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Watershed Hydrology

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Introduction

Watershed hydrology is determined by the local climate, land use, and pathways of water flow. At the Marcell Experimental Forest (MEF), streamflow is dominated by spring runoff events driven by snowmelt and spring rains common to the strongly continental climate of northern Minnesota. Snowmelt and rainfall in early spring saturate both mineral and organic soils and feed water to streams and the regional groundwater aquifer. However, large rains in midsummer or late fall can cause significant stormflow as well as groundwater recharge persisting into January. Defining and measuring water flow in each pathway reveal its importance to evapotranspiration, streamflow, groundwater recharge, and to the movement of nutrients, minerals, heavy metals, and organics (Chapter 8).

Significant changes in climate at the MEF in the last 12,000 years and the results of these millennial steps are discussed in Chapter 4. During the Wisconsin Glaciation period, annual precipitation was half (400 mm) that at glacial retreat (800 mm). Annual precipitation has since ranged from 700 to

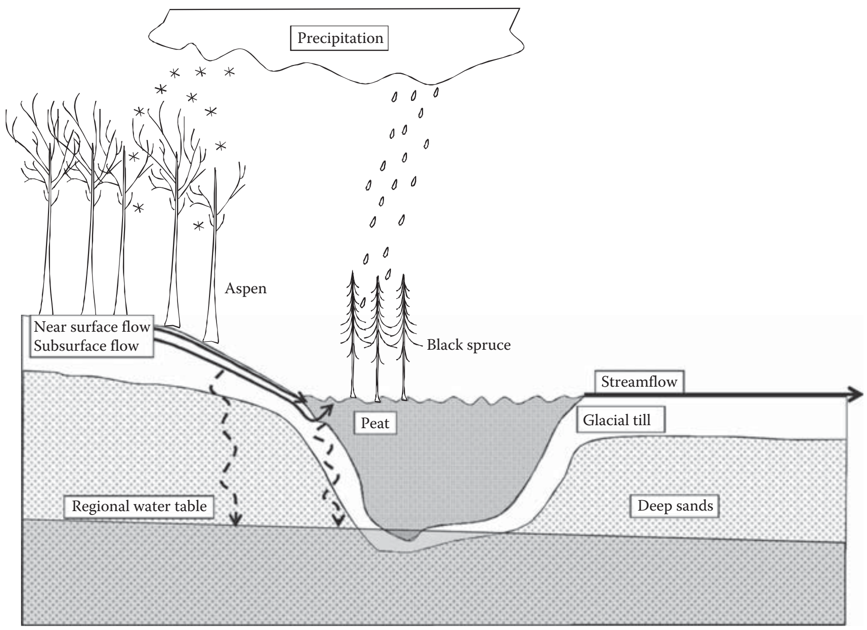

FIGURE 7.1
Net precipitation at perched watersheds in the MEF is rain or snow entering either the mineral or organic soil after interception through tree, shrub, and herbaceous layers. (Redrawn from Verry, E.S., Estimating water yield differences between hardwood and pine forests, Research Paper NC-128, USDA Forest Service, St. Paul, MN, 1976.)

800 mm and currently averages 782 mm. Mean temperature in July has varied 5°C since glacial retreat and now averages 18.9°C.

Once snow and rain reach the forests of MEF, interception, sublimation, throughfall, and stemflow are the first pathways to partition meteorological inputs between evaporation, storage, and runoff as mediated by forest type. Net precipitation (Figure 7.1; Verry 1976) is the rain or snowmelt that passes through the forest canopy and infiltrates through the litter to the forest floor and ultimately reaches the soil surface. Net precipitation is a key variable driving the amount of streamflow or groundwater recharge leaving the watershed.

Net Precipitation

In watersheds with deep sandy soils, net precipitation percolates directly to the regional groundwater system unless transpired or stored as soil water in the root zone. Fen peatlands in these watersheds receive net precipitation as a direct input to the regional water table, which usually flows from the fen as streamflow, but the direction of groundwater flow can reverse leading to groundwater recharge. In upland soils derived from glacial tills, net precipitation becomes (1) nearsurface flow occurring in the forest litter (generally because of underlying soil frost), (2) subsurface flow occurring in the saturated A and E horizons (often derived from eolian fine sands that overlie the tills), (3) deep seepage via unsaturated or saturated flow through the upland mineral soil into the regional water table, or (4) flow into bog peatlands at the wet edge (lagg). Flow from the peatland feeds streamflow in first-order channels.

Aspen (*Populus tremuloides*), red pine (*Pinus resinosa*), and black spruce (*Picea mariana*) are the major forest types at the MEF. The upland forests include an understory component dominated by the shrub hazel (*Corylus* sp.) and the herb bracken fern (*Pteridium* sp.). Bog peatlands have acid water (~pH 3.8) with depauperate flora characterized by low-*Ericaceae* shrubs and *Sphagnum* mosses. Tree species are primarily black spruce and tamarack (*Larix laricina*). Fen peatlands have near neutral pH water (~6.0–6.5) with a rich herbaceous flora, tall alder (*Alnus incana*) shrubs, and trees that include northern white cedar (*Thuja occidentalis*) along with black spruce and tamarack. Precipitation intercepted on the various vegetation layers is either evaporated or, on cold, sunny, and winter days, sublimated back to the atmosphere. Interception, stemflow, and throughfall were investigated at MEF with studies of hazel, aspen, and black spruce interception (Verry 1976) as well as literature data for red pine (Rogerson and Byrnes 1968).

In Lower Saxony, Germany, Delfs (1967) proposed a practical assessment of interception or net precipitation data: "Preliminary results suggest that it is chiefly interception which is responsible for the effect of beech and spruce stands on the water regime (streamflow) – the beech area discharged over 200 mm more than the spruce area." The hypothesis of Delfs was supported by findings of Swank and Miner (1968) who reported the effect on streamflow of converting a hardwood watershed to white pine (*Pinus strobus*) at the Coweeta Experimental Forest in North Carolina. Swank et al. (1972) concluded: "Increased interception loss occurs when hardwood-covered watersheds are converted to white pine, causing a significant reduction in streamflow (178 mm)."

At the MEF, the solution of 23 stemflow and throughfall equations for individual rainstorms and for annual snowfall, measured in the snowpack beneath each stand, yielded a nearly linear relationship between annual gross precipitation (rain plus snow) and annual net precipitation for each forest type and various basal areas (Figure 7.2).

The upland portion of S6 at MEF was converted from aspen to a mixture of red pine and white spruce (*Picea glauca*) in 1981. This provided a test of how well differences in net precipitation mirror differences in streamflow under hardwood and conifer forests. The S6 watershed contains 2.0 ha of black spruce bog and 6.9 ha of upland mineral soil. The mature aspen forest was harvested in the early winter and spring of 1980. Aspen regeneration was consumed by cattle grazing during the next three summers after which the upland was planted to a mixture of red pine and white spruce. When

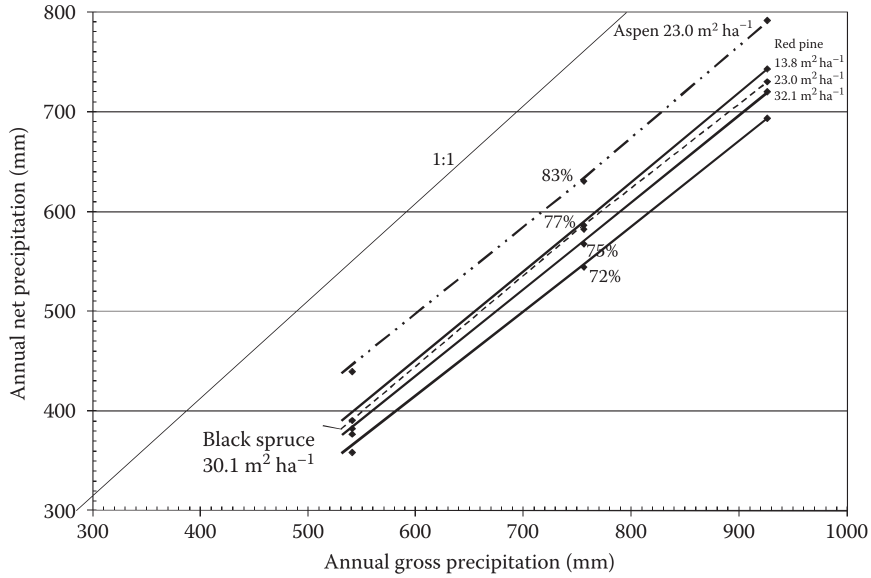

FIGURE 7.2
The relationship between net and gross annual precipitation for aspen (basal area of 23 m^2 ha^{-1}), red pine (three basal areas: 13.8, 23, and 32.1 m^2 ha^{-1}), and black spruce (basal area of 30.1 m^2 ha^{-1}) forests. Average precipitation at MEF is 783 mm. Annual net precipitation is 650 mm for aspen forests and 587 mm for red pine forests. (From Verry, E.S., Estimating water yield differences between hardwood and pine forests, Research Paper NC-128, USDA Forest Service, St. Paul, MN, 1976.)

conifers reached age 17, S6 produced 65 mm less streamflow than the watershed with mature aspen on the uplands (Chapter 13). Adjusted for a change in streamflow from just the upland area, there was a decrease in upland streamflow of 84 mm. The difference in net precipitation between mature aspen and red pine was 63 mm (Figure 7.2). The additional 18 mm reduction is attributed to transpiration from conifers, which retain their needles as opposed to aspen, which has leaves for about 150 days each year. Detailed net precipitation equations for aspen, spruce, and red pine are given in Verry (1976). These equations have been used to quantify the amounts of nutrients entering the forest floor as throughfall and stemflow (Timmons et al. 1977; Verry and Timmons 1977, 1982) and to partition the cycling of mercury and organic carbon through aspen and spruce forests (Kolka et al. 2001).

Water Flow in Mineral Soil

Frost affects the runoff of water from upland mineral soils to peatlands, and frost depth is inversely proportional to the snow depth in early winter. There is little frost when snow accumulates over 30 cm before air temperatures fall below −18°C. Net precipitation entering deep sands at the MEF satisfies soil-moisture storage and then percolates to the regional groundwater table via macropores and unstable wetting fronts. Water also may flow through the thawed upper sand horizons when soil frost at depth persists during snowmelt. The same is true of sandy loam and clay loam soils that comprise the glacial tills at the MEF. Vegetation cover type also affects frost occurrence due to canopy effects on snow depth. Annual variation in soil frost at the MEF is shown in Figure 7.3. In 1969, 1970, 1971, and 1996, little frost occurred when maximum snow depths exceeded 64 cm. The 1989 through 1995 sequence illustrates the inverse relationship between snow depth and frost depth.

Link between Uplands and Peatlands

The mineral soils of the glacial till upland and the organic soils of the peatland are linked at the lagg by water-flow pathways developed over millennia. Data collected on these pathways help us interpret how (1) upland water flows to the peatland, (2) upland and bog water mix in the lagg, (3) peatlands influence the generation of streamflow, and (4) the various components of the watershed contribute to deep, regional groundwater recharge via seepage.

Characteristics of the Upland Mineral-Soil Horizons in the Lagg

Water infiltrates the upland soil and flows downslope through the saturated A and E horizons. Figure 4.8 shows water flowpaths from the uplands to the peatland when the surface mineral soil horizons are saturated. In 1984, the soil horizons on the S2 upland were identified as the Warba Sandy Loam soil series in the Itasca County Soil Survey (Nyberg 1987). The A, E1, E2, E/B, and B/E horizons developed in an eolian loess cap of fine sandy loam that covers the glacial Koochiching clay loam till (Table 7.1). These horizons are sandy or fine sandy loam, medium to strongly acid in reaction, and friable with moderate to medium platy structure in the E horizons (Nyberg 1987; Tracy 1997). The eolian loess cap fell on the S2 uplands and some fine sand swirled into the open water of a tundra pool that covered the bottom of the ice block depression about 9300 cal years BP as suggested by pollen correlates of climate variables (Table 4.3). In subsequent millennia, peat filled the ice-block depression and eventually covered (paludified) the eolian deposits from an elevation of 420.2–422.0 m. The persistence of the A and E horizons under the peat allows a direct conduit for subsurface flow from the upland into the peatland lagg. When soil pits are excavated at the outside edge of the

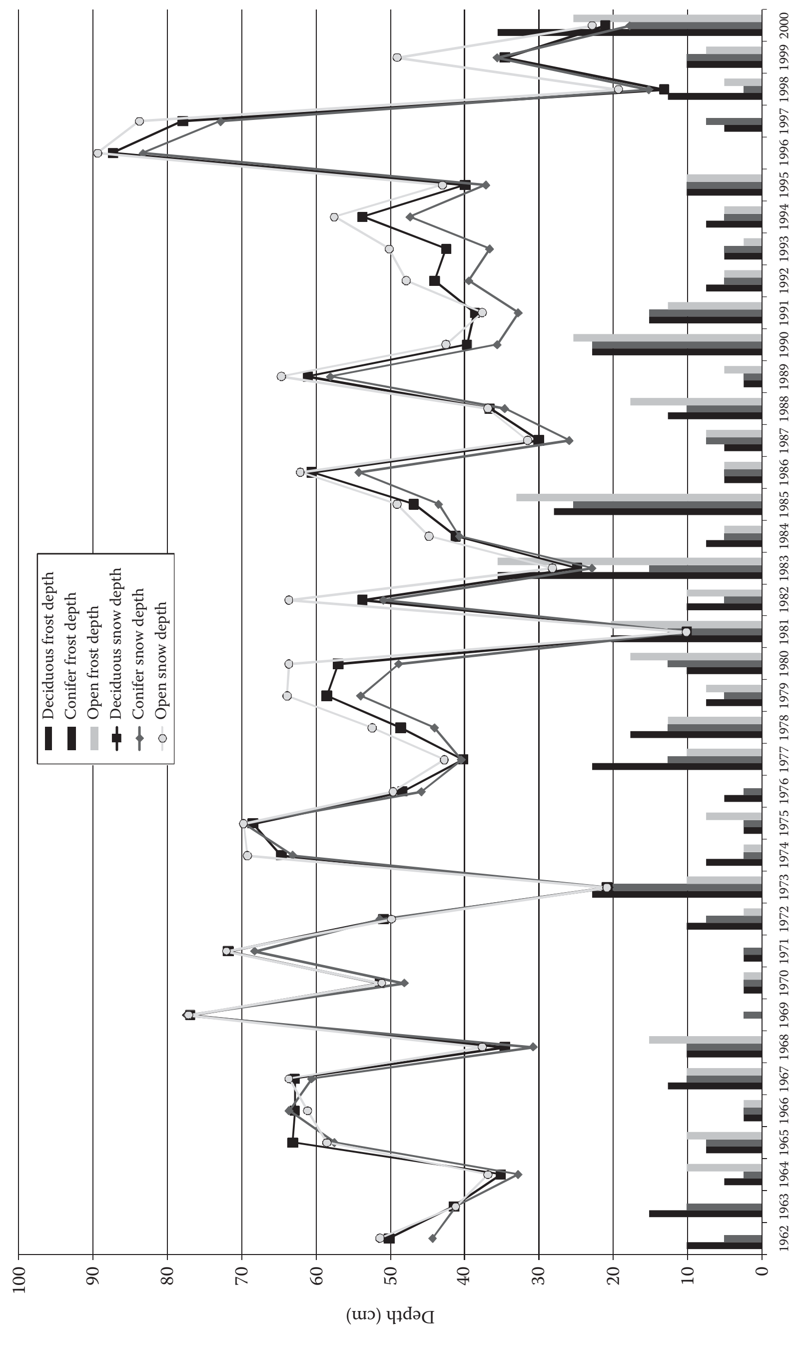

FIGURE 7.3
Annual variation in the depth of soil frost (bars) and maximum snow depth (lines) under deciduous (aspen), open (grasses), and conifer (black spruce on organic soils with some red pine on sands) forests.

TABLE 7.1

Description of the Warba Fine Sandy Loam Soil Series in Pit A at Eastern End of the S2 Bog

Horizon	Depth (cm)	Dry Color	Moist Color	Texture	Comment
0	1–0				Leaves and twigs
A	0–2	10YR 4/1	10YR 3/1	Sandy loam	Medium acid
E1	2–10	10YR 6/2, 7/2	10YR 4/2	Sandy loam	Strongly acid
E2	10–15	10YR 7/2	10YR 5/3	Sandy loam	Strongly acid
E/B	15–23	10YR 6/2		Fine sandy loam	Strongly acid
B/E	23–29			Fine sandy loam	Strongly acid
Bt1	29–54	10YR 4/3	10YR 4/4	Clay loam	Many distinct clay films (dry color)
Bt2	54–79	10YR 3/3, 3/2	10YR 4/3	Clay loam	10YR 3/2 clay films (dry color)
Bt3	79–98	Many faint 10YR 4/4 clay films	10YR 4/3	Loam	Common 10YR 4/4 and 7.5YR 3/2 root channel fillings; 10YR 8/2 filamentous carbonate masses
C	>98 (to sand)	No films	10YR 5/3	Loam	10YR 8/2 filamentous carbonate masses

Note: Information is for the mineral soil above and east of bog edge.

lagg, water flows freely from the interface of the peat and the underlying fine sandy loam (loess cap).

The Bt1 horizon has a moderate fine to moderate, angular, blocky structure that is firm, hard, and strongly acidic. Clay films coat the ped faces and sand sticks to the top of the clay boundary. The Bt2 horizon has a moderate coarse, prismatic structure with clay films, organic stains on the surface of peds, and a medium acidity. The Bt3 has a coarse subangular structure with few roots, organic stains, and a neutral pH. The C-horizon is massive, sandy clay loam, has few roots, and contains shale fragments in 1%–2% of its volume. These soils effervesce slightly with 10% HCl and are mildly alkaline (Nyberg 1987; Brooks and Kreft 1991; Tracy 1997).

Water moves through the upland soils into the deep sands beneath as evidenced by the filamentous carbonate masses following cracks in the Bt3 and C horizons. However, the amount of water loss is not sufficient to wash all the amorphous carbonate crystals from the soil. By contrast, all the mineral soil beneath the peatland lagg (Figure 4.8, horizontal meter scale 399.5–409.5 m) is free of carbonate masses (Brooks and Kreft 1991) and is acid in reaction. The eolian loess cap is up to 1.5 m beneath the peat surface (see 420 m contour in Figure 4.8).

Characteristics of Organic Soil Horizons at the Lagg

Tracy (1997) instrumented five sites around the perimeter of the S2 bog with piezometer nests. At each site, three nests were established on a transect from the upland mineral soil through the lagg and into the bog. Three piezometers were open to selected horizons within each type of profile except on of some mineral sites where only two piezometers were installed. Sand-point piezometers were also installed in the lagg areas to determine hydraulic pressure within the upper meter of the underlying sand. Piezometers screens were placed at elevations of 422.5, 421.5, and 419.6 m. In all cases, when bog water levels were normal to high (and no large amount of water came as subsurface flow from the upland), flow lines constructed between lines of equipotential water levels in the piezometers showed water at the outside edge of the lagg flowing downward and outward from the peatland. This water flows through the eolian loess horizon beneath the peatland lagg, through the Koochiching till that is stretched thin at the shoulders of the ice block depression, and finally through the mixed sandy clay loam layer (Figure 4.8) into the deep sands of the Rainy Lobe drift.

Water flows freely in the upper, less decomposed, organic soil horizons in S2. Values of field-measured hydraulic conductivity in the S2 bog are related to von Post degree of decomposition and range from 2.2×10^{-1} (von Post H1) to 1×10^{-6} (von Post H9) cm s^{-1} (Päivänen 1973; Gafni 1986; Gafni and Brooks 1990; Chapter 5).

Water arrives at the lagg from both the upland and the central dome of the bog (Figures 4.8, 7.4, and 7.5). The water-flow lines illustrated in Figure 7.4 occur when the upland mineral soils are below field capacity (unsaturated) and do not contribute water to the lagg. Figure 7.5 is a plan view of flow lines at the water table surface from the bog dome to the lagg. Flow through the lagg coalesces into a stream that flows from the southwest corner of the peatland.

Streamflow Generation at the S2 Bog

We separated the water leaving the bog as streamflow into a portion originating in the bog and a portion originating in the upland using a hydrograph-separation technique (Verry and Timmons 1982; Nichols and Verry 2001; Figure 7.6). Near surface and subsurface flow from the upland are combined as upland flow in Figure 7.6.

Figure 7.6 depicts the desynchronization of flow in a mixed basin (upland/peatland), an example of the variable source areas for streamflow. This variable source area concept was developed for streams by Hewlett (1961) to estimate the area of land directly contributing to streamflow as a saturation to the surface expands outward and upstream as precipitation increases. In upland/peatland mixed basins, the peatland water table is closer to the surface, making the peatland the primary source area for streamflow during late spring and summer. In a 4 cm rainstorm in July 1967, the S2 bog water

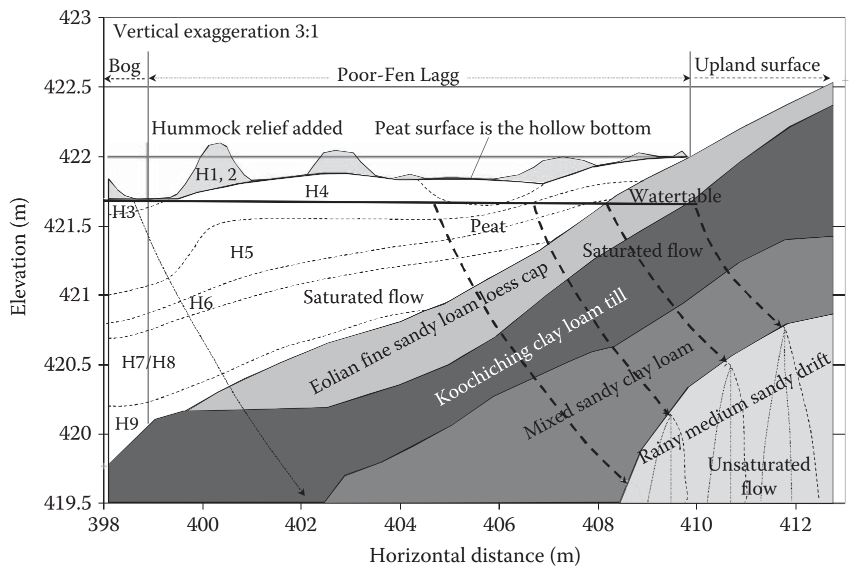

FIGURE 7.4
Organic (von Post H values) and mineral-soil layers beneath the lagg on the east end of S2. Water-flow lines represent a no-flow period from the upland; the water table in the bog slopes slightly downward from left to right.

table and the stream leaving the bog responded immediately to rainfall, but the peak flow from the upland was delayed 1–2 h depending on aspect (Verry and Kolka 2003). During snowmelt in early spring, the peatland and upland contributed equally to streamflow. However, streamflow from the peatland precedes and exceeds flow from the upland in this till-capped peatland basin (Figure 7.7).

Deep Seepage in the S2 Watershed

Some studies of water balance and nutrient yield on small watersheds have documented the tightness of the watershed and show or assume that virtually all liquid water leaves via streamflow (e.g., Likens and Bormann 1977). Despite this common assumption, some subsurface water flows to the regional water table. Penman (1963) stated that deep seepage "is frequently ignored altogether in catchment studies in the quiet hope that it is in fact, zero."

Nichols and Verry (2001) documented significant amounts of deep seepage from 10 to 50 ha watersheds on the MEF. The amount of deep seepage varies with the portion of glacial clay loam till and deep outwash sand in each watershed, the temporal distribution of precipitation, and air temperature.

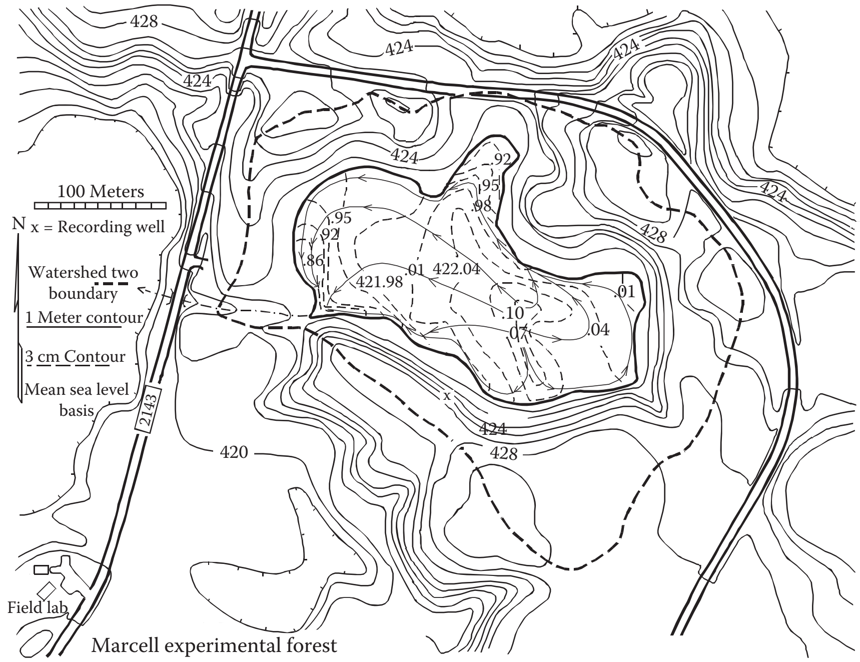

FIGURE 7.5
Much of the surface water leaving the S2 bog is from the dome. It flows from the dome to the poor-fen lagg (fine-lined arrows on the bog surface) where the water leaves the bog as a stream.

On average, deep seepage constitutes about 40% of the total water yield (streamflow plus seepage to groundwater). Adjacent watersheds (S4 and S5) on the North Unit of the MEF contain 15% deep outwash sands (Rainy Lobe drift) and 85% Koochiching till cap. Deep seepage ranges from 5 to 20 cm year^{-1} and averages 13 cm year^{-1}. The S2 watershed is completely covered with the Koochiching till cap; deep seepage ranges from 5 to 15 cm year^{-1} and averages 9 cm year^{-1}. Figure 7.8 shows the relationship using only annual precipitation (November–October) for 30 years on the S2 watershed. Nine centimeter of deep seepage across the entire surface area of the S2 watershed amounts to 8845 m^3 of water.

We estimated seepage to groundwater under watershed S2 using saturated hydraulic conductivity of the major material layers in the basin. We divided the surface area into upland, lagg, and bog and used average hydraulic-conductivity values for the soil material (K_{sat} values) to estimate seepage from each landscape unit. The critical impeding layers in the watershed include the Koochiching till covering the uplands, where the impeding layer is a series of clay-rich Bt horizons, the unweathered, and neutral-gray glacial flour (0.9 m thick; 10% very fine sand, 60% silt, and 30% clay) beneath both the bog and the lagg zone at the edge of the bog. Beneath the lagg above an elevation of 419.8 m, the stretched (and thus thinner) Koochiching till is mixed with

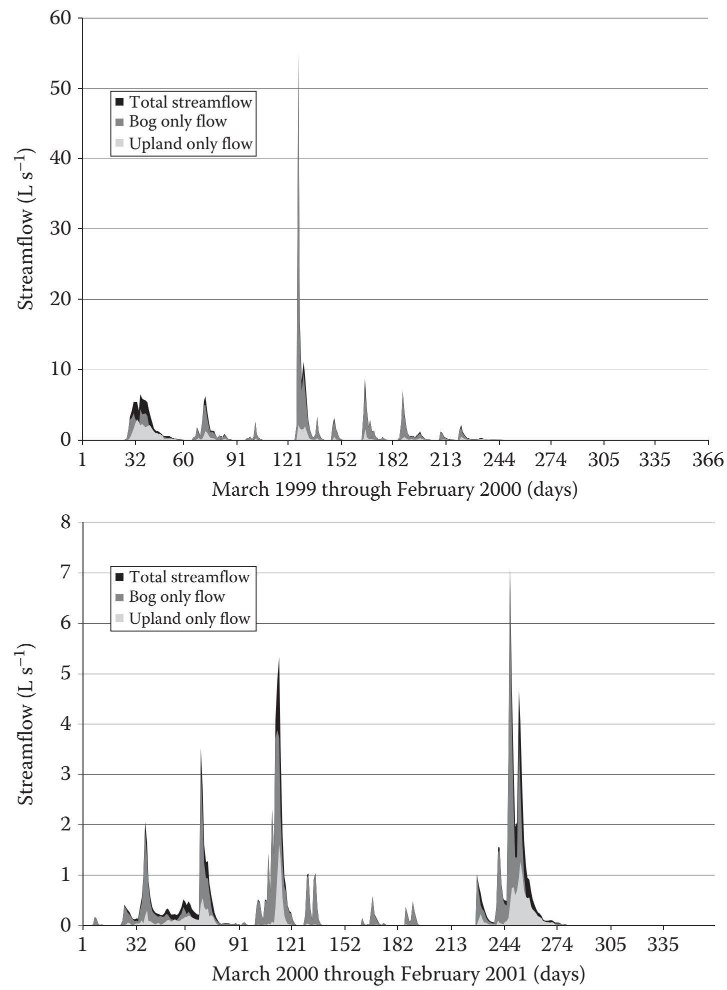

FIGURE 7.6
Annual hydrographs with upland and bog contributions separated from the total watershed streamflow. The year 1999 (top) illustrates a normal spring snowmelt peak and an extraordinary storm event on July 4, 1999. The year 2000 (bottom) illustrates an extraordinary wet period in the fall; note that the bog contribution precedes and exceeds the upland contribution.

sand on its bottom side (Figure 4.8). Table 7.2 shows K_{sat} values measured at the MEF by Tracy (1997) for mineral soils along with pertinent literature values for organic-soil K_{sat} values (see Tables 5.10 through 5.12). We used 16 years (1985–2000) of hydrograph separation data for S2 combined with rates of hydraulic conductivity and the area of the upland, lagg, and bog to calculate the amount of seepage for each watershed component (Figure 7.7).

In addition to the rate of seepage (K_{sat}), the time that free water is available to percolate must be considered. In the bog and lagg this is virtually

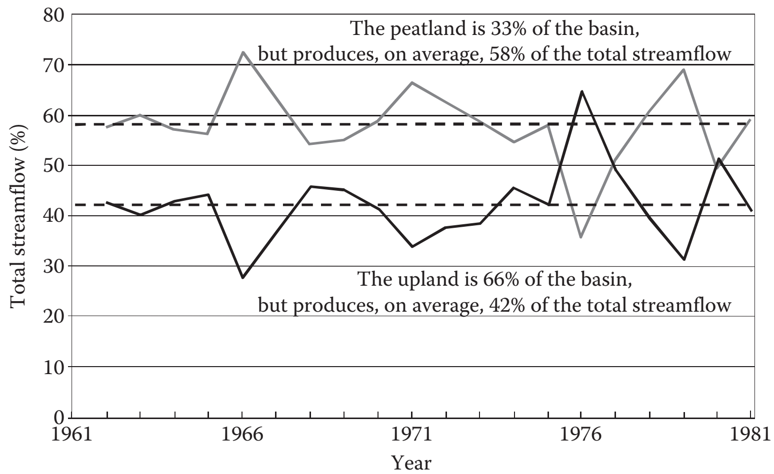

FIGURE 7.7
Relative contribution of the peatland and upland to streamflow over a 20 year period. Data for individual years were accumulated from daily data. The relative role of upland and peatland reverses in extreme drought years (e.g., 1976 and less so in 1980).

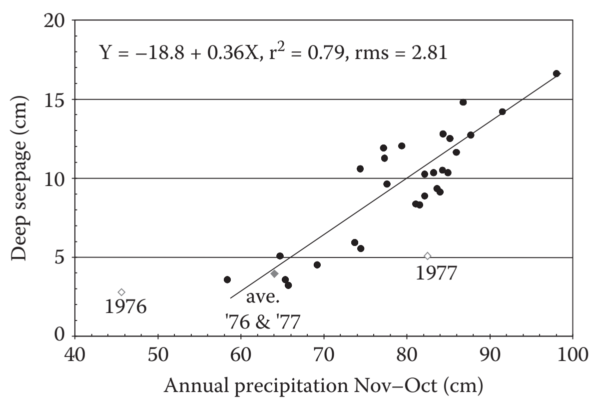

FIGURE 7.8
Deep seepage from S2 as a function of annual precipitation. The consecutive years of 1976 (dry with 45 cm) and 1977 (wet with 83 cm) were averaged due to pronounced hysteresis. (Modified from Nichols, D.S. and Verry, E.S., *J. Hydrol*., 245, 89, 2001.)

every day (365 days year^{-1}) except for years with severe droughts, such as 1976. In the upland, the number of days when the A and E horizons (and thus the Bt horizons) are saturated is highly variable. We estimated that deep seepage through the upland soil occurred on any day from April to June when the upland hydrograph exceeded 0.028 L s^{-1}, a threshold of saturation in the Bt horizon (Figure 7.6). In the months of high evapotranspiration that required substantial filling of the soil moisture deficit in the A, E, and B

TABLE 7.2

Hydraulic Conductivity (K_{sat}) Values for MEF Samples and from the Literature

Description	Bulk Density ($g\ cm^{-3}$)	K_{sat} ($cm\ s^{-1}$)	K_{sat} ($cm\ year^{-1}$)	Source
Mineral Bt, field	1.59	1.7×10^{-6}	54	Tracy, 1997, S2
Bt, field	1.59	1×10^{-6}	32	Tracy, 1997, S2
Bt, field	1.59	7×10^{-7}	22	Tracy, 1997, S2
Bt, lab	1.59	1.6×10^{-6}	50	Tracy, 1997, S2
Bt, lab	1.59	7×10^{-7}	22	Tracy, 1997, S2
Mixtures of sand, silt, clay	—	5×10^{-6}	160	Todd, 1959
Glacial till	—	5×10^{-7}	16	Todd, 1959
Stratified clays	—	2×10^{-7}	6	Todd, 1959
Unweathered clay, high	—	1×10^{-7}	3	Todd, 1959
Unweathered clay, mid	—	3×10^{-8}	1	Todd, 1959
Unweathered clay, low	—	1×10^{-9}	0.04	Todd, 1959

horizons (July through October), we estimated that deep seepage occurred on any day when the upland hydrograph exceeded 0.567 L s^{-1}. These rates are confirmed by the measurement of water in tanks that collect subsurface runoff from plots (Chapter 2). The number of days available for deep seepage through the upland soils ranged from 53 to 134.

We used the 30 year average water budget for S2 (Nichols and Verry 2001) and average or rounded K_{sat} values (Table 7.2) to estimate the relative amount of deep seepage from the upland, bog, and lagg (Table 7.3). A sum of watershed-component seepage rates to groundwater totaled 9916 m^3 $year^{-1}$. This is 12% more than that based on groundwater table recession curves. An agreement by two methods within 12% on S2 contrasts with errors associated with more general methods for estimating groundwater seepage that exceed 50% (Winter 1981). Table 7.3 also shows the relative importance of each watershed component to groundwater seepage.

The upland portion of the watershed accounts for slightly less seepage than its area (59 vs. 67%). The lagg accounts for 12 times the seepage its area would suggest (38 vs. 4%), and the bog accounts for only a tenth of what its area would suggest (3 vs. 29%). Plausible groundwater flow lines based on this analysis are shown in Figure 7.9.

The upper active layer of the peatland referred to as the acrotelm has a von Post degree of decomposition ranging from H1 to H8 and includes a low dome of *Sphagnum* moss peat (Figure 7.5). This layer has a separate, thin, subsurface water flow system that discharges water radially, downslope from the bog water table to the lagg and the stream (Figure 7.5), but also discharges water via shallow arcs to the peatland lagg (solid flow lines in the upper peat in Figure 7.9).

The deeper, hydrologically inactive zone is the catotelm that consists of highly decomposed von Post H9 peat. Such H9 *Sphagnum* peat transmits

TABLE 7.3

Deep Seepage for Each Component of the S2 Watershed

Watershed Component	Area (ha)	Area (%)	K_{sat} (cm s^{-1})	K_{sat} (cm $Year^{-1}$)	Saturated (Days)	Effective K_{sat} (cm $year^{-1}$)	Annual Seepage (m^3)	Annual Seepage (%)
Upland	6.48	67	1.14×10^{-6}	36	91	9	5832	59
Lagg	0.4	4	3×10^{-6}	95	365	95	3800	38
Bog	2.84	29	3×10^{-8}	1	365	1	284	3
Total watershed	9.72	100	—		—		9916[a]	100

[a] About 8845 m^3 based on water table recession and an aquifer specific yield of 0.2 (Nichols and Very 2001).

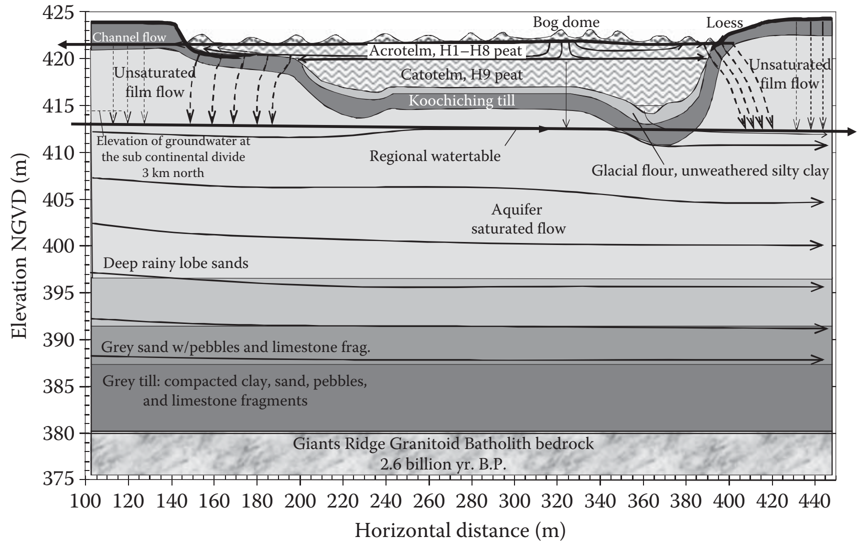

FIGURE 7.9
Plausible water-flow lines (solid lines show saturated flow; dashed lines show unsaturated flow) illustrate the relationship between the bog water table and the regional groundwater table when there is no upland subsurface flow. When upland subsurface flow occurs, unsaturated flow from uplands through the unsaturated zone is similar to the unsaturated flow that originates from the lagg. The dome peaks along the distance scale at 320 m.

21 cm water year^{-1}. However, a layer of unweathered, smooth, silty clay beneath the peat transmits less than 1 cm water year^{-1}. Although it is effectively an aquiclude, this layer does connect the saturated deep peat of the bog with the saturated regional aquifer.

Concrete Frost and Water Movement of Colloidal Organic Matter

Another mechanism of water and organic-carbon flow in organic soils is related to concrete frost in both bog and fen organic soils, although the process is more important in bog soils. When concrete frost forms in organic soils, colloidal organic matter is excluded preferentially from the ice crystal lattice and moves progressively deeper to the bottom of the concrete frost. The upper part of concrete frost is clear and the bottom layer is dark brown. The process of freezing moves colloidal organic matter and associated elements on nearly an annual basis from the acrotelm (where organic matter regularly contributes to dissolved organic carbon and nutrients leached in streamflow) to the catotelm (where they can be stored for millennia).

Streamflow from Upland–Peatland Watersheds

The high-pH water in fens is derived from the regional groundwater aquifer. It dissolves calcium and magnesium from gravel-size pieces of limestone in the deepest glacial ground moraine (gray till for the Itasca Lobe that originates from the Hudson Bay Dome Ice Center). The dissolved limestone imparts the near neutral pH and signals the inflow of large amounts of groundwater discharge into the fen peatland. The duration of surface streamflow from two watersheds at the MEF (S3 and S5) illustrates the effect of groundwater discharge on the duration and amount of streamflow (Figure 7.10). S3 and S5 have the same surface area, but the groundwatershed for S3 is 10 times the size of its surface watershed (Sander 1976). Because water from the regional groundwater aquifer discharges into the S3 fen, streamflow is nearly constant 70% of the time and an order of magnitude higher than streamflow from S5. Streamflow at S3 is perennial. Streamflow from the S5 watershed has a steep flow-duration curve with no flow about 25% of the time during

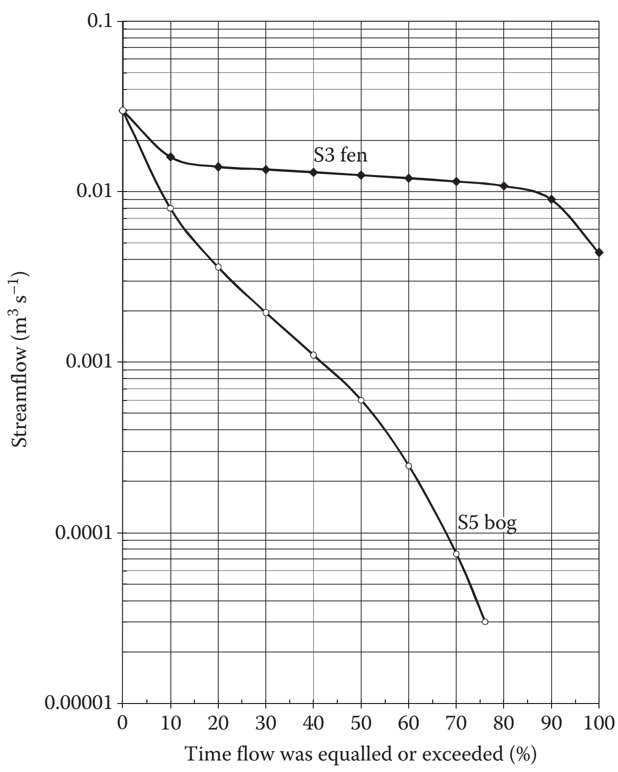

FIGURE 7.10
Streamflow-duration curves for the S3 watershed (groundwater-fed fen, solid circles) and watershed S5 (bog peatland, open circles). Both watersheds have the same surface area.

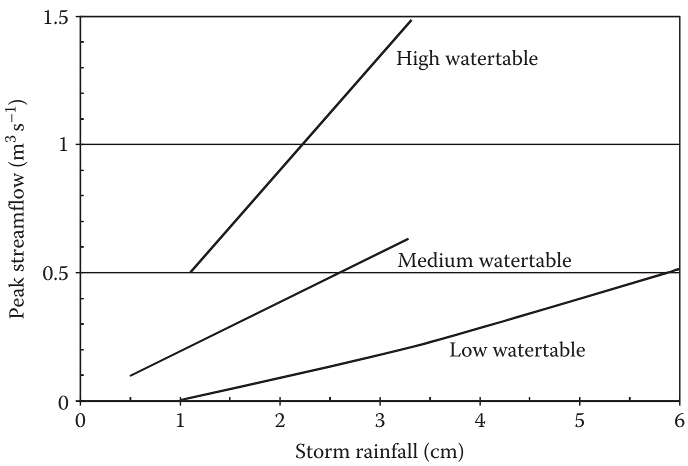

FIGURE 7.11
Water table position in the peatland at the time of a rainstorm affects the magnitude of the streamflow peak flow for the same size of rainstorm. Streamflow from peatlands with high water table is most responsive rainfall. (From Bay, R.R., *J. Hydrol*., 9, 90, 1969.)

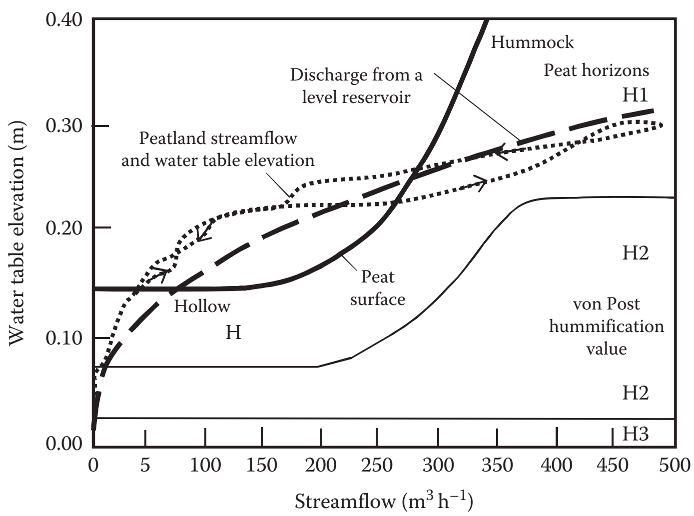

FIGURE 7.12
Peatland streamflow and measured water table elevation (returning dashed line) are imposed on a cross section of the hollow/hummock microtopography showing the porous von Post H1–H3 peat horizons. The direction of the arrow shows the progression of the hydrograph with hysteresis loop at the high flows. The measured stormflow response in the peatland is nearly identical with the rating curve for a level reservoir (heavy dashed line).

summer months and winter. Note that the maximum streamflow rate for both watersheds is the same (flow-duration curves converge in upper left in Figure 7.10), because annual streamflow peaks are driven largely by spring snowmelt occurring in the surface watershed.

Stormflow from rain in a bog or fen watershed is dependent on the position of the water table in the peatland at the time of the rain. Bay (1969) reported this phenomenon (Figure 7.11) and is supported by the hydrographs in Figure 7.6.

The largest peak streamflow (486 m^3 h^{-1}) from rain (17 cm) was caused by a large storm cell over a 35 h period on July 2 and 3, 1979. Water table response in the hummock and hollow topography of the S2 bog illustrates a hysteresis effect on the rising and falling leg of the hydrograph and its near match to the rating curve for a level reservoir (Figure 7.12). This similarity of reservoir and peatland stormflow responses is not surprising given that both peatlands and reservoirs have flat surfaces. The loop in the peatland response curve also occurs in reservoirs and is caused by the movement of a wedge of water through the reservoir (Verry et al. 1988). This proof of concept has been used to model rainstorm response in peatlands (Chapter 15).

Summary

- Intermittent streamflow from bog watersheds occurs when rainfall or snowmelt is rapidly routed at times of high water tables in bogs and laggs. Most of the annual stream water yield occurs in response to spring snowmelt. Although peakflows may be similar between fen and bog watersheds, the streamflow from the fen varies over about one order of magnitude in contrast to bog streamflow that varies up to four orders of magnitude.
- Mineral soils on uplands and organic soils in peatlands have distinct effects on streamflow from peatland watersheds. Studies at the MEF have quantified the relative magnitude of water movement along various flowpaths for watersheds with large central bogs and partitioned hydrological outflows to show that evapotranspiration (52 cm $year^{-1}$) on average accounts for the largest output of water followed by streamflow (17 cm $year^{-1}$) and deep seepage (9 cm $year^{-1}$).
- The magnitude and timing of streamflow from peatland watersheds at the MEF is a function of the hydrogeological setting. The surface elevation of perched bogs is at least several meters above the regional groundwater table. Saturated organic soils maintain a permanent hydraulic gradient that results in slow, but constant deep seepage of water to the regional aquifer from bogs and laggs. Streamflow from fens with outlet streams is augmented by the inflow of groundwater

from the regional aquifer in addition to precipitation inputs on upland soils. In contrast to bog watersheds that lose significant amounts in the annual water budget to deep seepage, groundwater inflow to fens may cause annual stream-water yields that are orders of magnitude larger than water yields from bog watersheds.

- Although laggs around bogs in perched watersheds occupy a small fractional area of the landscape, laggs are particularly important to the hydrology of a watershed. Laggs receive rainfall and snowmelt melt inputs from direct inputs as well as bog and upland soils. Flow through the lagg routes these waters to outlet streams when water tables are high. In addition, most of the deep seepage for the entire watershed occurs through soils that lie beneath laggs.

References

Bay, R.R. 1969. Runoff from small peatland watersheds. *Journal of Hydrology* 9(1):90–102.

Brooks, K.N. and D.R. Kreft. 1991. Hydrologic linkages between uplands and peatlands. Final Report to Water Resources Research Center, University of Minnesota, College of Natural Resources, St. Paul, MN.

Delfs, J. 1967. Interception and stemflow in stands of Norway spruce and beech in West Germany. In *International Symposium of Forest Hydrology*. Penn. State Univ. New York: Pergamon Press, pp. 179–185.

Gafni, A. 1986. Field tracing approach to determine flow velocity and hydraulic conductivity of saturated peat soils. PhD dissertation. St. Paul, MN: University of Minnesota.

Gafni, A. and K.N. Brooks. 1990. Hydraulic characteristics of four peatlands in Minnesota. *Canadian Journal of Soil Science* 70:239–253.

Hewlett, J.D. 1961. Soil moisture as a source of base flow from steep mountain watersheds. Research Paper 132. Asheville, NC: USDA Forest Service. Southeastern Forest Experiment Station.

Kolka, R.K., D.F. Grigal, E.A. Nater, and E.S. Verry. 2001. Hydrologic cycling of mercury and organic carbon in a forested upland-bog watershed. *Soil Science Society of America Journal* 65(3):897–905.

Likens, G.E. and H.P. Bormann. 1977. *Biogeochemistry of a Forested Ecosystem*. New York: Springer Verlag.

Nichols, D.S. and E.S. Verry. 2001. Stream flow and ground water recharge from small forested watersheds in north central Minnesota. *Journal of Hydrology* 245(1–4):89–103.

Nyberg, P.R. 1987. *Soil Survey of Itasca County, Minnesota*. St. Paul, MN: USDA Soil Conservation Service.

Päivänen, J. 1973. Hydraulic conductivity and water retention in peat soils. *Acta Forestalia Fennica* 129:1–70.

Pennam, H.L. 1963. Vegetation and hydrology. Technical Communication No. 53. Harpenden, UK: Commonwealth Bureau of Soils.

Rogerson, T.L. and W.R. Byrnes. 1968. Net rainfall under hardwoods and red pine in central Pennsylvania. *Water Resources Research* 4(1):55–57.

Sander, J.E. 1976. An electric analog approach to bog hydrology. *Ground Water* 14(1):30–35.

Swank, W.T. and N.H. Miner. 1968. Conversion of hardwood-covered watersheds to white pine reduces water yield. *Water Resources Research* 4:947–954.

Swank, W.T., N.B. Goebel, and J.D. Helvey. 1972. Interception loss in loblolly pine stands of the South Carolina Piedmont. *Journal of Soil and Water Conservation* 27(4):160–1264.

Timmons, D.R., E.S. Verry, R.E. Burwell, and R.F. Holt. 1977. Nutrient transport in surface runoff and interflow from an aspen-birch forest. *Journal of Environmental Quality* 6(2):188–192.

Todd, D.K. 1959. *Groundwater Hydrology*. New York: Wiley.

Tracy, D.R. 1997. Hydrologic linkages between uplands and peatlands. PhD dissertation. St. Paul, MN: University of Minnesota.

Winter, T.C. 1981. Uncertainties in estimating the hydrologic balance of lakes. *Water Resources Bulletin* 17(1):82–115.

Verry, E.S. 1976. Estimating water yield differences between hardwood and pine forests. Research Paper NC-128. St. Paul, MN: USDA Forest Service.

Verry, E.S., K.N. Brooks, and P.K. Barten. 1988. Streamflow response from an ombrotrophic mire. In *Proceedings, International Symposium on the Hydrology of Wetlands in Temperate and Cold Regions*. Helsinki, Finland: International Peat Society/The Academy of Finland, pp. 52–59.

Verry, E.S. and R.K. Kolka. 2003. Importance of wetlands to streamflow generation. In *Proceedings, First Interagency Conference on Research in the Watersheds*, ed. K.G. Renard, S.A. McElroy, W.J. Gburek, H.R. Canfield, and R.L. Scott. Benson, AZ, October 27–30, 2003. Tucson, AZ: USDA Agricultural Research Service, pp. 126–132.

Verry, E.S. and D.R. Timmons. 1977. Precipitation nutrients in the open and under two forests in Minnesota. *Canadian Journal of Forest Research* 7(1):112–119.

Verry, E.S. and D.R. Timmons. 1982. Waterborne nutrient flow through an upland-peatland watershed in Minnesota. *Ecology* 63(5):1456–1467.