Gregory & Petty, 1973). This also permits gas under high pressure to redissolve and the water column to be re-established following disruption by freezing or desiccation (Hammel, 1967). In general a forest of mature trees has greater internal reserves to draw upon than a younger forest with equal leaf area. This results in the older forest being less susceptible to climatic extremes affecting the water budget.

Transpiration. The process of transpiration refers to water loss through the stomata or cuticle of the leaves. Transpiration is directly proportional to the leaf area and the water vapor gradient between the air and leaf surfaces. It is inversely proportional to the canopy resistance, encompassing stomatal, cuticular resistance, and a boundary layer resistance. Usually the biological portion of these resistances are pooled as leaf resistance or its reciprocal, conductance.

The water flux from a canopy can be estimated with knowledge of the surface resistance using the Penman–Monteith equation (Monteith, 1965). If the surface area of the foliage is multiplied by the surface resistance of the canopy, an estimate of the average foliage resistance is obtained, subtracting any boundary layer resistance (Stewart & Thom, 1973). Fig. 4.15 illustrates the correspondence between these two expressions.

The Penman–Monteith equation, although complex, does not require specific knowledge of foliage temperatures because in the process of combining both mass and conductive transfer of heat and water vapor the

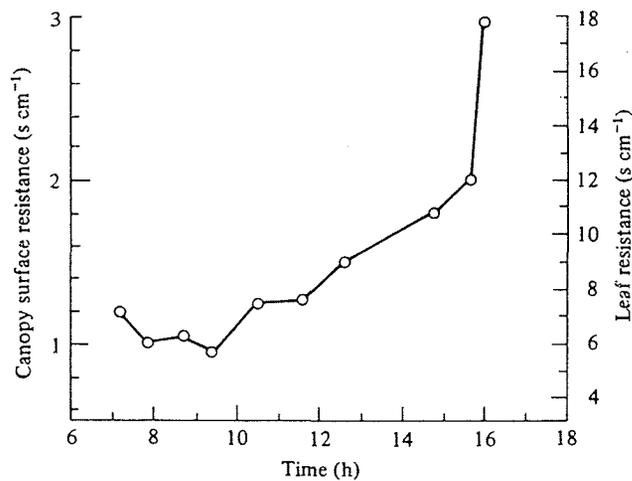


Fig. 4.15. In a Scots Pine (*Pinus sylvestris*) plantation canopy surface resistance increased throughout the afternoon corresponding to a stomata resistance change from 6 s cm⁻¹ to more than 18 s cm⁻¹. Values calculated for projected surface, not total leaf area. LAI=6. After Stewart & Thom, 1973.

temperature terms cancel. This makes the equation particularly useful in forests with broad leaf vegetation, for such leaves exhibit large boundary layer resistances resulting in leaf temperature well above ambient. The needle-shaped leaves of conifers, in contrast, have a small boundary layer resistance (usually less than 0.5 cm s^{-1}) and are thus within a few degrees of ambient temperature (Gates, 1968).

It is possible for broad leaf plants to exhibit some stomatal control without a reduction in transpiration. In fact, if the leaf temperature is elevated sufficiently, transpiration may even be increased in spite of partial stomatal closure (Gates, 1968).

Stomatal control by light, temperature, and humidity. Stomata respond to critical levels of light, temperature, and humidity, as well as to internal water deficits. The first three factors are mainly external in that they operate upon the guard cells of the stomata and need not affect the internal water balance of the leaf or other tissue.

Most plants have a minimum level of light which is necessary to trigger opening of the stomata. This is often equivalent to the carbon dioxide compensation point. Plants differ in this respect and also in the time it takes to respond to changes in light conditions. In general, fast growing pioneer species usually require greater amounts of light to open than slow growing advanced successional species. The former are also less responsive to sudden changes in light than the latter (Woods & Turner, 1971).

Light attenuation is difficult to estimate under a variable canopy but tends to follow Beer-Lambert's law in relation to increasing foliage density (Fig. 4.16). Lower foliage in the canopy can be expected to be more sensitive and to require somewhat less light for stomata to open.

Increasing the temperature may result in further opening of stomata, at least in some species (Schulze *et al.*, 1973). On the other hand, temperature causing freezing of the guard cells or any part of the conducting system will result in stomatal closure (Drew, Drew & Fritts, 1972). Indirectly, low root temperatures may affect stomata by inhibiting water uptake, resulting in a water deficit within the leaves (Bababola, Boersma & Youngberg, 1968; Havranek, 1972).

Low humidity, or more precisely a high evaporative demand can affect guard cells of the stomata and cause partial closure (Lange *et al.*, 1971; Watts, Neilson & Jarvis, 1976).

These external environmental controls may help in explaining differences in water use by forests. Climates characterized by high evaporative demand may have forests that exhibit stomatal control even with adequate soil water. This means that a higher proportion of the solar energy is dissipated as sensible or conductive heat loss than by way of evaporation. Similarly,

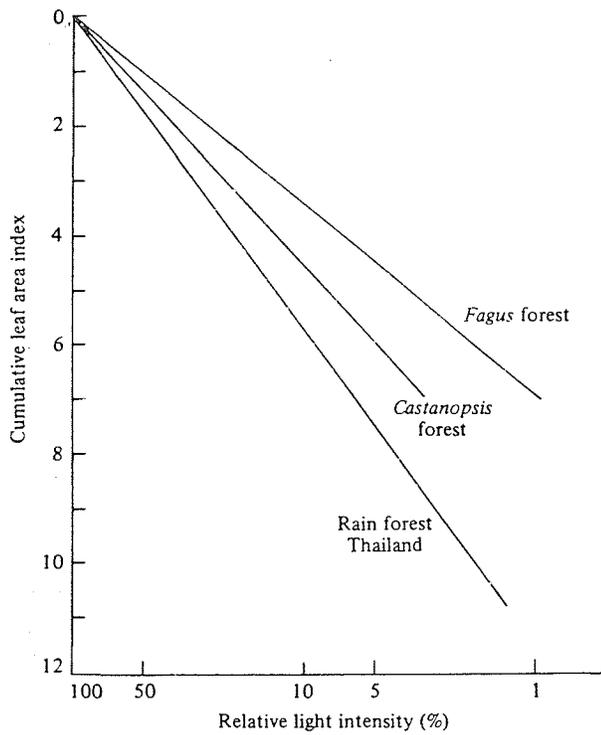


Fig. 4.16. Light attenuation in relation to leaf area follows a negative exponential curve as predicted by Beer-Lambert's Law. Kira, Shinozaki & Hozumi, 1969.

light limitations due to cloudy weather or a shaded position may result in less transpiration than expected by an analysis of evaporative demand alone. For modeling purposes, threshold and limits can be established to identify conditions that may be expected to alter the canopy resistance abruptly (Waring & Running, 1976).

How plant water potential and internal water deficits affect stomata. The evaporative demand, soil water supply, and resistance to water flow through the conducting system interact to affect the water balance of leaves. If insufficient water is supplied to the leaves to meet the evaporative demand, the immediate result will be an increase in the tension on the water column throughout the conducting tissue. This is indicated by an increase in the water potential gradient. Additional conducting tissue may be brought into use with the net result that water will flow faster and that uptake will balance demand. In such a situation there will be little volumetric change in extensible tissue, stomata will remain open, and the water potential may, at

least in some deciduous species, fall linearly with increasing transpiration (Landsberg *et al.*, 1975) as illustrated in Fig. 4.17.

If, however, the internal resistance becomes too great to supply water at the necessary rate, the deficit will be met first by withdrawal of water from the extensible tissue near the evaporative sites. As water is withdrawn it becomes more difficult to obtain additional water and untapped reserves in tissue further from the evaporative sites must be utilized (Gibbs, 1958; Doley, 1967; Schnock, 1972; Jarvis, 1975). The water potential gradients necessary to extract water from different tissue vary but curves such as those presented in Fig. 4.18 are typical of many forests (Hellkvist, Richards & Jarvis, 1974). Often a diurnal change of 10% (Clausen & Kozlowski, 1965; Gary, 1971) in the relative water content is observed in the foliage of trees. Under extreme conditions more than a 40% change in volume or relative water content occurs in some plants (Huck, Klepper & Taylor, 1970). As water becomes more difficult to obtain, the guard cells of the stomata are themselves depleted and this causes at least partial stomatal closure. The point where water becomes limiting can often be related to a particular water potential gradient. On forest samplings of *Pseudotsuga*

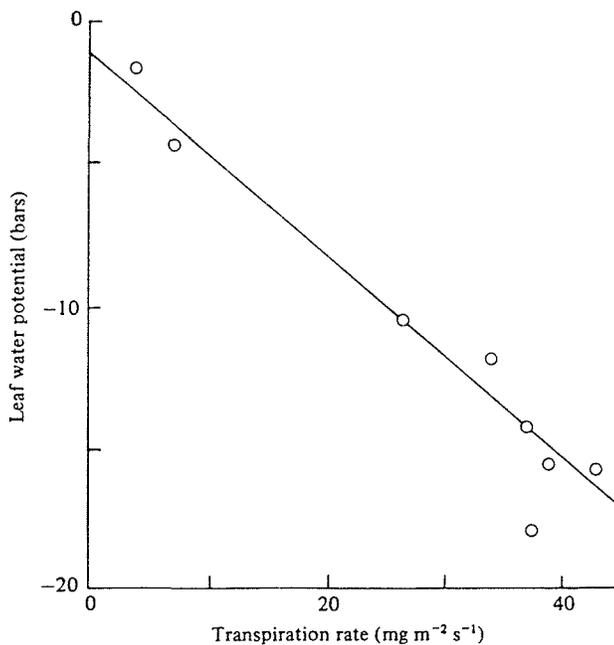


Fig. 4.17. In apple trees, increased transpiration may be sustained through an increased water potential gradient. After Landsberg *et al.*, 1975.

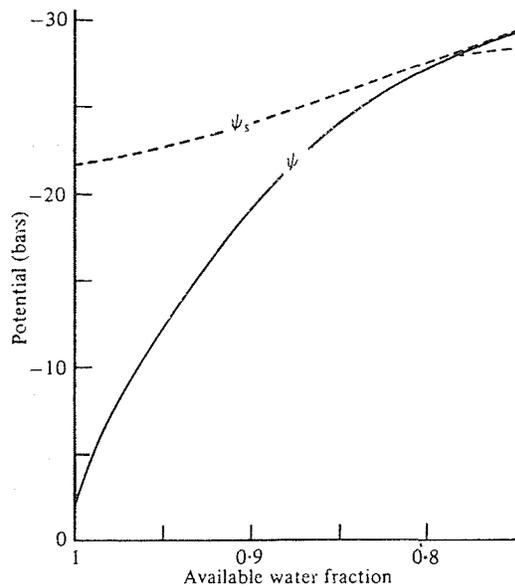


Fig. 4.18. Pressure extraction curve of water from twig of *Picea sitchensis*. After Hellkvist *et al.* 1974. ψ is water potential, ψ_s is solute potential.

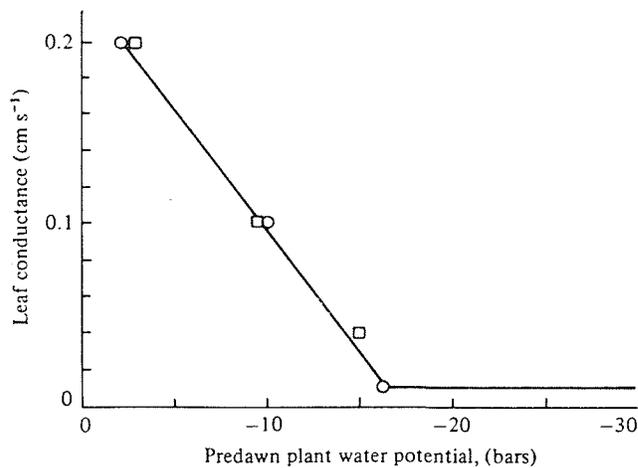


Fig. 4.19. Predawn plant water potential appears to control the maximum opening of stomata as illustrated by the relationship with leaf conductance of both hardwoods (Dinger, personal communication) and conifers (Running, 1976). Conductance, $1/r$, is calculated for all leaf surfaces.

menziesii and *Pinus ponderosa* the value is between -18 and -20 bars (Running, 1976; Lopushinsky, 1969). There are, of course, internal adjustments by the plants with height and season as a result of increased osmotic concentrations in the leaves (Richter, Halbwachs & Holzner, 1972).

A useful relationship has been established between predawn water potential and the maximum leaf conductance for both hardwoods and conifers (Fig. 4.19), (Running, 1976; Dinger, personal communication). This suggests that the water uptake by roots at night fully rehydrates the extensible tissue and that the relative water content of foliage controls the initial leaf conductance. If the evaporative demand during the day remains low, leaf conductance remains essentially constant (Running, 1976). On the other hand, a high evaporative demand may result in internal water deficits and lead to at least partial stomatal closure. A family of curves can be drawn or the average leaf conductance estimated for different kinds of days and initial conditions (Running *et al.*, 1975).

If root uptake become inadequate to balance losses through transpiration, more and more water is withdrawn from the sapwood. Most of the water is initially withdrawn from the interior zone with the outer 0.5 cm remaining saturated (Chalk & Bigg, 1956). Extraction may continue until only about 40% of the void spaces are filled with water. The water potential gradient may continue to increase to -50 or -60 bars which corresponds with lethal limits for conifer seedlings.

Key plant characteristics. From the above discussion we recognize certain key features of the vegetation that aid in interpreting how plants affect the hydrologic cycle. Structural features include an analysis of the vertical distribution of the canopy together with rooting volume and sapwood storage. Seasonal changes in leaf area, if large, should also be estimated. Some knowledge of threshold temperature, light and humidity effects are helpful if the vegetation is shown to react to typical climatic patterns. Monitoring of pre-dawn and mid-afternoon plant water potentials is desirable in order to indirectly calculate canopy resistances. These general features are incorporated in the structure of the simulation models discussed in later sections and were employed to estimate transpiration by individual trees by Running *et al.*, 1975).

Soil water balance

The physical processes controlling water movement in the soil are the result of strong surface and capillary forces with which the porous media hold water against the force of gravity. The soil water potential is a measure of the energy status of water in soil. This varies primarily with soil water content but is also influenced by temperature and chemical composition of the

water. The hydraulic conductivity is a measure of the ability of a soil to conduct water in response to gradients in soil water potential. It also varies strongly with soil water content. Both soil water potential and hydraulic conductivity are strongly influenced by physical characteristics of the soil such as texture, structure and pore size distribution. An example of soil water potential and hydraulic conductivity relation with water content is shown in Fig. 4.20 and 4.21.

Water moves from areas of high potential to areas of low potential at a rate dependent on the potential gradient and the conductivity. The direction of movement is controlled by the gradient. The conductivity is highest when the soil is saturated and can decrease by many orders of magnitude as the soil moisture decreases to field capacity (0.1 bars tension). This is because of the increased tortuosity of the path of water movement through the soil, as water must flow along the thin films on the soil particles.

Here we again need to recognize the possible existence of a channel system within the capillary system, as discussed previously. Evaporation and root uptake processes occur in the capillary system. Infiltration and percolation may occur in either the capillary or channel system. Storage may occur in either system but storage in the channel system would generally be very temporary. Lateral flow is a large scale form of redistribution. However, redistribution generally occurs within the capillary system or from the channel system to the capillary under the influence of potential gradients.

Forest soils are rarely isotropic and homogeneous, but are generally layered. Movement of moisture in layered soils is complicated by large changes in conductivities between strata. The boundaries between layers

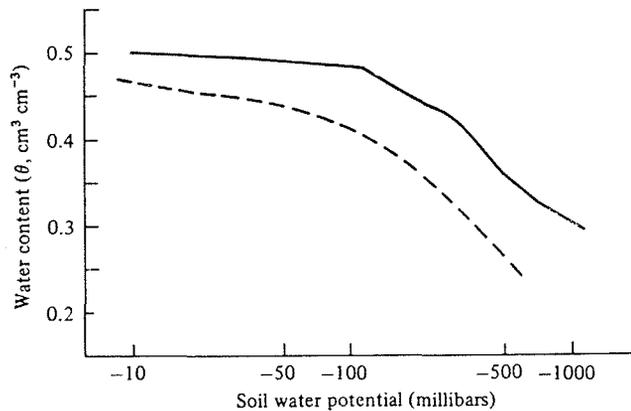


Fig. 4.20. Volumetric water content as a function of soil water potential for two surface horizons of a Honeywood silt loam. From Elrick, 1968.

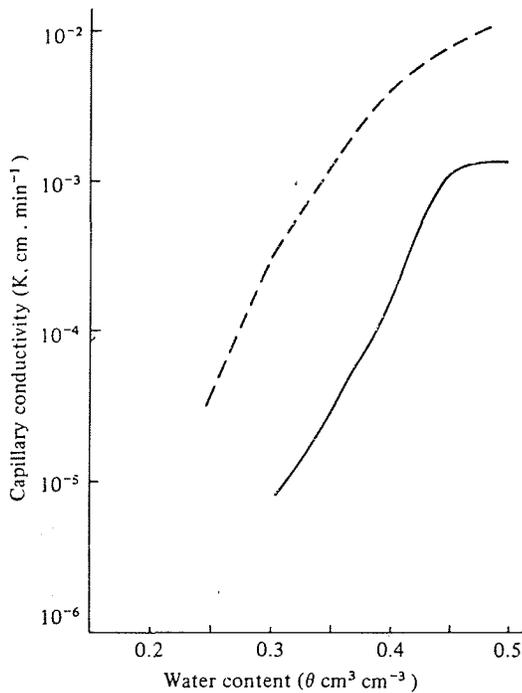


Fig. 4.21. Capillary conductivity as a function of water content for two surface horizons of Honeywood silt loam. From Elrick, 1968.

may be sharp or gradual. Infiltration into layered soils has been studied by Hanks & Bowers (1962), Miller & Gardner (1962), Miller (1969), Whisler, Watson & Perrens (1972), and Aylor & Parlange (1973). We find a significant difference in behavior between sharp and gradual stratification, particularly when a stratum of low conductivity overlies a coarse stratum of high conductivity. If the stratification is sharp the upper stratum must nearly saturate before significant water can move into the coarse stratum. In any case, the coarse stratum may never saturate in a very deep soil. For the reverse situation, infiltration to depths proceeds very rapidly with the eventual saturation of the upper layer. In either case, percolation to deeper layers is controlled by conductivity of the limiting layer.

The permeabilities of forest soils developed in place generally decrease with depth. Furthermore, they always have higher combined conductivities in a direction parallel to the slope than at right angles to the slope. This difference with direction is termed anisotropy. Its causes are discussed by Bear *et al.* 1968, and include the effects of biological activity as well as the formation of litter and humus layers. This almost assures that overland flow is a rare occurrence on forest watersheds (Freeze, 1972b)

Hence the infiltration process is generally similar to that of Figs. 4.22 and 4.23.

This situation is conducive to lateral downslope movement of water in the upper, high conductivity layers of the soil. This is particularly true in forested areas with high rainfall and steep slopes. Many investigations have found that the hydrologic response of a forested hillslope to rain is dominated by the lateral downslope movement of water within the soil mineral layers (Whipkey, 1967*a,b*; Hewlett & Hibbert, 1967; Dunne & Black, 1970; Nutter & Hewlett, 1971; Weyman, 1970, 3; & Megahan, 1972; Hornbeck, 1973; Hewlett, 1974). The phenomenon has been investigated on artificial slopes by Hewlett (1961) and Hewlett & Hibbert (1963).

Simulation studies were performed by Freeze (1972*a,b*) for hypothetical slopes and Stephenson & Freeze (1974) using actual data. Results indicate that subsurface lateral flow will be a significant contribution to storm runoff only under the conditions where convex hillslope feeds steeply incised channels. The surface soil horizons must have high permeabilities and as permeabilities decrease a threshold is reached below which subsurface flow cannot be an important component of storm runoff. These conditions are illustrated in Fig. 4.24. Freeze found that on convex slopes with low permeabilities and on all concave slopes, the storm runoff was dominated by direct runoff of precipitation on transient wetlands near the channels. These wetlands expand during storms as surface saturation occurs from

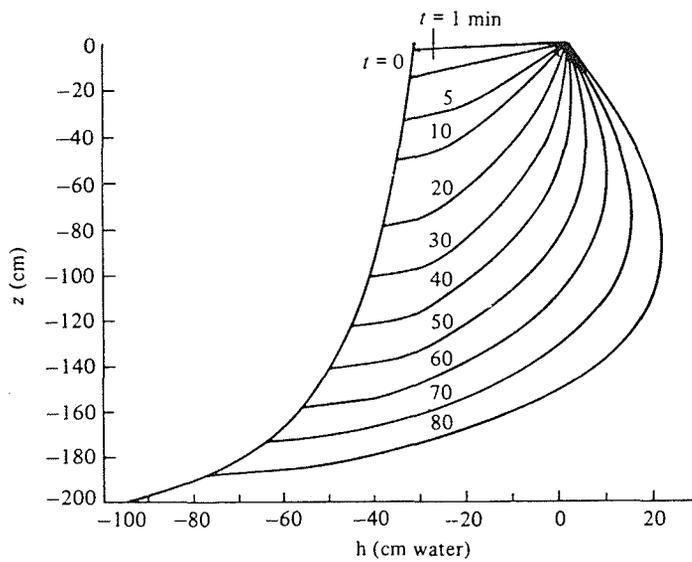


Fig. 4.22. Pressure head profiles during infiltration into a heterogeneous porous medium with saturated conductivity of 0.5 cm min^{-1} at the top uniformly decreasing to 0.05 cm min^{-1} at the bottom (-200 cm). From Whisler *et al.* 1972.

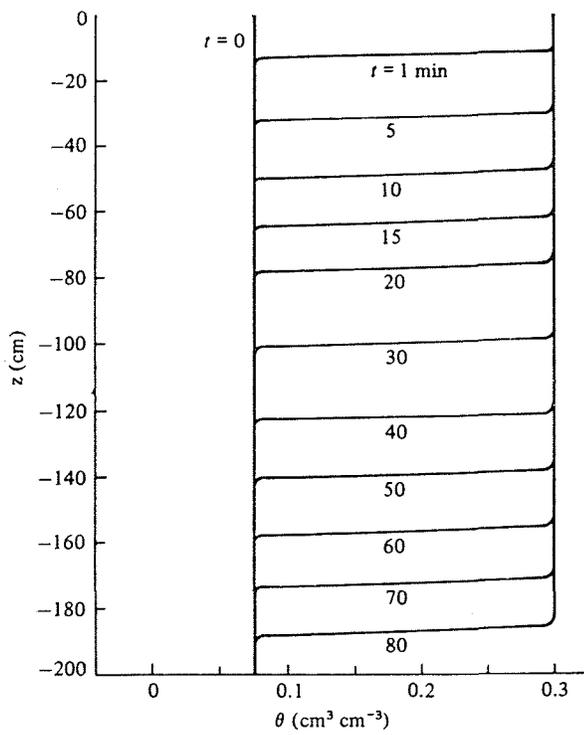


Fig. 4.23. Water content profiles for the infiltration case shown in Fig. 4.22. From Whisler *et al.* 1972.

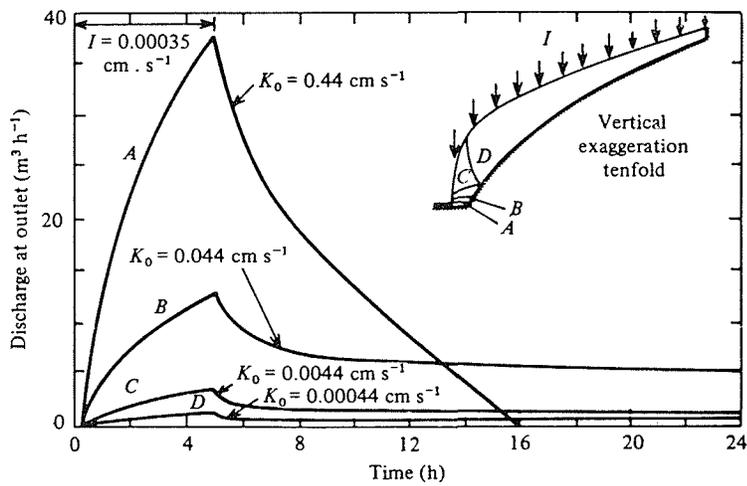


Fig. 4.24. Simulated stream hydrographs of runoff generated by subsurface stormflow at outlet of upstream source area. Source area has convex, 7.5% hillslope with maximum soil thickness of 100 cm. Results for four saturated conductivities are shown for a rainfall event of five hours duration and intensity (I) of 0.00035 cm/s. The inset show the cross-section of hillside and position of water table for each case at end of five hours. From Freeze, 1972b.

below due to infiltration toward very shallow water tables, and not from lateral subsurface flow. The behavior is compatible with the variable source area concept (Hewlett & Hibbert, 1967).

Freeze also concludes that a watershed just permeable enough to sustain subsurface flow to a channel, without formation of wetlands, will exhibit the minimum storm runoff response. In some cases this permeability may be so low that subsurface lateral flow just satisfies the evaporation demands from the channel.

For water balance purposes, it is useful to divide the soil into at least two layers: a root zone and a subsoil. Biological processes influence water movement in the root zone but purely physical processes are acting in the subsoil.

Root zone. The change in water content of the root zone is the sum of (1) infiltration, (2) rise of water from the subsoil, (3) subsurface lateral flow in, minus (4) evaporation, (5) uptake by roots, (6) percolation into subsoil, (7) subsurface lateral flow out, and (8) exfiltration. We have previously discussed infiltration, subsurface lateral flow, root uptake, and percolation, and will briefly discuss evaporation, rise, and exfiltration.

The drying of soil by evaporation has been studied by Lemon (1956), Philip (1956), Gardner (1959), and Kimball & Jackson (1971). There are three recognizable stages in the drying of soil. During the first stage, there is a rapid loss of moisture while capillary flow to the surface is adequate to meet evaporation demands. During the second stage, the surface dries and there is a rapid drop in the rate of loss of moisture because the rate of movement to the surface is controlled by intrinsic hydraulic characteristics of the soil. During the third stage the rate of loss is very low and is controlled by adsorptive forces at the liquid-solid interfaces. A litter layer can significantly reduce soil temperature and evaporation of soil water. Tomanek (1969) summarizes much of the work on effects of litter on evaporation from soil.

Water can move upward in response to potential gradients within the soil. This process is termed capillary rise. It most often occurs in response to water being either removed by evaporation and uptake processes, or added by the upward movement of a saturated zone due to a net lateral flow in. The process of water movement from the soil to the surface is termed exfiltration. It can be significant in maintaining a wet evaporating surface when a saturated zone is present close to the surface.

Subsoil. The change in water content of the subsoil is the sum of (1) percolation from the root zone, (2) subsurface lateral flow in, (3) rise from regional groundwater, minus, (4) rise to the root zone, (5) subsurface lateral

flow out, and (6) percolation (or seepage) to the regional groundwater. We previously discussed these processes with the exception of rise from and seepage to regional groundwater.

The interaction between the subsoil on a small forested watershed and the regional groundwater is dependent to a large extent on the geology of the area, the permeability of the bedrock beneath the subsoil, and the characteristics of the regional groundwater aquifers. Since we are restricting ourselves to small watersheds within a larger basin, we will not deal with capillary rise to the subsoil from regional groundwater, since in mountainous areas this probably is not significant. However, percolation can occur through the bedrock at a rate limited by the permeability of the bedrock. Freeze & Witherspoon (1966, 7, 8) and Freeze (1969a,b) provide comprehensive discussion and analysis of regional groundwater.

If the bedrock has low permeability, a saturated zone can form in the subsoil. This is termed perched groundwater because it exists or is 'perched' above the regional groundwater. It is often the source of small springs or seeps which may dry up in extended periods of drought.

Summary

This completes our discussion of the hydrology of an individual terrestrial unit. In summary, precipitation entering a unit passes first through a canopy consisting of foliage, branches, and stems. Some is held and evaporated and the balance directly or indirectly reaches the ground surface. It may be stored in a snowpack which eventually melts and delivers water to the surface. At the surface it may be held in the litter and evaporate or percolate through the litter to the surface where it may infiltrate. Excess water on the surface may runoff through the litter as overland flow, but this is rare in forested watersheds. Once in the soil, water can percolate downward and move laterally in response to potential gradients. Percolation can deliver water to regional groundwater. Within the root zone, water in excess of the wilting point is available for uptake by plants. This water may not immediately leave the system by transpiration but can be stored in the sapwood and extensible tissue reserves of the vegetation. Transpiration withdraws water from these reserves, reducing the conducting capabilities of the vascular system, which in turn creates a water deficit in the foliage and reduces transpiration by affecting stomata.

Coupling of terrestrial and aquatic units to describe a watershed system

Overland and subsurface lateral flow provides the basis for coupling terrestrial units to other terrestrial units as well as to aquatic units. Surface

Dynamic properties of forest ecosystems

topography determines the path of overland flow and is an indication of the probable paths of subsurface flow. The actual paths of subsurface flow may be modified by the local geology. Flow paths through the aquatic system are more easily determined.

Knowledge of the flow paths can be used to route both overland and subsurface flow from upper units to lower units and into the stream or aquatic units. This has been done in simulation studies by Simons *et al.* (1975), US Forest Service (1972) and Rogers (1973).

Comparison of total watershed behavior

We have described the water balance processes of individual terrestrial units. We are able to describe the path of movement of excess water between individual units and into the aquatic system. By putting this all together mathematically, in the form of a general hydrologic model, we will be able to reproduce and better explain a number of phenomena which we have observed in practice. This would contribute substantially to our theoretical understanding of the hydrology and water relations of watershed ecosystems. We will discuss two efforts in this direction – PROSPER and ECOWAT.

Comparison of the PROSPER and ECOWAT hydrology models

By comparing PROSPER and ECOWAT, we hope to illustrate the structure and capabilities of the models, and also clarify the major difference in order to place each in perspective.

PROSPER

PROSPER was developed within the Eastern Deciduous Forest Biome and is described by Goldstein, Mankin & Luxmoore (1974). The objective in development of the model was to simulate water relations in the soil-plant-atmosphere system emphasizing the evapotranspiration processes. Initial work was done in a deciduous forest watershed. Hence, PROSPER models only a portion of the hydrologic system depicted in Fig. 4.1; specifically the foliage, root zone and plant water relations components for a simple hydrologic unit. The current implementation does not consider internal storage in the sapwood and is generally for the deciduous forest area where it was developed. However this could be changed.

Swift *et al.* (1975) modified PROSPER slightly and used it to simulate evapotranspiration from both mature and clearcut, deciduous forest watersheds, and from a young pine plantation. We will review the results (p. 248),

ECOWAT

ECOWAT is a result of extensive refinements and revision of a model originally developed by Rogers (1973). It incorporates component models for plant water relations developed within the Coniferous Forest Biome. ECOWAT models all of the terrestrial water balance processes depicted in Fig. 4.1 and described on p. 241 with the exception of groundwater. It can model many units simultaneously, and routes excess water between units to the aquatic system. (It does not yet route flows through the aquatic system.) It can model units with either coniferous or deciduous vegetative cover. The current version of ECOWAT is documented by Rogers (1975) and we briefly describe the component models used in the following sections.

Canopy water balance component. The model which we use is similar to that developed by Rutter *et al.* (1975). We assume that the water holding capacity of the canopy is a function of the leaf, branch and stem area, as well as vegetation type (conifer or deciduous). We also assume that the capacity is the same for rain and snow, under the assumption that while more snow may be caught, the excess will eventually fall off before significant loss occurs. The model keeps track of amounts of rain and snow caught by the canopy, passing directly through without striking the canopy, and excess dripping from the canopy as well as evaporative losses from the canopy storage. We assume evaporation proceeds at the potential rate until storage is exhausted as determined by either Penman-Monteith equation, or Fick's Law (method is user selected).

Snowpack component. A good snowpack model requires consideration of both the water and energy balances of the snowpack. The model we use is based primarily on the work of Anderson (1968), with some modifications to allow for pack settlement, density changes, albedo changes, and water holding capacities.

Briefly, we consider the energy balance of the thin surface layer of the snowpack. Heat gained in this layer produces melt which, together with any rain, percolates into the pack. If the pack has a heat deficit, the water will freeze and release heat which warms the pack, eventually bringing it to an isothermal condition. When the pack is isothermal additional water will fill the water holding capacity and the excess will runoff through the bottom of the pack. A heat loss from the surface layer will cool the pack, and excess water will freeze to satisfy any heat deficit until all excess water is frozen and the pack develops a heat deficit.

Litter water balance. The litter water balance operates very similarly to the canopy water balance. It recognizes the several drying stages of litter. The

first is where moisture evaporates at the potential rate, the second where the rate of evaporation is limited by an increasing surface resistance, and the third where the hygroscopic properties are important and moisture transfer may be in either direction. The model gives amounts evaporated from storage, passing through the litter, and passing through openings in the litter without being caught by the litter.

Surface water balance component. This component deals with the processes of overland flow, detention storage and infiltration. It is closely coupled with the soil water balance component. We currently assume that anything in excess of the detention storage capacity flows out of the unit. A provision for real time routing will be incorporated later. Hoitani's equation for infiltration as used in the USDAHL-70 model (Hoitani & Lopez, 1971) is the basis for estimating infiltration. Routing of overland flow between units and to the stream is also handled by this component.

Soil water balance component. This component incorporates the processes of percolation, lateral flow, and non-capillary flow. Uptake is incorporated in the plant water relations component. We subdivide the soil horizon into up to four layers, depending on the rooting depth and soil properties. We assume that all units have the same number of layers, but not necessarily of the same thickness.

We assume that the soil water conductivity-soil moisture relationship is a log-log relationship. This could be replaced, but we use this relation because data are rarely available under most conditions. We assume that all water in excess of the capillary water holding capacity (which may be the saturated water holding capacity) flows laterally down to the next unit. Water left then moves at a rate determined by the conductivity. It is divided into percolation down to the next layer, and lateral flow out to the next unit, on the bases of slope and conductivities of both layers. The method for doing this is derived from work of Zavlavsky & Rogowski (1969). We iterate to get the average conductivity for each time step. Lateral flow between units and to the stream is also handled in this component.

Plant water relations. This component incorporates the processes of uptake, transpiration, internal storage, moisture stress, leaf resistance and surface resistance. It is applicable to deciduous cover and coniferous cover with the primary difference being that we assume deciduous cover has no effective internal storage of water. In other words deciduous plants essentially depend on uptake to meet transpiration demand, while coniferous plants have the capability of satisfying demand from internal storage, primarily in the sapwood.

The method used is essentially an adaptation of an individual tree model

developed by Running *et al.* (1975). We first estimate plant moisture stress as a function of temperature and readily available moisture. We then estimate leaf resistance and surface resistance. For coniferous species we then take up water from the soil to satisfy the internal storage, or sapwood water deficit. Uptake proceeds first with readily available water from the top layers downward. The water holding capacity of the sapwood is a function of the leaf area and tree height. For deciduous species we determine transpiration demand and then uptake to meet the demand if possible, and the actual transpiration is then equal to uptake. For coniferous species we meet transpiration demand from sapwood storage if possible. Transpiration demand is determined on the basis of the Penman-Monteith equation or a modification of Fick's Law, both of which include surface resistance terms.

Discussion

It should be obvious that ECOWAT is directed toward a different type of application from PROSPER. PROSPER could be a component of ECOWAT, replacing the existing water relations routine. PROSPER can also be used to study water relations on a small site where other components not included in the model, such as snow, are not important. ECOWAT can be directed toward applications on both larger watersheds where all components and heterogeneity are important; and on small sites, where snow and other components are important.

Applications of the models

In order to evaluate the models we made initial applications of ECOWAT in three areas and PROSPER in one area. Both were applied to several vegetation cover conditions in each area, in order to evaluate the utility of the models for studying and predicting total watershed behavior. In general, the cover conditions studied in each area included (1) the original undisturbed vegetation, (2) a reduction in original vegetation density as a result of thinning or clearing, and (3) a change in type of vegetation cover.

Description of the watersheds

The areas in which the models were applied were (1) the Coweeta Hydrologic Laboratory in North Carolina, (2) the H J Andrews Experimental Forest in Oregon, and (3) the Beaver Creek Experimental Watershed in Arizona. These represent three sites with significant differences in climate, vegetation, soils, and geology.

The Coweeta Hydrologic Laboratory. The Coweeta Hydrologic Laboratory is located in western North Carolina. It was established in 1933 and contains a network of 31 gaged watersheds on which a series of vegetation alteration experiments have been made. These are described by Hewlett & Hibbert (1967), Swank & Miner (1968), and Swank & Helvey (1970).

In general, the area is located in the high precipitation (2% snow) region of western North Carolina. It is characterized by deep, permeable soils underlain by granite bedrock, steep slopes, and an original cover type of oak-hickory.

Our tests were made on Watershed 18 using climatic and streamflow data for the period May 1971, through April 1973. Watershed 18 is a 12.5-ha control watershed with mature oak-hickory forest which has not changed in over 40 yr. The growing season leaf area index is 5 (one sided) and dormant season is 0.5. It has a north facing slope with elevation ranging from 721 to 977 m. Precipitation for 1971-2 water year (May-April) totaled 199 cm and 1972-3 was 234 cm. Complete climatic data are available.

Soils data were obtained from Luxmoore (1972). Albedo values were taken from Swift (1972), and interception storage data for canopy and litter from Helvey (1967). The watershed is very homogeneous and was represented by one response unit with no stratification.

The H J Andrews Experimental Forest. The H J Andrews Experimental Forest is a 6000-ha watershed in the western Cascades of Oregon. The drainage is characterized by steep topography, with about one-fifth of the study area consisting of more gentle slopes or benches. Elevation varies from 457 m to more than 1523 m. Mean forest air temperature varies from 2°C in January to 18°C during the summer months. Annual precipitation ranges from 225 cm at lower elevations to 350 cm at the highest ridges. Highest elevations are characterized by extensive snowpack during the winter, while rain predominates at lower elevations.

Watershed 10 is 10.2 ha and is located on the edge of the H J Andrews Experimental Forest. Elevations on the watershed range from 430 m at the stream gaging station to about 670 m at the highest point. Slopes on the watershed average about 45% but frequently exceed 100%.

The study site is located in an area underlain by volcanic tuff and breccia. Soils of the watershed are derived from these materials and range from gravelly, silty clay loam to very gravelly clay loam. The <2-mm fraction of these soils ranges from 20% to 50% clay and contains gravel amounting to 30-50% of the soil volume. The forest floor ranges from 3 to 5 cm thick and is classified as a duff-mull.

The present overstory vegetation is dominated by a 60- to 80-m-tall, 450-year-old stand of Douglas fir (*Pseudotsuga menziesii*) containing small islands of younger age classes. Distribution of understory vegetation reflects topography and slope-aspect on this watershed. Dry ridge-tops and south-

facing slopes have an understory composed primarily of chinkapin (*Castanopsis chrysophylla*), Pacific rhododendron (*Rhododendron macrophyllum*), and salal (*Gaultheria shallon*). More mesic parts of the watershed support an understory of vine maple (*Acer circinatum*), rhododendron, and Oregon grape (*Berberis nervosa*, with a well-developed intermediate canopy of *Tsuga heterophylla*). Subordinate vegetation of the moist areas along the stream and on north-facing slopes is primarily vine maple and sword fern (*Polystichum munitum*). The average leaf area index is 25 to 30 (two sided) for all sites. Tree basal area averaged 76 m² ha.

The climate of Watershed 10 is typical for the western Oregon Cascades. Average annual precipitation is 230 cm per year with over 75% of the precipitation falling as rain between October and March. Snow accumulations on the watershed are not uncommon, but seldom last more than two weeks. Based on two years' data, the average daytime temperature for July is 21°C and for January is 0°C. Observed extremes have ranged from a high of 41°C in August to a low of -20°C in December. Complete climatic data are available.

Stream discharge varies from around 0.23 litre sec in the summer to about 140 litre sec during winter freshets. The uppermost forks are intermittent during the summer months. Mean width of the stream channel ranges from 0.25 m in the upper reaches to 0.75-1.0 m at the base of the watershed.

The Beaver Creek watershed. The 110 000-ha Beaver Creek watershed is located in Central Arizona. It was established in 1957 and now contains a network of about 44 watersheds ranging in size from about 10 ha to about 6400 ha. About one-third of the watershed is ponderosa pine and the balance Utah and Alligator juniper.

In the pine type, elevations vary from about 2073 m to 2439 m. Underlying bedrock consists of igneous rock of volcanic origin. Soils are developed on basalts and cinders, and are mostly silty clays and silty clay loams less than 0.75 m deep. January temperatures average -3°C and July temperatures average 19°C. Precipitation averages 63.5 cm annually with 64% falling during winter, October-April. Winter snow regimes are quite variable and range from a continuous snowpack in heavy snowfall years, to an intermittent pack in light snowfall years. Average annual streamflow has varied from 0.5 to 41.6 cm with an overall average of 13.5 ± 1.3 cm with 93% coming during winter (October-April).

The forest is composed of a mixture of ponderosa pine (*Pinus ponderosa* Laws.), Gambel oak (*Quercus gambelli* Nutt.) and alligator juniper (*Juniperus deppeana* Steud). Of a combined basal area of 26 m² ha⁻¹ pine occupies 21 m², oak 4 m², and juniper 1 m². A more detailed description is available in Brown *et al.* (1974).

Leaf area data were lacking on Beaver Creek, as were solar radiation

wind and good dew point data. We estimated solar radiation, used a constant wind speed and estimated dew point from available humidity data of unknown quality. Good data should be available for later years. Our leaf area estimate of 4.0 (all sides) is very tentative and we hope to improve this using relationships of foliage mass to sapwood area developed by Grier & Waring (1974).

Comparison of reduction in vegetation density

A reduction in cover density may take two forms (1) uniform reduction by thinning the vegetation in a uniform way, and (2) non-uniform reduction by removing clumps, strips or patches of the vegetation so as to create openings in the original cover. The two forms may be combined. The significant difference is that a light uniform thinning may not create openings in which the soil is unoccupied by roots, while non-uniform thinning does if the openings are large enough. This can result in significant differences in transpiration. A very heavy thinning will create openings and of course both can be carried to the extreme of clearing of all vegetation.

Other effects include (1) reduction in surface area of foliage, branches and stems, (2) increased penetration of solar radiation through the canopy to the snow or soil surface, (3) possible effects on distribution of snowpack and (4) possible disturbance or exposure of the soil surface which could lead to reduced infiltration capacity (Goodell, 1967).

One-year-old clearcut at Coweeta. Both PROSPER and ECOWAT were used to simulate a one-year-old regrowing clearcut on Coweeta Watershed 18. This clearcut was represented as a stand of sprout and seedling regrowth with growing season leaf area index of 0.75 (one sided) and dormant season of 0.5. Simulations for the original hardwood forest cover provide a base for comparison. Results are given in Table 4.2 with ECOWAT runs being *e*- and *i-j* for original hardwoods and clearcut, respectively; and PROSPER runs being *g-h* and *k-l*, respectively. Not all evaporation and transpiration results were available for the PROSPER runs. The differences in total precipitation should be noted when comparing ECOWAT and PROSPER runs.

For the original oak-hickory conditions the maximum error in simulated streamflow was 8.1% (11 cm) with all others within 2%.

Three clearcut treatments on north- or east-facing Coweeta watershed produced first year streamflow increases ranging from 36.0 to 41.3 cm. The increases determined by simulation for water years 1972 and 1973 were 36.0 and 38.0 cm, respectively, for ECOWAT; and 36.5 and 42.2 cm, respectively, for PROSPER. Differences in climatic conditions and the shape, aspect and elevation of the treated watersheds prevent direct comparison of the

Table 4.2. Summary of the results of simulation runs under different conditions (all values in cm)

Watershed	No. and condition	Water year	Precipitation		Evaporation				Transpiration					Total	Streamflow			
			Rain	Snow	Canopy	Litter	Snow	Total	Infil- tration	A	B	C	D	Total	evapotrans- piration	Predicted	Actual	
Beaver Creek																		
	<i>a.</i>	Original	1965	48	50	10.6	1.2	2.5	14.3	69.7	33.7	8.8	8.3	6.4	57.2	71.5	26.0	33.1
	<i>b.</i>		1973	45	74	13.7	1.5	3.9	19.1	70.3	28.6	7.8	8.3	4.9	49.5	68.6	56.3	55.4 ^a
	<i>c.</i>	Thinned	1973	45	74	10.6	1.6	3.0	15.2	67.8	23.3	5.8	6.4	1.4	36.9	52.1	67.6	64.3
	<i>d.</i>	Shrubs	1973	45	74	9.7	1.6	3.6	14.9	67.5	18.8	5.8	6.0	0.4	31.0	45.9	71.7	70.1 ^a
Coweeta																		
	<i>e.</i>	Hardwood	1972	192	3	24.4	1.4	0.0	25.8	168.0	53.7	4.4	4.3	---	62.4	88.2	103.0	104.3
	<i>f.</i>		1973	235	5	24.2	1.3	0.0	25.3	205.0	51.0	6.6	0.4	---	58.0	83.3	147.0	136.0
	<i>g.</i>	Hardwood	1972	199	0	ε	ε	ε	24.5	ε	ε	ε	ε	---	66.6	91.1	103.6	104.3
	<i>h.</i>		1973	234	0	ε	ε	ε	22.8	ε	ε	ε	ε	---	66.0	88.8	138.0	136.0
	<i>i.</i>	Clearcut	1972	192	3	8.6	2.4	0.0	11.0	182.0	41.9	0.0	0.0	---	41.9	52.9	139.0	142.5 ^d
	<i>j.</i>		1973	235	5	8.5	2.2	0.0	10.7	218.0	34.3	0.0	0.0	---	34.3	45.0	185.0	ε
	<i>k.</i>	Clearcut	1972	199	0	ε	ε	ε	ε	ε	ε	ε	ε	---	ε	ε	140.1	142.5 ^d
	<i>l.</i>		1973	234	0	ε	ε	ε	ε	ε	ε	ε	ε	---	ε	ε	180.2	ε
	<i>m.</i>	Pine	1972	192	3	30.9	1.4	0.0	32.3	163.0	73.4	4.2	0.0	---	77.6	109.9	84.4	84.1 ^d
	<i>n.</i>		1973	235	5	30.4	1.3	0.0	31.7	200.0	65.3	6.3	1.8	---	74.2	105.9	122.0	117.7 ^d
	<i>o.</i>	Pine	1972	199	0	ε	ε	ε	34.2	ε	ε	ε	ε	---	82.2	116.4	83.4	84.1 ^d
	<i>p.</i>		1973	234	0	ε	ε	ε	31.9	ε	ε	ε	ε	---	76.5	108.4	121.1	117.7 ^d
H J Andrews																		
	<i>q.</i>	Original	1973	119	48	32.9	0.6	-1.5	32.0	134.0	30.0	7.5	5.0	---	42.5	74.5	82.7	80.1
	<i>r.</i>		1974	206	98	22.3	0.7	-2.2	20.8	269.0	19.0	7.5	7.1	---	33.6	54.4	267.0	258.5
	<i>s.</i>	Clearcut	1973	119	48	2.3	3.3	-2.4	3.2	163.0	18.8	0.1	---	---	18.9	22.1	130.0	ε
	<i>t.</i>	Upper 37% cut	1973	119	48	21.0	1.7	-1.7	21.0	145.0	25.9	5.0	2.7	---	33.5	54.5	98.0	ε
	<i>u.</i>	Lower 10% cut	1973	119	48	29.9	0.8	-1.7	29.0	137.0	28.9	6.6	4.5	---	40.1	69.1	87.9	ε

^a Estimates based on results from other watersheds at Beaver Creek (Brown *et al.* 1974).

^b Results of simulations with PROSPER as given in Swift *et al.* (1975).

^c Not available.

^d Estimates based on results from other watersheds at Coweeta as presented in Swift, Swank *et al.* (1975).

simulation results with measured treatment effects. Water year 1973 was a record year for precipitation at Coweeta.

The ECOWAT simulations showed reduction in evaporation of 58% for both years and in transpiration of 33 to 41%. The reduction in evaporation is due to reduced storage capacity of foliage, branches and stems. The reduction in transpiration is proportionally less than that of evaporation because increased exposure of the remaining foliage results in much higher rates of transpiration per unit of leaf area (Landsberg *et al.*, 1975).

Heavy thinning at Beaver Creek. We used ECOWAT to simulate measured treatment effects on the 120 ha Watershed 17 which is typical of the pine watersheds on Beaver Creek. This watershed was thinned by group selection in 1969. 75% of the initial 27.5 m² basal area was removed by thinning leaving even-aged groups with average basal area of 7 m² ha for all species. All Gambel oaks over 38 cm dbh except den trees were removed, leaving 1 m² ha. All junipers were removed. Slash was windrowed. We used a leaf area index of 4.0 (all sides) for 1965 and of 2.0 for 1973. Pre- and post-treatment conditions were simulated for the water year October 1972 to September 1973, and pre-treatment conditions for water year 1965. For simulation purposes the watershed was stratified into three response units. The results are shown as runs a-c in Table 4.2. The model underestimated flow by 21% for water year 1965 and overestimated flow by 5% for water year 1973.

The estimated increase streamflow due to thinning for water year 1973 is 8.9 cm. The simulation showed an increase of 11.3 cm or 20%. This was due to a 20% reduction in evaporation and a 25% reduction in transpiration. These reductions are attributable to reduced storage capacity of foliage, branches and stems, and increased surface resistance to transpiration.

Almost all transpiration at H J Andrews and Coweeta is taken from the top soil layer. The pine stand at Beaver Creek, however, is dependent on lower layers to meet transpiration demand as shown in Fig. 4.25. Uptake proceeds first from the top 20 cm which is soon exhausted and withdrawal from lower layers begins. Rain from summer storms recharges the top layer in July and August and this then provides part of the demand.

Both the Coweeta and H J Andrews site rarely have snowpacks which persist. However, Beaver Creek can have a persistent pack and Fig. 4.25 represents a record year for both precipitation and snow. The peak in infiltration in May represents the melting of the snow pack.

Partial cutting at H J Andrews. We simulated effects of clearing different parts of a watershed on the H J Andrews Watershed 10. Original conditions were simulated for water years 1973 and 1974 and compared to measured

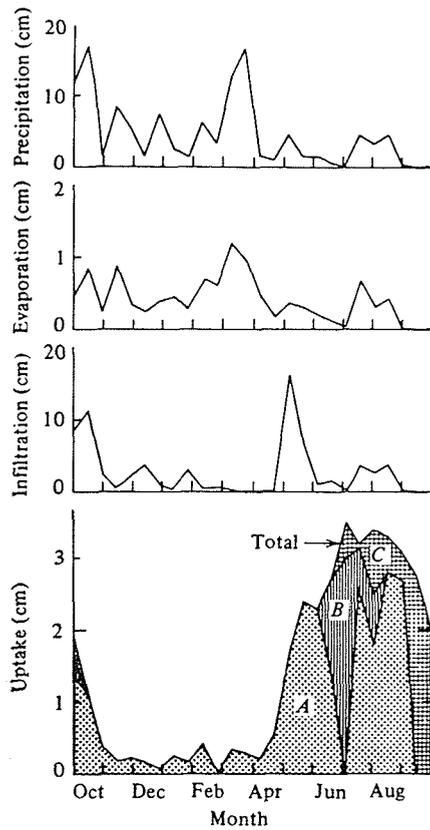


Fig. 4.25. Actual precipitation, and simulated evaporation from foliage, infiltration and root uptake from A 0 to 20 cm, B 20 to 40 cm, and C 40 to 100 cm at 14-day intervals for Beaver Creek Watershed 17, October 1972 through September 1973. The simulation represents a thinned stand of Ponderosa pine with total leaf area of 2 and average height of 32 m.

results. We then simulated effects of (1) clearcutting the entire watershed, (2) clearcutting the upper slopes, and (3) clearcutting the lower slopes.

The watershed was stratified into six units for purposes of simulation. Essentially the north and south slopes were divided into upper (dry ridgetops), middle (mesic) and lower (moist) zones. All flow from the upper units passes through the lower units before reaching the stream.

Clearcutting a unit was represented by a deciduous stand with leaf area of 0.75 in summer and 0.5 in winter. This was the same as in the Coweeta clearcut. Clearcutting of the upper units affected 37% of the total watershed area while clearcutting the lower units affected 10% of the area.

The results for the original conditions are given as runs *q* and *r* in Table 4.2. The predicted streamflow was high by 1.6 cm (2.0%) for water year 1973

and high by 8.5 cm (3.3%) for water year 1974. We used the 1973 simulation (run *g*) as the base for comparing treatment effects.

The results of simulations of complete clearcut, upper slope clearcut, and lower slope clearcut are given in Table 4.2 as runs *s*, *t* and *u*, respectively. The total clearcut resulted in a 47.3 cm (57.2%) increased streamflow with a reduction in evaporation of 90% and in transpiration of 55.5%. Clearcutting the upper 37% of the watershed resulted in a 15.3 cm (18.5%) increased streamflow with an overall reduction in evaporation of 34.4% and in transpiration of 21.2%. Clearing the lower 10% of the watershed resulted in 5.2 cm (6.3%) increased streamflow with an overall reduction in evaporation of 9.4% and in transpiration of 5.6%. Cutting the lower 10% of the watershed resulted in a larger relative response with respect to the area cut than cutting the top 37%. Hence, streamflow response is not necessarily directly proportional to the area cut.

The difference in response was caused by differences in orientation of the areas cut which caused differences in evaporation and transpiration. Under the original conditions, transpiration from the two upper units was 43 and 41 cm; and from the two lower units, 49 and 40 cm. Under the cut condition, transpiration was 20 and 18 cm from the upper units and 20 and 19 cm from the lower units. Hence the lower units had a larger decrease in transpiration (and in evaporation) than the upper units.

This emphasizes the need to stratify watersheds into response units in areas where heterogeneity of soils and topography has significant effects on the vegetation and the hydrologic response of those units.

It should be noted that in order to simulate a change in vegetation density, it is only necessary to change the values of no more than five input to ECOWAT for each unit. These are (1) leaf area for growing season, (2) leaf area for dormant season, (3) stem area, (4) average tree height, and (5) litter capacity. Of these the leaf areas have by far the greatest effect and demonstrate the need for obtaining good estimates of this variable. Change in stem area and litter capacity have some effect but are minor compared to changes in leaf area.

Comparison of changes in vegetation type

Changes in cover type can have many forms and be the result of natural (successional), catastrophic (fire), timber harvest, or other management activities such as revegetation, seeding, planting, or brush control activities. In any case, they can have more significant effects on water balance processes than simple reduction in leaf area. Effects can be due to change in (1) amount and seasonal distribution of leaf area, (2) density and depth of rooting, (3) season and duration of active growth, (4) albedo and aerodynamic characteristics of the vegetation. These and others are discussed below.

Goodell (1967), Swank & Miner (1968), and Swank & Douglas (1974).

Conversion to conifers at Coweeta. We simulated conversion of an oak-hickory forest to white pine at Coweeta using both ECOWAT and PROSPER. The conversion was represented by a stand of 16-year-old white pine with growing season leaf area index of 12 (surface with stomata) and dormant season of 6. The simulations for the original forest provide a base for comparison. Results are given in Table 4.2 with ECOWAT runs being *e-f* and *m-n* for original hardwoods and conifer, respectively; and PROSPER runs being *g-h* and *o-p*, respectively. Not all evaporation and transpiration results were available for PROSPER runs, and total precipitation was different for the ECOWAT and PROSPER runs.

The simulation results are compared to measured results from a watershed with a white-pine plantation on a north-facing slope adjacent to the simulated Watershed 18. Simulation showed decreases in streamflow for water years 1972 and 1973 of 18.6 and 25.0 cm, respectively, for ECOWAT; and 20.2 and 16.9 cm, respectively, for PROSPER. The measured decreases were 20.2 and 18.3 cm, respectively.

The ECOWAT simulations showed increases in evaporation of 25% for both years and in transpiration of 24 to 28%. PROSPER showed evaporation increases of 40% and transpiration increases of 16 to 23%. The increases in total evapotranspiration were 25 to 27% for ECOWAT and 22 and 28% for PROSPER.

The increase in evaporation is due to increased foliar storage capacity during the dormant season when hardwoods are normally bare. Summer transpiration by pine and oak-hickory was similar except that pine transpiration rates increased about one month earlier than hardwoods. Simulation indicates more than half of the increase in total evapotranspiration from pine was due to dormant season transpiration.

Conversion to deciduous shrubs and grass at Beaver Creek. We also simulated a conversion from mature ponderosa pine to grasses, forbs, shrubs and oak sprouts on Beaver Creek Watershed 7 using ECOWAT. This represented the current condition on Beaver Creek Watershed 12 which was cleared in 1966. The conversion was represented as a deciduous stand with growing season leaf area of 2 (one sided) and dormant season of 0.5. The results are shown in Table 4.2 with the original condition as run *b* and the conversion as run *d*.

The simulation shows an increase in streamflow of 15.4 cm (27%) compared with an estimated increase based on measured results of 14.7 cm (27%). Evaporation decreased by 22% (4.2 cm) and transpiration decreased by 37% (18.5 cm). This was due to reduction in foliar storage capacity, increased surface resistance, and reduced leaf area in winter.

Conclusions

We believe that models which represent the processes that control water movement and storage, in a way consistent with the structure and function of terrestrial ecosystems, simulate the hydrologic regimes of these ecosystems under a variety of conditions. The last objective provided convincing evidence to support our belief. The processes described and incorporated into the ECOWAT and PROSPER models gave very encouraging results. Fig. 4.26 compares predicted versus observed annual streamflow for sites at which ECOWAT was tested. These predictions were made without benefit of prior calibration of the model. With both models the annual totals of simulated streamflow agreed closely with measured streamflow for the three areas under different vegetation and climatic conditions.

Streamflow from a watershed is influenced by both events in past months and current climatic and vegetative conditions. However, vegetative conditions are also dependent on events in past months as well as the current hydrologic and climatic conditions. The ability of the models to estimate day-to-day changes in the hydrologic regime can be improved by including

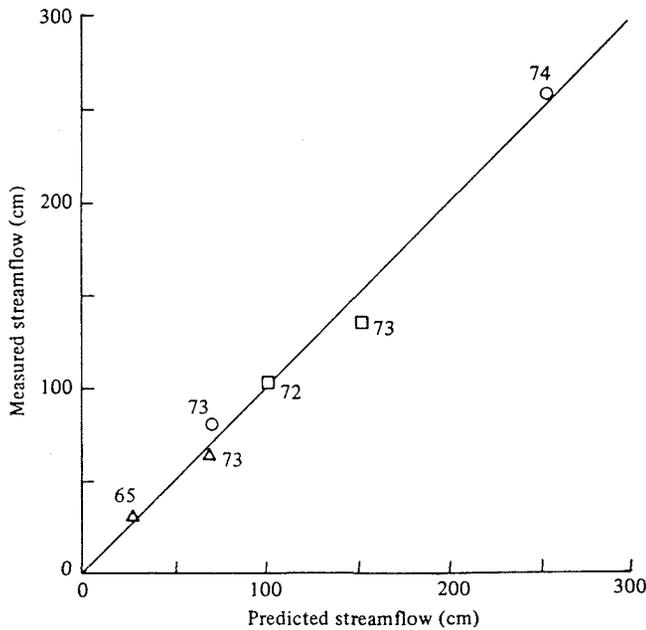


Fig. 4.26. Comparison of measured annual streamflow versus streamflow predicted by ECOWAT. The line represents perfect prediction. The r^2 is 0.995. □, Coweeta, Watershed 18; △, Beaver Creek Watershed 17; ○, HJ Andrews, Watershed 10.

corporating models for the short and long term response of the vegetation on the watershed to both the treatment and to the current hydrologic and climatic conditions.

Both simulations revealed research needs. The models are sensitive to the leaf area and particularly to leaf conductance, surface resistance and moisture stress. These types of data are generally lacking for vegetation on forest watersheds.

Better data, both environmental and structural, should improve both resolution of predictions and general understanding. We encourage further integrative watershed ecosystem studies. These provide a complete set of data which we use both to improve our understanding of structure and function, and completely test and evaluate models which synthesize our knowledge and understanding.

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