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Stream flow and ground water recharge from small forested watersheds in north central Minnesota

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Abstract

In hydrologic studies of forested watersheds, the component of the water balance most likely to be poorly defined or neglected is deep seepage. In the complex glaciated terrain of the northern Lake States, subsurface water movement can be substantial. On the Marcell experimental forest (MEF) in north-central Minnesota, ground water table elevations measured in observation wells in recharge areas were used to calculate rates of ground water recharge. In northern Minnesota winters, precipitation is stored on the surface as snow and ground water recharge ceases. Water table elevations in recharge areas decline over winter at calculable rates. Deviations from these rates during other times of the year are due to ground water recharge. On 10–50 ha watersheds on the MEF, ground water recharge varies among watersheds but constitutes about 40% of the total water yield. Annual ground water recharge amounts were found to vary linearly with precipitation. Even in high precipitation years, the infiltration capacity of the watersheds was not exceeded. Regression equations were developed relating yearly ground water recharge, stream flow, and total water yield, to seasonal precipitation amounts, summer and autumn precipitation during the previous year, and non-winter air temperature. Published by Elsevier Science B.V.

Keywords: Ground water; Recharge; Water yield; Runoff

1. Introduction

Beginning with the work of Engler (1919) in Switzerland and Bates and Henry (1928) at Wagon Wheel Gap in Colorado, dozens of studies have described the water balance of forested watersheds and assessed the effects on water yield of various timber management activities. See, for example, reviews by Hibbert (1967), Bosch and Hewlett (1982), and Hornbeck et al. (1993). In such studies, the component of the hydrologic balance most likely to be neglected or poorly defined is the subsurface movement of water from the catchment by deep seepage. Penman (1963) stated that deep seepage "is frequently ignored altogether in catchment studies in the quiet hope that it is, in fact, zero".

In some catchments with thin soils over bedrock, this hope is probably justified. However, in the complex glaciated terrain of the northern Lake States (Minnesota, Wisconsin, and Michigan), bedrock may be overlain by many layers of till and outwash that may be as much as 120 m thick overall. In this setting, subsurface water movement can be substantial. Few data exist, however, from which to calculate its magnitude or significance. This study was undertaken

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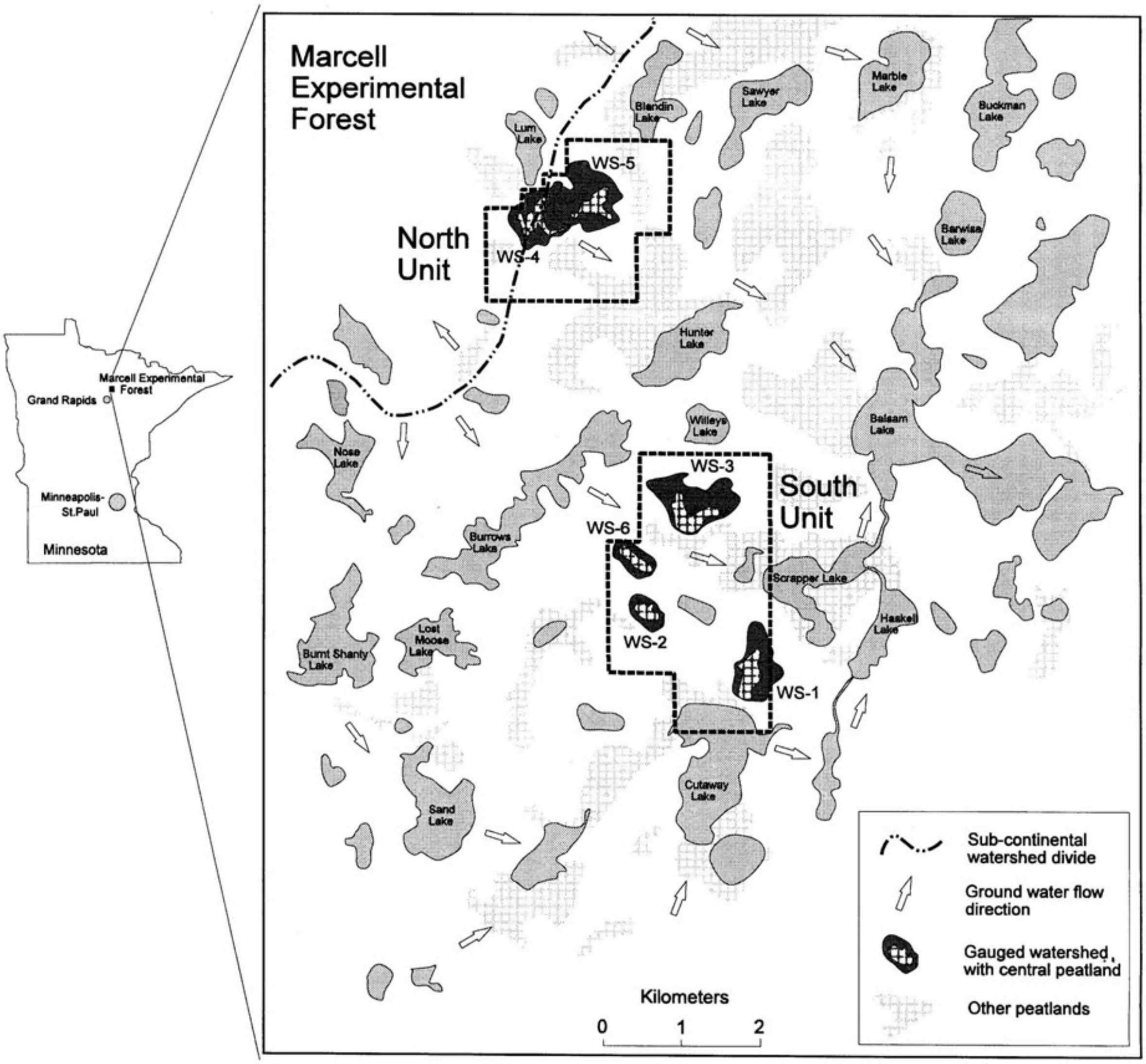

Fig. 1. Map of Marcell experimental forest.

to determine the amounts and timing of deep seepage to ground water from three small forested watersheds in north-central Minnesota.

2. Methods

2.1. The study area

The study was conducted on the Marcell experimental forest (MEF) in north-central Minnesota (ca. 47° 32′N, 93° 28′W), about 40 km north of Grand Rapids (Fig. 1). The glaciated terrain is one of rolling hills dotted with numerous small lakes and peatlands. The uplands are covered by forests of aspen–birch (*Populus–Betula*) with smaller amounts of northern hardwoods and balsam fir (*Abies balsamea* (L.) Mill). The peatlands are forested with black spruce (*Picea mariana* (Mill) Britton et al.) and tamarack (*Larix laricina*(Du Roi) K. Koch), with some open areas where high water levels prevent tree growth. This landscape is typical of large areas in the northern

Lakes States and in Canada. The area has a marked continental climate, with short, warm summers; long, cold winters; and wide temperature variation. The average annual temperature is about 3.2°C, with yearly highs in the 30–35°C range and lows from −35 to −40°C. Mean annual precipitation is about 78 cm, of which about 20–25% falls as snow.

The data presented in this paper are from gauged watersheds on the MEF: watersheds 4 and 5 (ws-4 and ws-5) on the north unit of the MEF and watershed 2 (ws-2) on the south unit (Fig. 1). All three watersheds are more or less circular. Each contains a centrally located peatland surrounded by mineral soil uplands. Each watershed is drained by an intermittent stream that emanates from the central peatland. Ws-2 and ws-5 drain southward to the Mississippi River system. Ws-4 sits astride the subcontinental divide and is drained by two streams, one flowing north via the Rainy River system to Hudson Bay and the other south to the Mississippi. Ws-2 is about 9.7 ha in area, of which about 3.3 ha is peatland. Ws-4 is about 34 ha including a 8.1-ha peatland. Ws-5 is about 52.6 ha in size, with a central peatland of about 6.1 ha. Ws-5 contains two additional peatlands with a total area of 6–8 ha in semi-closed basins that fill up and overflow into the central peatland during wet periods.

The predominant upland soils on the watersheds are Warba fine sandy loam on ws-2 and the similar Nashwauk fine sandy loam on ws-4 and ws-5. These are deep, well drained, slowly to moderately slowly permeable soils, formed in till. They consist of a 25–35 cm layer of fine sandy loam over heavier textured clay loam, sandy clay loam, silt loam, and loam. Ws-4 and ws-5 also contain about 10–15% excessively well drained, rapidly permeable Menahga coarse sand soils. The clayey till layer in which the Warba and Nashwauk soils are formed, ranges from about 2–5 m thick. It is underlain by 10–40 m of sand.

The uplands of all three watersheds are forested. On ws-2 and ws-5, the forest cover is primarily mature (80 + years) aspen (*Populus tremuloides* Michx., *Populus grandidentata* Michx.) with lesser amounts of paper birch (*Betula papyrifera* Marshall), and balsam fir. Some of the sandier areas in ws-5 also support scattered red pine (*Pinus resinosa* Ait.). The upland forest of ws-4 is similar in composition, but is younger, having been harvested in 1970 and 1971.

The peatlands in each watershed contain deep deposits of well-decomposed organic material overlain by 30–100 cm of poorly to moderately decomposed material. All are acidic, ombrotrophic (nutrient-poor) bogs. Although they are ponded or saturated most of the year, these bogs are perched several meters above the regional ground water table and receive water only from direct precipitation and runoff from the surrounding uplands. Their acidic nature (pH ≈ 4.0) reflects their separation from the carbonate-rich ground water. The bog vegetation consists of an overstory of 130-year-old black spruce, some 12–15 m in height; an understory of low, ericaceous shrubs — Labrador tea (*Ledum groenlandicum* Oeder) and leatherleaf (*Chamaedaphne calyculata* (L.) Moench); and sphagnum moss (*Sphagnum angustifolium*, *Sphagnum magellanicum*). The ws-4 bog contains an area of open water in the center.

2.2. Hydrologic measurements

Measurements of various components of the hydrologic balance for these watersheds began in the early to late 1960s. Surface flow from each watershed is recorded continuously at 120-degree sharp-crested weirs installed on each stream. Each weir is tied to a concrete cut-off wall that extends several feet into the clayey soil of the stream banks and stream bottom to minimize leakage around or below the weir. We assume that none of the water seeping to the regional water table is measured at the weir. Precipitation is collected in US National Weather Service standard 20.3-cm rain gauges located at from two to four sites in each watershed. Precipitation on each watershed is calculated as a weighted average of the gauge readings in that watershed (Thiessen, 1911). Water levels in each bog are measured in 10-cm-diameter PVC wells equipped with continuous level recorders. Available soil moisture is measured three times a year at two sites in each watershed using a neutron meter inserted into 3.8-cm-diameter aluminum access tubes. The elevation of the regional ground water table below each watershed is measured monthly in 3.2-cm-diameter galvanized steel observation wells by use of a steel tape. The end of the tape is coated with carpenter's chalk and lowered into each

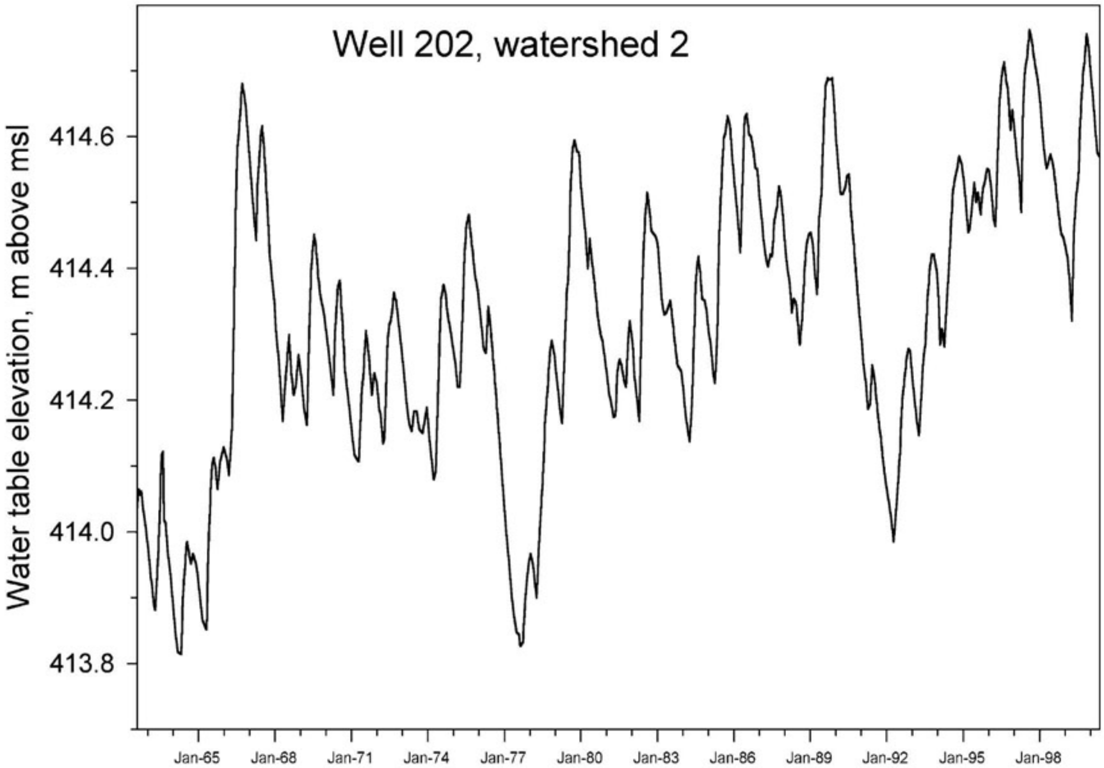

Fig. 2. Year-to-year variation in water table elevation below watershed 2 (water table elevation is measured monthly to ±0.01 ft with a chalked steel tape).

well to just below the water level. The distance from the top of the well to the water mark on the chalk is then subtracted from the elevation of the well top. The elevation of the top of each well relative to msl is surveyed annually to confirm well stability.

3. Results and discussion

3.1. Ground water recharge

The MEF lies literally at the 'top of the watershed' The subcontinental divide between water flowing to Hudson Bay and water flowing to the Gulf of Mexico runs through ws-4 and just a few hundred meters to the northwest of ws-5. Ws-2 is about 3.5 km southeast of the divide. Ground water flow paths through the MEF area tend to follow surface topography. The surface water divide is the ground water divide as well. On the Mississippi River side of the divide, ground water flows generally to the southeast. Ground water discharges into Hunter, Cutaway, Scrapper, Haskell, and Balsam Lakes, which feed into the Prairie River, a tributary to the Mississippi (Fig. 1) (Hawkinson and Verry, 1975).

Changes in ground water elevation reflect local recharge and discharge. Fig. 2 shows the hydrograph of observation well 202 in ws-2. Hydrographs of the other wells at the MEF are similar. During the 38-year period of record, the water table at this location has varied over an elevation range of about 0.9 m. Depending on surface topography, the water table generally lies 5–20 m below the land surface. In most years, snow melt and spring rains result in significant surface water flows in April and May. Seepage water from these events reaches the water table one to three months later, causing a rise in the water table surface that peaks in June or July. From July through October, the water table may rise or fall depending on whether growing-season (May–August) precipitation is greater than or less than evapotranspiration (ET) losses. Heavy autumn rains may result in an additional rise in ground water levels in November and December. During late fall and winter, precipitation accumulates as snow. Surface flow from the watersheds dwindles to zero as stream channels and peatlands freeze. Any soil water in excess of field capacity drains away, and ground water recharge stops. From January through March in most years, in the absence of recharge, the water table declines steadily as ground water flows down-gradient and discharges into lakes and steams. In some years, due to late season rains or later than normal freeze-up, some ground water recharge continues into the winter.

For the three ground water observation wells representing ws-2, ws-4, and ws-5 (wells 202, 401, and 501) and for two additional wells on watersheds 6 and 3 (Fig. 1) (wells 601 and 304), we calculated the rate of water table decline during each winter. At all wells, in the absence of recharge, the rate at which the water table declined during a particular winter depended on the elevation of the water table that winter (Fig. 3). When water tables were higher, rates of winter decline were higher, presumably because ground water gradients to discharge points were steeper and rates of discharge were greater. On the north unit of the MEF, at wells 401 and 501, winter water table decline ranged from about 0.10 cm day^{-1} when ground water levels were low, to 0.18 cm day^{-1} during winters with high water tables. At south unit wells, 202, 601, and 304, the rates were slightly lower at 0.08–0.16 cm day^{-1}. For all five wells, the equations for the regression of rate of water table decline on water table elevation had similar slopes and had r-squared values of 0.72–0.84.

We estimated recharge to ground water in the vicinity of each well by using that well's regression equation (Fig. 3) to calculate the distance the water table would have fallen, in the absence of recharge, between elevation measurements. During most winters and in some extended dry periods at other times of the year, the actual change in water table elevation between measurements equaled that predicted for conditions of no recharge. For other times, differences between the calculated no-recharge elevation and the actual elevation were assumed to be due to recharge (Fig. 4). The sum of these differences over a year multiplied by an estimated specific yield of 0.2 for the sand aquifer (Heath, 1983; Myette, 1986) equals the total recharge for the year in centimeters of water. This technique is not valid where water table elevations are influenced by significant ground water movement from up slope. The peatlands in ws-3 and ws-1 on the MEF (Fig. 1) are ground water discharge areas. Water table elevations measured in wells near these peatlands did not exhibit predicable winter declines.

Fig. 5 shows monthly amounts of stream flow and seepage to ground water from ws-2 and from the contiguous areas of ws-4 and ws-5 (ws-4/5) (average of ws-2 and ws-4/5, 1981–1999). Stream flow and seepage to ground water are minimal during December through March. By far, the greatest monthly amount of stream flow occurs in April, as a result of snow melt and early spring rains. On average, stream flow is concentrated in the spring and early summer, with 30% of the annual stream flow coming in April, almost 50% coming in April and May, and 72% occurring in April through July. The amount of water reaching the water table increases in April and peaks in May. Because of the varied seepage pathways and travel times, and the ponding of water in wetlands, seepage to ground water is distributed somewhat more evenly through the spring, summer, and fall than is stream flow. Sixty percent of the seepage water reaches the water table in April–July and 32% in August–November. However, we must emphasize that the monthly values shown in Fig. 5 are averages. For both stream flow and seepage to ground water, maximum monthly amounts generally exceed average values by two to four times. Zero monthly stream flow has been recorded for every month except April and May, and zero seepage has occurred in every month but April, May, and June.

3.2. Water balance

Annual water balances for ws-2 and ws-4/5 are shown in Table 1. Data are presented for 1967–1972 and 1978–1999 for ws-4/5 (groundwater data were not available for ws-4/5 between 1973 and 1977), and for 1970–1999 for ws-2. Shown are annual amounts of precipitation (P), stream flow (Q), ground water recharge (G), change in soil storage (∇S_s), change in peatland storage (∇S_p), and ET. ET was calculated as: $ET = P - Q - G - \nabla S_s - \nabla S_p$. For

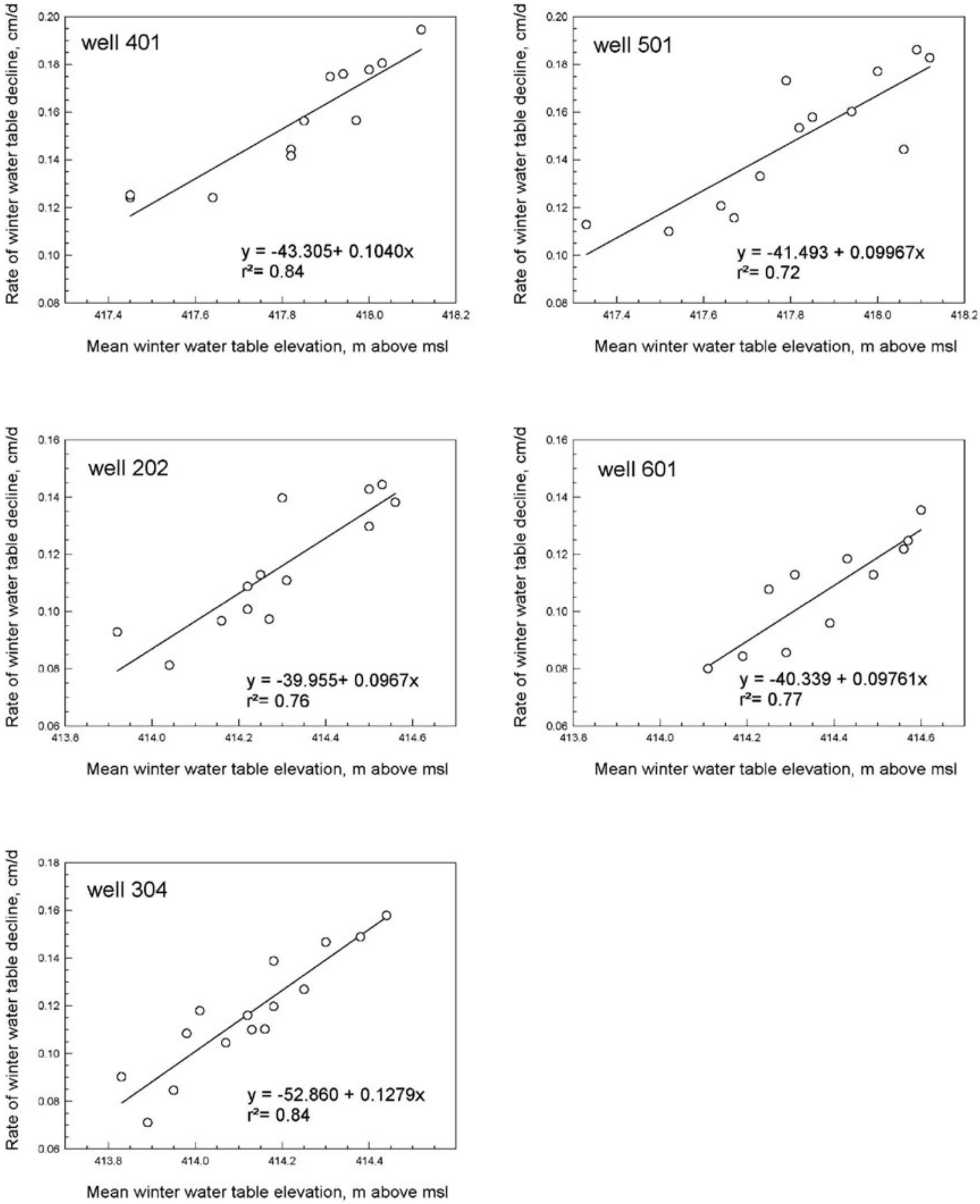

Fig. 3. Winter decline of water table elevations in five upland wells.

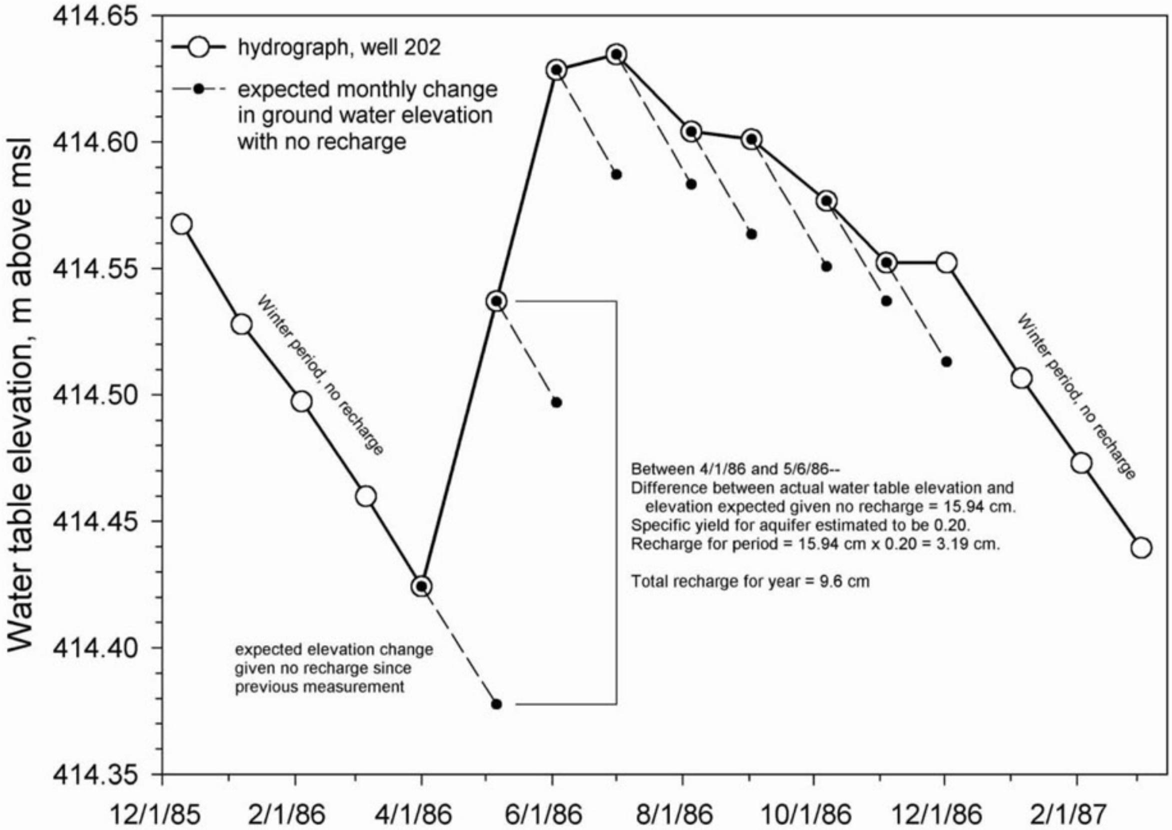

Fig. 4. Method of calculating monthly ground water recharge.

comparison, Thornthwaite potential evapotranspiration (PET), the maximum amount of moisture that would be lost to the atmosphere via ET under optimal soil moisture conditions (Thornthwaite and Mather, 1955) is also shown. Thornthwaite PET, based on air temperature and day length, does not account for evaporation when the mean monthly air temperature is 0°C or below, which at the MEF is November through March. During that 5-month period, the MEF landscape is usually snow-covered, and some water is lost to the atmosphere by sublimation. The ET term in Table 1, calculated by difference, includes sublimation. Based on measurements of winter precipitation and maximum snowpack water content on the MEF (Verry, 1976), we estimate that annual water loss by sublimation averages about 1.8 cm.

On the MEF, precipitation generally occurs as snow during November through March and is thus stored until it melts in the spring. Annual precipitation was calculated on a November through October basis. As seen in Fig. 5, both stream flow and ground water recharge are generally negligible during the winter and begin in earnest after snow melt gets under way. The small amounts of runoff or recharge that may occur in January or February are most likely attributable to the previous year's precipitation. For water budget purposes, annual runoff and recharge amounts were calculated from March through February. In most years, soil water content is high following snow melt, is decreased during the growing season by ET, and then is recharged to some extent by autumn rains. Changes in soil water storage were calculated as the year-to-year differences in available water in the soil to a depth of 150 cm measured in November. The peatlands in ws-2 and ws-4/5 have their own local water tables that are perched above the regional water table. Changes in peatland storage were based on November measurements of peatland

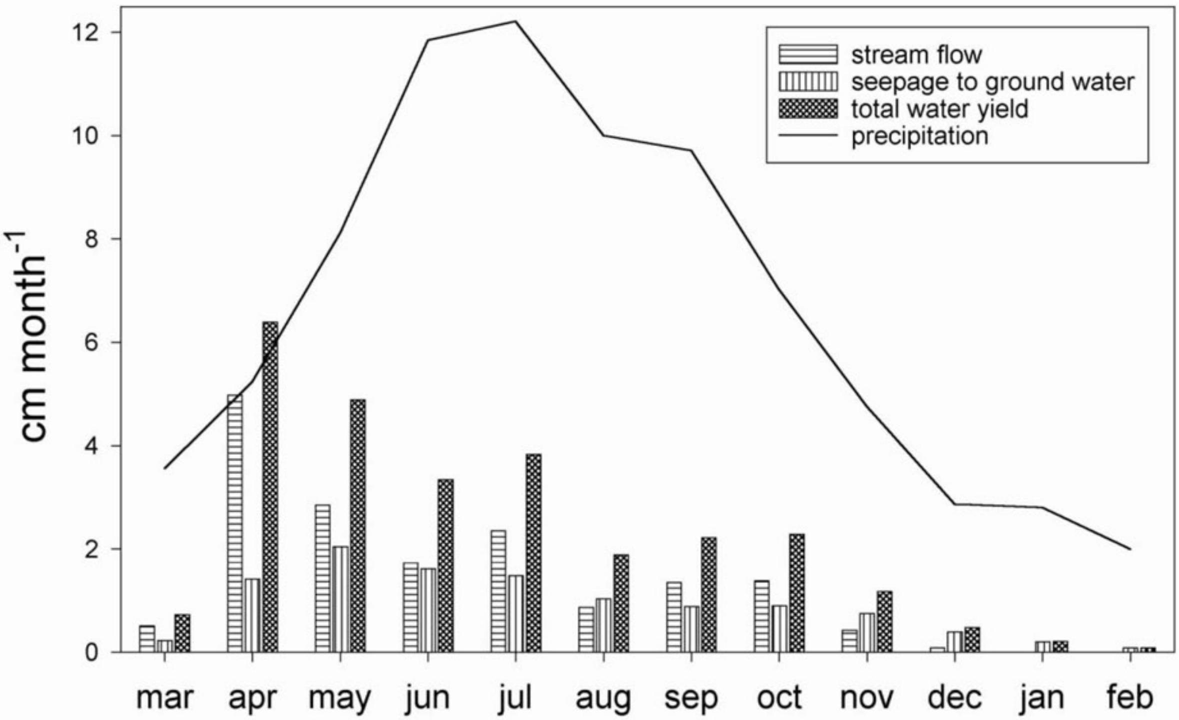

Fig. 5. Average (1981–1999) monthly stream flow and ground water recharge, watersheds 2 and 4/5.

water table elevations, taking into account the specific yield of the peat, which varies with depth and averages about 0.25.

The following discussion of water yield focuses on 1981 through 1999. Before 1981, water yield from ws-4 was affected by timber harvesting, and in some years not all data are available. The upland portion of ws-4 was clearcut during the winter of 1970–1971. For 1971–1974, 1975–1977, and 1978–1980, stream flow from ws-4 was increased by 47, 37, and 18%, respectively, over the flow that would have been expected, by comparison with the uncut ws-5, if ws-4 had not been cut. By 1981, ws-4 stream flows had essentially returned to normal, averaging only 1.4 and 1.5% above expected unharvested values for 1981–1985 and 1986–1990 (Hornbeck et al., 1993). We assume that by 1981 any effects of harvesting on ground water recharge had likewise diminished.

From 1981 through 1999, precipitation averaged 80.1 and 80.2 cm yr^{-1} for ws-4/5 and ws-2, respectively; total water yield — runoff plus seepage to ground water — was 28.5 and 27.2 cm yr^{-1}; and calculated ET averaged 51.8 and 53.0 cm yr^{-1} (Table 1). Of the total water yield, paired *T*-tests showed that the distribution between surface runoff (stream flow) and seepage to ground water was significantly different between the two watersheds. For ws-4/5, stream flow averaged 15.9 cm yr^{-1} (55% of total yield) and seepage to groundwater 12.6 cm yr^{-1} (45%). Stream flow from ws-2 averaged 17.4 cm yr^{-1} (64% of total yield) and ground water recharge was 9.8 cm yr^{-1} (36%). The probable reasons for this difference are the occurrence in ws-4/5 of about 10–15% Menahga coarse sand soil, with its high infiltration rate, and the inclusion of about 10% of the watershed area in semi-closed basins that contribute proportionately less stream flow than does the rest of the watershed.

For both watersheds, calculated ET averaged about 3.3 cm less than PET (Table 1). As previously mentioned, ET includes, but PET does not include, the average 1.8 cm of moisture estimated to be lost from the watersheds via sublimation each year. In most years, actual ET from peatlands is about equal to Thornthwaite PET (Verry, 1988; Kim and Verma, 1996). ET from uplands is generally considered to be

Table 1
Water balance for Marcell experimental forest, watersheds 2, 4, and 5[b]

	Watersheds 4 and 5								Watershed 2							
	Precipitation (cm)	Stream flow (cm)	Seepage to ground water (cm)	Total water yeild (cm)	Change in soil storage (cm)	Change in peatland storage (cm)	Calculated ET (cm)	Potential ET (cm)	Precipitation (cm)	Stream flow (cm)	Seepage to ground water (cm)	Total water yeild (cm)	Change in soil storage (cm)	Change in peatland storage (cm)	Calculated ET (cm)	Potential ET (cm)
1967	61.6	10.2	7.6	17.8	−4.1	−0.4	48.3	48.5								
1968	85.7	11.7	9.2	20.9	9.0	0.5	55.3	51.5								
1969	80.2	16.6	12.8	29.4	1.3	0.0	49.5	51.3								
1970	62.2	11.4	9.7	21.1	−2.1	−0.1	43.3	51.9	64.7	16.3	5.1	21.3	−0.6	−0.4	44.4	54.2
1971	86.6	17.7	13.8	31.4	2.8	0.3	52.0	52.2	84.0	17.2	9.1	26.3	0.7	0.6	56.4	54.2
1972	84.5	17.5	13.8	31.3	−0.9	−0.2	54.3	53.2	82.1	18.0	8.9	26.9	−1.8	−0.6	57.6	51.7
1973									74.4	15.5	5.6	21.0	0.9	0.0	52.5	55.5
1974									74.4	18.0	10.6	28.6	−1.5	0.1	47.2	51.7
1975									83.2	21.4	10.4	31.8	−3.9	−0.5	55.8	54.8
1976									45.6	5.6	2.8	8.4	−5.2	−3.5	45.9	56.7
1977									82.5	16.9	5.1	22.0	9.8	3.3	47.4	57.8
1978	89.6	20.2	17.2	37.4	−3.2	−0.1	55.5	54.8	85.2	21.	12.5	34.4	−2.4	−0.3	53.5	55.8
1979	80.8	22.5	16.4	38.9	0.2	0.1	41.6	51.4	86.8	28.5	14.8	43.3	0.0	0.9	42.6	51.3
1980	67.6	8.2	5.4	13.6	1.7	0.1	52.2	56.4	65.4	8.6	3.6	12.1	1.8	−0.6	52.1	56.4
1981	78.9	14.3	10.8	25.1	−0.9	0.1	54.6	55.3	81.6	18.6	8.3	27.0	−0.5	0.3	54.8	54.8
1982	77.0	20.9	14.4	35.3	0.4	−0.1	41.4	54.5	77.2	18.3	11.9	30.2	−0.2	−0.2	47.4	53.3
1983	68.9	6.3	6.8	13.1	−1.2	0.0	57.0	57.4	69.2	7.5	4.5	12.0	0.0	0.0	57.2	55.8
1984	78.3	15.8	11.9	27.8	0.9	0.0	49.7	57.7	82.2	18.4	10.3	28.7	−0.3	0.1	53.7	56.1
1985	87.2	21.9	15.2	37.1	−0.2	0.0	50.3	54.9	91.5	25.2	14.2	39.3	−0.6	−0.2	53.0	54.9
1986	81.2	18.1	13.5	31.5	0.8	0.1	48.7	55.7	77.6	16.5	9.6	26.2	−0.3	0.1	51.6	55.6
1987	78.0	13.0	9.7	22.7	−4.3	−0.1	59.7	59.0	81.1	17.9	8.4	26.3	−3.5	−0.2	58.5	58.7
1988	89.2	22.1	13.7	35.9	5.0	0.0	48.4	59.6	83.6	21.1	9.3	30.4	5.5	−0.2	47.9	59.4
1989	88.5	21.7	17.2	38.9	−1.1	0.0	50.7	57.1	84.4	20.3	12.8	33.1	−1.1	0.2	52.2	55.5
1990	59.0	8.6	7.7	16.3	−5.9	−0.1	48.7	56.0	58.4	6.5	3.6	10.1	−4.7	−0.3	53.3	55.3
1991	64.6	6.2	4.8	11.0	−2.9	0.0	56.5	58.8	65.7	7.5	3.2	10.7	−1.0	0.1	55.9	58.7
1992	79.7	14.1	13.3	27.4	5.4	0.1	46.8	50.5	84.3	20.8	10.5	31.3	5.0	0.0	48.0	51.9
1993	79.9	15.9	14.4	30.3	0.8	0.1	48.7	48.7	77.3	16.8	11.3	28.1	−1.9	0.3	50.8	51.3
1994	83.8	16.7	15.5	32.3	1.4	0.1	50.1	53.0	87.7	20.2	12.7	33.0	5.3	0.2	49.2	56.1
1995	86.3	17.3	13.4	30.7	0.7	0.1	54.8	52.1	84.9	19.2	10.3	29.6	−2.0	0.1	57.2	54.6
1996	77.9	16.7	13.4	30.1	−2.7	0.0	50.5	53.7	79.4	18.6	12.1	30.6	−0.6	0.2	49.2	53.2
1997	85.6	17.7	14.4	32.1	−0.6	−0.2	54.3	54.7	86.0	19.0	11.6	30.6	−1.2	−0.3	56.9	54.7
1998	73.3	8.3	8.1	16.4	0.3	−0.1	56.7	59.3	73.7	11.1	5.9	17.0	0.9	−0.3	56.1	59.4
1999	104.8	26.8	20.2	47.0	0.3	0.1	57.4	57.2	98.1	27.4	15.4	42.8	0.4	0.2	54.7	57.1
Mean, all years	80.0	16.1	12.6	28.7	−0.2	0.0	51.6	55.4	78.4	17.3	9.1	26.4	−0.1	0.0	52.1	55.2
Std. deviation	9.6	5.6	4.0	9.5	2.6	0.1	4.7	2.9	10.5	5.7	3.6	9.1	3.1	1.0	4.3	2.3
Coef. Variation	0.12	0.35	0.31	0.33	−11.03	9.90	0.09	0.05	0.13	0.33	0.40	0.35	−31.36	−31.76	0.08	0.04
Mean, 1981–99	80.1	15.9	12.5	28.5	−0.2	0.0	51.8	55.5	80.2	17.4	9.8	27.2	−0.0	0.0	53.0	55.6
Difference[a]	ns	*	**	ns			ns	ns								

[a] Between 1981–1998 means for ws-4/5 and ws-2.
[b] ns, no significant difference; *, difference significant at 0.05 level; **, difference significant at 0.01 level.

somewhat less than PET due to periodic conditions of less than optimal moisture availability. In this study, non-winter ET averaged about 5 cm less than PET. The difference between ET and PET was not significantly related to annual precipitation or growing season precipitation amounts. During dry growing seasons, ET was maintained by the capacity of the forest vegetation to draw upon water in the soil profile. The available soil moisture pool was almost always fully recharged each year by the combination of rain in the fall and snow melt the following spring.

Although ws-4/5 yielded more ground water in proportion to surface runoff than did ws-2 (Fig. 6a), the ratio between stream flow and seepage to groundwater for each watershed was similar from year to year. The proportion of total annual yield occurring as seepage to ground water in each watershed did not change with total yield nor with annual precipitation (Fig. 6b). We had anticipated that we might see a curvilinear relationship between seepage to groundwater and precipitation, with seepage tending toward a plateau in wet years as the maximum infiltration rate for the watershed was approached. This, however, did not occur.

Regardless of the difference in the runoff/seepage ratios of the two watersheds, the relationships between annual precipitation and total water yield for ws-2 and ws-4/5 were almost identical and are shown with a single regression line in Fig. 6c. While about 84% of the annual variation in water yield was explained by a single term for total annual precipitation, a better predictor of water yield was obtained by dividing annual precipitation into seasonal amounts and considering them separately, along with factors for April through October air temperature and for precipitation during July through October of the previous year. Multiple regression of water yield on this combination of variables yielded the following equation with an r^2 value of 0.96:

$$\begin{aligned} Y_{(q+g)} = &-16.22 + 0.7587P_{\mathrm{nm}} + 1.0486P_{\mathrm{am}} \\ &+ 0.8107P_{\mathrm{jun}} + 0.0161P_{\mathrm{ja}}^2 + 0.8435P_{\mathrm{so}} \\ &+ 0.2346P_{\mathrm{pvjaso}} - 5.9515(T_{\mathrm{ao}}/5)^{1.514} \end{aligned} \quad (1)$$

where, $Y_{(q+g)}$ is the total annual water yield (stream flow + ground water recharge), cm yr^{-1}; P_{nm} the precipitation November through March, cm; P_{am} the precipitation in April and May, cm; P_{jun} the precipitation in June, cm; P_{ja} the precipitation July and August, cm; P_{so} the precipitation in September and October, cm; P_{pvjaso} the precipitation in July through October of the previous year, cm; and T_{ao} is the mean air temperature April–October, °C.

Every variable in the regression was significant at the $p = 0.0001$ level. There was essentially no difference in yield response between ws-2 and ws-4/5 (Fig. 6d). Eq. (1) was derived using data from both watersheds.

The coefficients for the variables reflect the relative proportional contribution to water yield of precipitation falling during different periods of the year. For example, as the snow melts in spring, some of the melt water is yielded to stream flow and ground water recharge, but some is retained in the watershed, filling available storage capacity in the soil or in the peatlands and other depressions. April and May precipitation probably falls on a watershed with limited storage capacity, so yield is proportionally greater. The relationship between precipitation amount and total water yield was found to be essentially linear for all periods of the year except July and August. In July and August, ET typically exceeds precipitation. Water is withdrawn from storage in depressions and in the soil. Much of the water supplied by rain is lost to evaporation or infiltrates into dry soils. A relatively large amount of precipitation is needed to produce runoff or seepage to ground water. Response of yield to July–August precipitation was seen to be exponential rather than linear, with an exponent of 2.0 giving us the best fit for the P_{ja} term in Eq. (1). For the period of record, July–August precipitation at the MEF ranged from about 6.8 cm in 1990 to 39.6 cm in 1999. Over this range, the yield predicted by Eq. (1), with the $P_{\mathrm{ja}}^{2.0}$ term, fit the actual yield data slightly better than values predicted by a similar equation with a $P_{\mathrm{ja}}^{1.0}$ term.

We found that the amount of precipitation during the summer and early autumn of one year had a significant effect on water yield the following year, as shown by the P_{pvjaso} term in Eq. (1). In spring following a particularly dry summer, it's likely that deficits in the soil moisture and depressional storage pools are filled by water from melting snow that would otherwise runoff as stream flow or seep downward to recharge the ground water. Conversely, after a wet

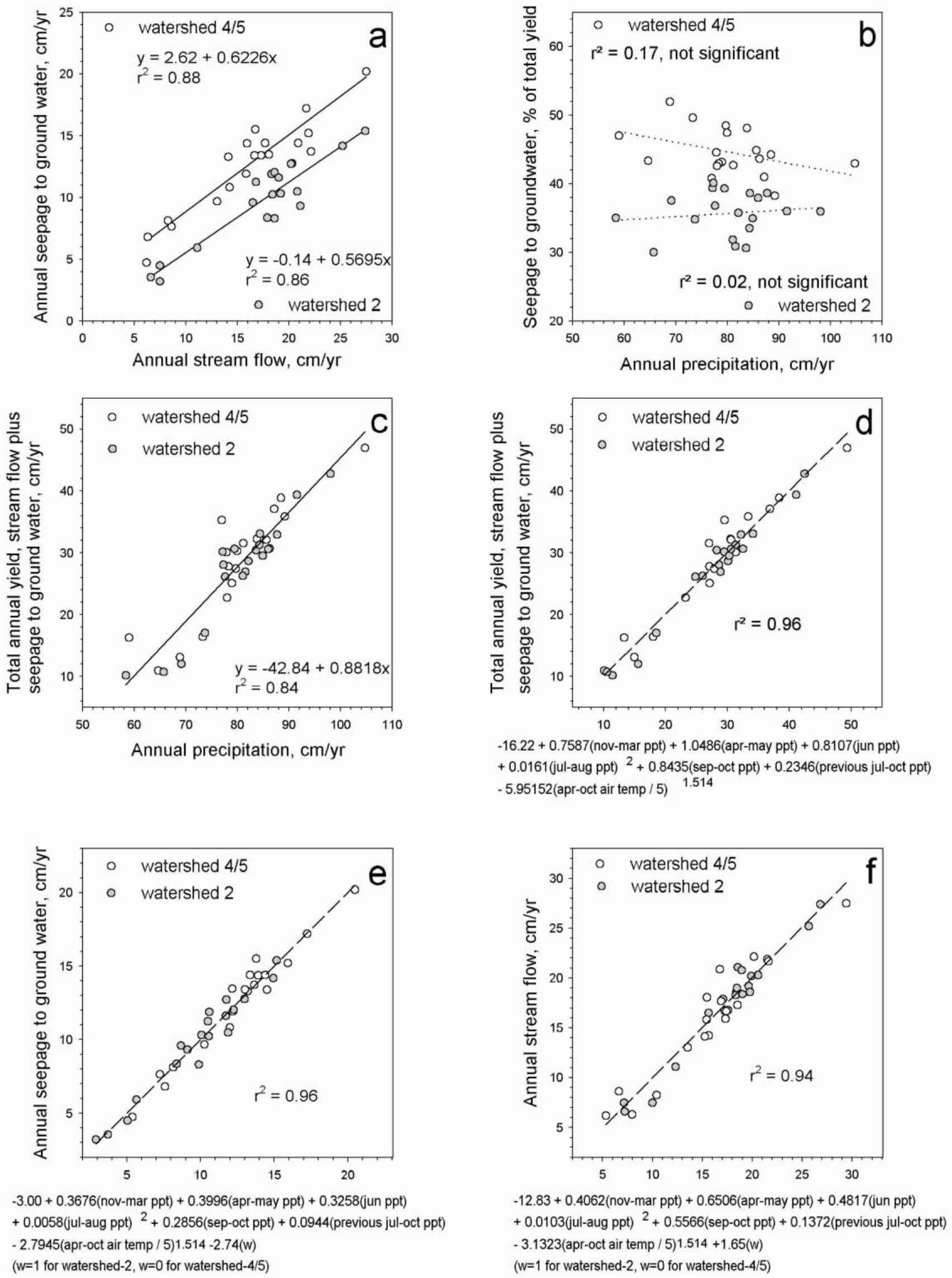

Fig. 6. Relationships among ground water recharge, stream flow, and precipitation, watersheds 2 and 4/5.

summer, complemented by autumn rains, a watershed may go into winter with saturated soils and filled depressions, leaving little capacity to hold water on-site in spring.

The final variable in Eq. (1), $(T_{ao}/5)^{1.514}$, accounts for the effect of air temperature on ET. Thornthwaite PET is related to a heat index, the sum of the monthly values of $(t/5)^{1.514}$, where t is the mean monthly temperature (for those months for which $t > 0$°C). While Thornthwaite PET was a significant variable in the water yield regression equation, we found that the term $(T_{ao}/5)^{1.514}$ fit the data just as well, and is easier to calculate.

We also derived separate regression equations for stream flow and ground water recharge.

$$Y_{(q)} = -12.83 + 0.4062P_{nm} + 0.6507P_{am} + 0.4817P_{jun} + 0.0103P_{ja}^2 + 0.5566P_{so} + 0.1372P_{pvjaso} - 3.1323(T_{ao}/5)^{1.514} + 1.6467\mathrm{WS}, \tag{2}$$

r-squared $= 0.93$, and

$$Y_{(g)} = -3.00 + 0.3676P_{nm} + 0.3996P_{am} + 0.3258P_{jun} + 0.0058P_{ja}^2 + 0.2856P_{so} + 0.0944P_{pvjaso} - 2.7945(T_{ao}/5)^{1.514} - 2.7380\mathrm{WS}, \tag{3}$$

r-squared $= 0.96$,where, $Y_{(q)}$ = annual stream flow, cm yr^{-1}; $Y_{(g)}$ = annual ground water recharge, cm yr^{-1}; and WS is a dummy variable with a value of 1 for ws-1 and 0 for ws-4/5.

The dummy variable was necessary because, although the two watersheds yield almost identical amounts of water, the relative proportions of surface water to ground water yielded are different. Except for the dummy variable, the variables are the same as in Eq. (1). For both Eqs. (2) and (3), all variables are significant at the $p = 0.001$ level and most are significant at the $p = 0.0001$ level. Both equations appear to be linear over the entire range of precipitation and summer air temperature conditions occurring between 1981 and 1999 (Fig. 6e and f). Even in 1999, when precipitation was, by far, the highest on record at the MEF, ground water recharge did not appear to be limited by the infiltration capacity of the watershed. On an average, water yield from the two watersheds is about 60% stream flow and 40% ground water. Comparison of the coefficients of the variables in Eqs. (2) and (3) indicates that winter precipitation is relatively more important to ground water recharge, while non-winter precipitation is more important to stream flow. However, precipitation in all seasons contributes to both recharge and stream flow.

Fig. 7 shows monthly precipitation, stream flow, available soil moisture, seepage to ground water, and water table elevation for ws-2 in 1988–1990. Fig. 7 shows how various components of the hydrologic balance function in relation to each other on the MEF, during a period with a diversity of precipitation conditions.

Total precipitation in water year 1988 was 83.6 cm, a little greater than the annual average of 77.9 cm. However, the distribution of precipitation throughout the year was highly skewed, with over 60% of the total occurring in a 50-day period in July and August. Soils were relatively dry in November of 1987, with several centimeters of unfilled water storage capacity available. Precipitation from November 1987 through May 1988 was about 70% of normal, and precipitation in June and July was about a third of the normal amount. As a result, stream flow from the watershed during April and May was low — 3.8 cm compared to an April–May average of 19.9 cm — and no flow at all was recorded in June and July. A similar pattern was seen in ground water recharge. Seepage to ground water was 2.0 cm from April–June, and was essentially zero in July. Normal recharge for that 4-month period is 6.2 cm. August and September brought 53.1 cm of precipitation, almost three times the average rain fall for those 2 months. In response, August–September stream flow was 16.7 cm, about seven times the August–September average and almost 80% of the total for the year. Following that precipitation outburst, the last 10 days of August and the month of October were dry, and stream flow fell to below average conditions and stayed that way for the rest of the year. The response of ground water recharge to the heavy August–September rains was more subdued and longer lasting than that of stream flow. With the rain, dry soils became saturated and depressional storage areas filled. As this excess water drained through the soil, ground water recharge was sustained at about twice the average rate from August through

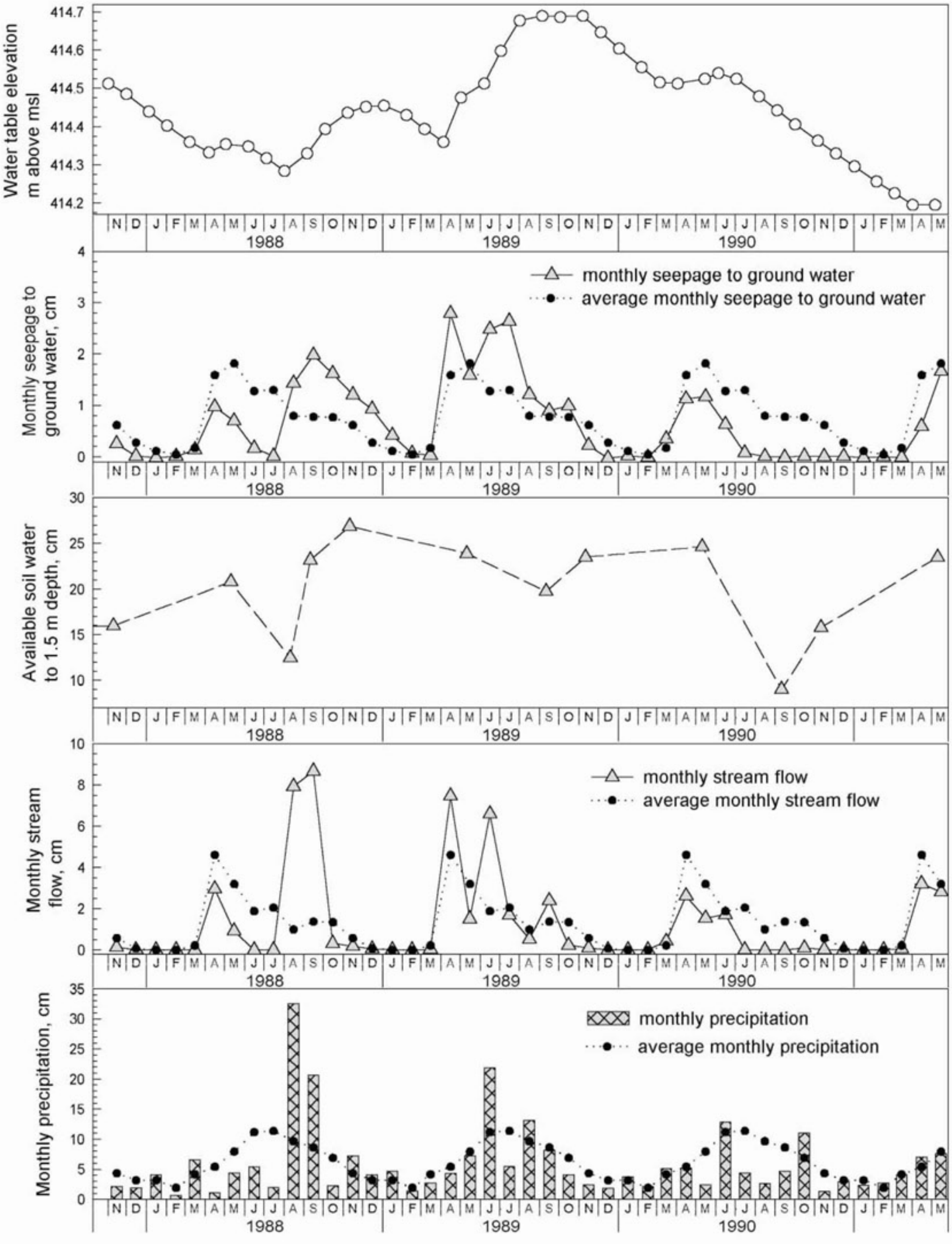

Fig. 7. Water balance, 1988–1990, watershed 2.

February. The water table, which had declined to its low point for the year in midsummer, rose continuously through the autumn and early winter and reached its highest elevation for the year in January.

As in 1988, precipitation in water year 1989, 84.4 cm, was also a few centimeters above average, but its distribution pattern through the year was much closer to normal. The combination of wet soils the previous autumn and slightly higher than average (5%) amounts of snow and rain from November through May resulted in stream flow about 15% above normal and ground water recharge about 28% above normal during April and May. A wet June, a dry July, a more or less normal August and September, and a dry October are reflected in the pattern of stream flow from the watershed. The effects on ground water recharge were somewhat delayed; recharge was sustained at high levels through July and at somewhat higher than normal rates through October before diminishing to close to zero by December. The water table reached its highest elevation for the year in November and then declined steadily through the winter.

Water year 1990 was the second driest year since record keeping began for ws-2 in 1962, with a total precipitation of 58.4 cm compared to an annual average of 77.9 cm. November–March precipitation totaled 15.2 cm, slightly less than the average of 16.7 cm, but April–May precipitation was little more half the 13.4 cm average. The resulting March–May stream flow totaled 4.7 cm, compared to an average of 8.0 cm. June precipitation, 12.8 cm, was slightly above the 12.2 cm average, and June stream flow, 1.7 cm, was close to the average 1.9 cm. Seepage to ground water from late March through June was about 3.4 cm, compared to an average of about 4.9 cm. Precipitation during July, August, and September totaled 11.7 cm, about half of the normal 22.7 cm, and much less than the 31.7 cm of PET for that period. As a result, both stream flow and seepage to ground water were essentially zero for the rest of the year and did not begin again until spring of 1991. Water table elevation peaked in July and then declined steadily until early May of 1991. By September, available soil moisture was greatly depleted. Above average precipitation in October (11.0 cm compared to an average 6.9 cm) only partially replenished soil moisture and produced insignificant runoff (0.1 cm) and no ground water recharge.

The water yields reported in this paper compare well with stream flows measured downstream. Most of the water yielded by the MEF, both surface and ground water, flows to the Prairie River, a tributary of the Mississippi River. Between 1967 and 1982, the U.S. Geological Survey measured daily flows in the Prairie River at a point about 18 km south southeast of the MEF. At this point the river is 35–40 m lower in elevation than the top of the watershed on the MEF and has a watershed of about 930 km^2. Most of the watershed is forested, but there is some agricultural activity, mostly pasture and haying. Between 1967 and 1982, the average annual discharge of the Prairie River at the USGS gauging station was 21.8 cm, from an average annual precipitation over the watershed of about 72.1 cm. For the same period, we calculate the average total water yield (surface plus ground water) from ws-4/5 and ws-2 to have been 25.2 cm, out of 75.5 cm of precipitation. We assume that flow in the river at the gauging station represents almost all of the water leaving the watershed, with ground water flow being relatively minor. The station is about 2 km north of the Mesabi Iron Range, a bedrock high overlain by only a thin layer of glacial material.

4. Conclusions

On the MEF in north-central Minnesota, ground water table elevations measured in observation wells in recharge areas can be used to calculate rates of ground water recharge. Because winter precipitation is stored on the surface as snow and ground water recharge ceases, water table elevations in recharge areas decline over winter at calculable rates. Deviations from these rates during other times of the year can be attributed to ground water recharge. On 10–50 ha watersheds on the MEF, ground water recharge varies somewhat among watersheds but constitutes about 40% of the total water yield. We found that annual ground water recharge amounts vary linearly with precipitation. The ratio of ground water recharge to stream flow from the watershed was about the same in low precipitation and high precipitation years. On the MEF watersheds, 93–96% of the variation in annual ground water recharge, stream flow, and total water yield was explained by multiple regression equations in which seasonal precipitation amounts, summer and autumn precipitation during the previous year, and non-winter air temperature are the independent variables.

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