Measured river leakages using conventional streamflow techniques: the case of Souhegan River, New Hampshire, USA

Philip T. Harte · Richard G. Kiah

Abstract Multiple streamflow measurements were made at coupled discharge measurement stations to quantify rates of aquifer recharge and discharge on two reaches of the Souhegan River, New Hampshire, USA, flowing within a glacial-drift river-valley aquifer. The reaches included a predominantly losing (aquifer recharge) reach and a variable (aquifer recharge and discharge) reach located downstream of the former reach. River leakage, the differential between coupled upstream and downstream streamflow measurements along a reach, varied by almost 30 cubic feet per second (ft^3/s) ($0.85 m^3/s$) along the two reaches. The upper reach averaged $3.94 ft^3/s$ ($0.11 m^3/s$) loss whereas the lower reach averaged $4.85 ft^3/s$ ($0.14 m^3/s$) gain. At the upper reach, 13 losses were measured out of 19 coupled measurements. At the lower reach, ten out of 13 coupled measurements indicated gains in flow and suggest that this reach is primarily a gaining river reach. An important factor in river leakage appears to be antecedent trends in river stage. At the upper reach, gains were measured only during periods of declining river stage. Conversely, at the lower reach, streamflow loss was measured primarily during periods of rising river stage. Although some tendencies exist, several factors complicate the analysis of river leakage, most notably the inaccuracies in computed stream discharge.

Keywords Groundwater/surface-water relations · River leakages · Streamflow · Aquifer recharge · USA · Conceptual models · Rainfall/runoff

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Introduction

Measuring gains and losses in flow along river reaches is an important tool in quantifying potential rates and patterns of groundwater recharge and discharge in many aquifers. It is one of the few quantitative measurements of fluxes in groundwater hydrology. The flows at coupled upstream and downstream gaged stations are subtracted from one another and the difference (called river leakage in this report) is used to identify rates of streamflow gains or losses. The technique is often applied during periods of low-flow (Harte and Mack 1992) to ensure isolation of groundwater interactions (aquifer recharge and discharge) on streamflow, called baseflow conditions. However, this approach is problematic for several reasons. While measuring during low-flow improves the chances of measuring only baseflow conditions, it may lead to an underestimation (or in some cases overestimation) of aquifer recharge and discharge from river–aquifer interactions. For example, river loss to a valley-fill aquifer was found to increase (six-fold) when river stage increased along two tributaries within a valley draining Marsh Creek in north-central Pennsylvania, USA (Kontis et al. 2004). Therefore, low-flow river-leakage rates (defined in this report as streamflow exceedance rates more than 90% flow) underestimated the amount of potential aquifer recharge from these tributaries.

Single or a few measurements of river leakage will insufficiently represent the realm of hydrologic conditions. The transient nature of groundwater and surface-water interactions suggest that aquifer recharge and discharge data obtained from river-leakage measurements should be augmented with more periodic (monthly) measurements throughout the year so that representative rates of average annual groundwater recharge and discharge can be computed. Many water-supply studies of surface-water and groundwater resources typically quantify storage and flow under average annual conditions (Harte and Mack 1992). Furthermore, additional measurements of streamflow will allow for quantifying temporal variability and assessing inaccuracies in river leakage from streamflow measurements.

Previous studies documented in the literature lack a comprehensive data set of both temporal and spatial information on river leakage. Often, the amount of available data on either temporal or spatial information is inversely related. For example, temporal information on river leakage is available from the application of baseflow separation techniques to the record at continuous streamgage stations (Rutledge 1993). However, spatial coverage is lacking because streamgage stations are generally located far apart from each other and integrate the effects of runoff from large areas and long streamflow reaches; thus, information on specific reaches are unavailable and consequently rates of aquifer recharge from streamflow loss and patterns of aquifer recharge and discharge are unavailable. A good example of this approach is a study by Atkins et al. (1996), who estimated groundwater discharge in the coastal plain of Georgia. Conversely, spatial information on river leakage can be refined by making streamflow measurements along specific reaches, however, measurements of temporal variability are often lacking and there is a bias toward making the measurements during low-flow conditions. A good example of this approach is a study by Dysart and Rheaume (1999), in which detailed streamflow measurements were made adjacent to a well field in the glaciated northeastern United States.

Recently, some river-leakage studies have attempted to measure temporal variability under various streamflow conditions with a high spatial resolution. Kontis et al. (2004) showed that river leakage from induced infiltration of groundwater withdrawals occurred at several sites in the glaciated northeastern U.S. adjacent to nearby well fields and that large seasonal variability in river leakage occurs. Lee and Risely (2002) measured river leakage using coupled streamflow measurements for streamflow duration conditions of 50–85% streamflow exceedance and found aquifer discharge (streamflow gain) rates in the spring to be much higher than those in the summer. Simonds and Sinclair (2002) also found variations in streamflow gains/losses during three sets of measurements during 20, 50, and 70% streamflow exceedance.

Multiple coupled measurements of flow were made along two river reaches to compute river leakage and to assess potential rates of river-aquifer recharge and discharge in a glacial-drift valley aquifer in Milford, New Hampshire (NH), USA. The primary objective of the study was to improve estimates of average annual rates of aquifer recharge and discharge from streams and to quantify seasonal variations. This study was part of an effort to remediate a volatile organic compound plume that is present in the glacial-drift river-valley aquifer. The improved values of river leakage helped better quantify estimates of aquifer recharge and discharge, thus improving numerical transport simulations of the contaminant plume by providing ranges in river–aquifer interactions. This report describes the results of this effort, identifies issues related to the use of streamflow measurements to compute river leakage, and highlights some important processes affecting river leakage.

Reliability in river-leakage measurements

The differential (river leakage) between coupled streamflow measurements along a designated reach can incorporate a number of processes including groundwater recharge and discharge, runoff, bank storage, and interflow. Two main factors affect the use of river leakage in quantifying groundwater recharge and discharge; they are (1) inaccuracies in the computed leakage, and (2) processes other than baseflow.

The potential inaccuracies associated with computed river leakages can be large, given the uncertainty of the measurement themselves. Under the best streamflow measurement conditions, minimum potential errors (standard error between the true and measured streamflow) are ±2% of each measurement deemed “excellent” and ±5% for measurements deemed “good” based on qualitative ratings from field observations given by the hydrographer (Rantz 1982a). Because leakages are computed from coupled measurements, the inaccuracies are considered cumulative (4% for “excellent” and 10% for “good” rated measurements). This is a persistent problem with the coupled streamflow measurement technique as river-leakage values commonly are less than 10% of the total streamflow (Simonds and Sinclair 2002; Dysart and Rheaume 1999). Few studies have river-leakage measurements persistently above 10% of total streamflow. In most cases, rates of streamflow loss exceeding 10% of total streamflow occur in less humid climates (Williams-Sether 2004) characterized by precipitation to potential evapotranspiration ratios less than 0.50.

A quantitative assessment of measurement inaccuracies is possible following techniques described by Sauer and Meyer (1992), which may further refine estimates of inaccuracies beyond the qualitative field observations of the hydrographer. The quantitative assessment incorporates a summation of the standard error of multiple factors that affect the accuracy of the streamflow measurements. Sauer and Meyer (1992) suggest that most streamflow measurements are within 3–6% of their true value.

The reliability of streamflow measurements in the quantification of river leakage and identification of patterns of gains and losses along reaches is typically validated against alternative methods and data. These alternative methods include hydraulic head measurements, seepage meters, temperature profiling, and geochemical information. Dysart and Rheaume (1999) found that temperature profiling and thermal vertical one-dimensional modeling of the temperature profile resulted in a calculated streamflow loss of 1.8 cubic feet per second (ft^3/s) (0.051 cubic meters per second (m^3/s)) compared to a measured streamflow loss of 0.67 ft^3/s (0.019 m^3/s). Seepage meters, which are typically installed in the central part of the river channel to measure vertical flow between the river and aquifer, provide small-scale “point type” measurements of flow that may poorly correlate with river-leakage measurements due to variations in measurement scale (Kontis et al. 2004). Hydraulic head data that measure horizontal or vertical surface- and groundwater gradients provide information also on relatively small scales compared to the size of the reach unless a large data array is used.

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River-leakage processes and preliminary conceptual model of river–aquifer interactions

Factors that affect river leakage include geomorphology of the stream, streambed properties, temperature, groundwater evapotranspiration (ET), aquifer geometry and hydraulic properties, groundwater extraction in the aquifer, and local and regional groundwater flow patterns. Fryar et al. (2000) identified characteristic patterns of groundwater discharge to streams based on physical properties of watersheds. Their conclusions were based on streamflow at gaged stations computed from rating (stage-discharge) curves rather than discrete streamflow measurements. The effects of temperature changes on streamflow loss and gain were examined by Constantz et al. (1999). Large temperature ranges (20° Celsius) along reaches with streamflow loss affect the rate of infiltration and potentially can change infiltration rates by a factor of two. This same temperature range has a smaller effect on infiltration rates along reaches with streamflow gain.

Groundwater ET can occur from phreatophyte water uptake and may induce streamflow loss regardless of whether the river is gaining or losing (Meyboom 1964). Groundwater ET has been shown to occur in arid and semiarid environments (Chen and Shu 2006) and more humid environments (Batelaan et al. 2003). Direct evidence of groundwater ET has been observed from diurnal fluctuations of water levels in wells adjacent to streams (Rosenberry and Winter 1997).

An important factor in controlling river leakage is the head difference or relative position of the river stage and the adjacent groundwater head over time (Winter et al. 1998). Rates of river leakage are a function of the difference between river stage and aquifer head. Mathematical models representing confined and unconfined aquifer head and river stage show that river leakage is best represented by a combined linear and non-linear response function (Rushton and Tomlinson 1979).

River–aquifer responses differ depending on the predominant horizontal or vertical-head gradient between the river and aquifer. Where the river stage is higher than the aquifer head (the water-table surface in unconfined aquifers), such as along a predominantly losing reach (a decrease in streamflow downstream), computed river leakages that show streamflow loss are a strong indicator of groundwater recharge to the aquifer. In contrast, where the river stage is less than the aquifer head, such as along a predominantly gaining reach (an increase in streamflow downstream), runoff into the reach may contribute to streamflow gain, which complicates the computation of groundwater discharge to the reach. Thus, depending on the relation between river stage and aquifer head, river-leakage rates may be indicative of different hydrological processes and aquifer interactions.

An important process in river–aquifer interactions is bank storage, which is conceptualized to affect river leakage differently along losing and gaining reaches. Bank storage occurs when water flows temporarily into the aquifer during rising river stage, only to return to the stream during declining stage. Along losing reaches, aquifer head slopes away from the river and bank storage would tend to increase the gradient away from the river rather than reverse the gradient as along gaining reaches. While some of this infiltrated bank water may return to the river when the stage declines, it is hypothesized that a part will continue to be transported away from the river. Therefore it is plausible that bank storage is more prevalent along gaining reaches than losing reaches.

Certain reaches may appear to exhibit spatially variable river leakage, and the streamflow measurement may integrate river-leakage results from the larger scale. For example, reaches on a local scale (less than 100 feet (ft) (30.5 meters (m)) can be losing or gaining within a predominantly larger (1,000 ft) (305 m) gaining or losing reach, respectively. For this reason, streambed piezometer or seepage meters, which measure small scale “point type” tendencies, may not corroborate river-leakage results from streamflow measurements of larger reaches (Dumouchelle 2001).

Transient changes in river stage and aquifer head affect the rates and timing of river leakage. Because of the relative rapid response of river stage compared to aquifer head, large rates of river leakage or flow reversals (Winter et al. 1998) may occur. Climatic events that induce transient changes in river stage and streamflow from storms, precipitation and(or) melting of snow pack affect the relative contributions to streamflow from different processes. Bank storage, runoff, and interflow are temporally dependent and contribute different amounts of streamflow over time (Dunne and Leopold 1978). Harte et al. (1999) showed that baseflow contributions to streamflow varied by almost a factor of three between periods of low precipitation and(or) high evapotranspiration, and periods of high precipitation and(or) low evapotranspiration.

Study area

The field site for this study is in the Souhegan River valley in Milford, New Hampshire, USA (Fig. 1). The Souhegan River valley in the Milford area is relatively flat and gently sloping. Land surface elevations range from 230 (70.1 m) to 280 ft (85.3 m) a.s.l. The valley is drained by the Souhegan River and its tributaries, which include Tucker Brook, Purgatory Brook, Great Brook, Hartshorn Brook, and a number of small unnamed streams. A discharge ditch drained processed waters from several manufacturing companies in the southwest part of the study area from 1965 to 2002.

The valley is underlain by unconsolidated sediments identified as the Milford-Souhegan glacial-drift aquifer (Harte and Mack 1992), hereafter referred to as the MSGD aquifer. The MSGD aquifer is defined as the entire sequence of unsaturated and saturated alluvium, glacial drift and other unconsolidated deposits above the bedrock surface in the Souhegan River valley in Milford. The aquifer consists primarily of stratified sand and gravel

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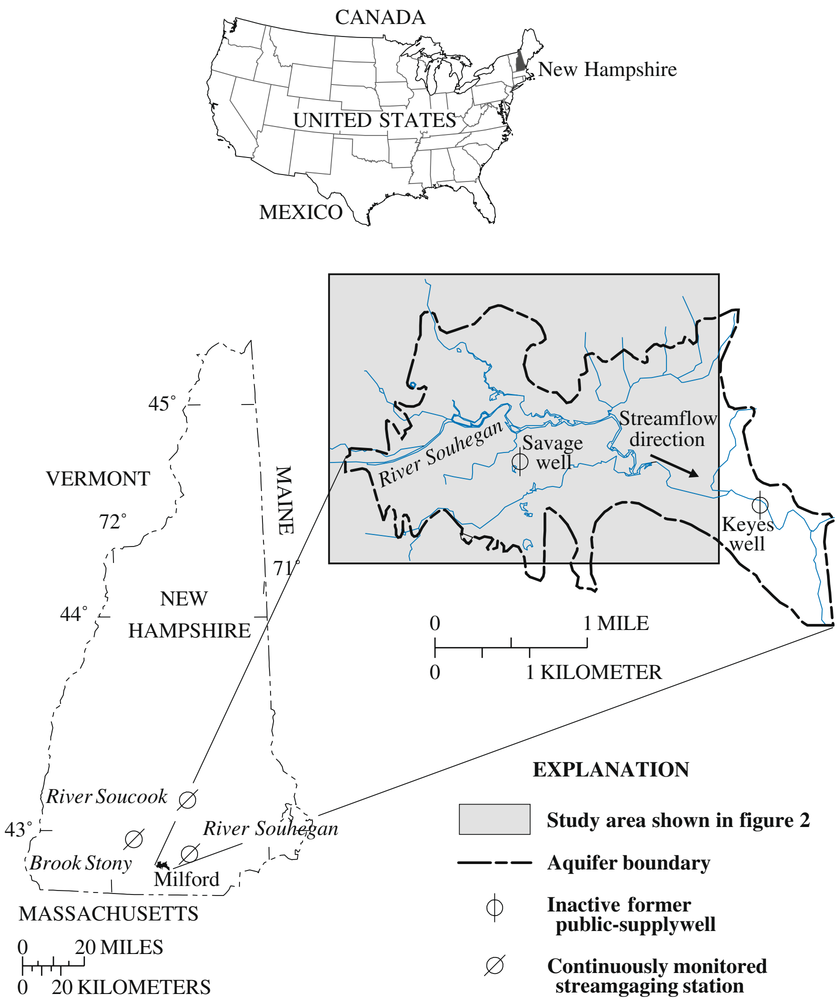

Fig. 1 Location map of study area in New Hampshire, USA

with some basal till, and is overlain in places by alluvium. The maximum saturated thickness of the aquifer exceeds 100 ft (30.5 m) in its eastern part, but generally ranges from 0 to 60 ft (18.3 m). Laterally, the aquifer is bounded by till-covered bedrock uplands. The aquifer is primarily unconfined and the water table ranges in depth from 8 (2.4 m) to 16 ft (4.9 m) below the land surface.

The stratigraphy in the near surface (upper 20 ft (6.1 m)) of the MSGD aquifer is comprised of poorly sorted sediment layers of cobbles, gravel, and fine sand interlayered with well-sorted sediment layers of medium to coarse sand in the western part of the aquifer. Gravel and sand layers are interlayered with medium sand layers in the eastern part of the aquifer.

Flow in the Souhegan River ranges from 10 (0.283 m^3/s) to 1,000 (28.3 m^3/s) ft^3/s and river stage varies by up to 7 ft (2.13 m) (Harte et al. 1997). There is a good hydraulic connection between the river and the aquifer and the aquifer receives recharge from the Souhegan River and its tributaries along some reaches. Along other reaches, groundwater discharges to the Souhegan River and its tributaries. Baseflow was estimated to contribute 71% of streamflow in the Souhegan River valley from July 1994 to September 1995, based on hydrograph separation techniques (Harte et al. 1997).

The primary groundwater flow direction in the valley is from west to east. The slope of the water table is about 0.006 ft/ft (0.006 m/m). Large groundwater withdrawals (approximately 1.55 ft^3/s (0.04 m^3/s) each) were made from two wells operated by the New Hampshire State Fish Hatchery in the north-central part of the valley; these withdrawals induce streamflow infiltration along the central reaches of the Souhegan River (Fig. 2). Groundwater flow near the large withdrawals is asymmetrically radial because of the large amounts of induced infiltration that limit drawdowns to the south, near the Souhegan River. Relatively small groundwater withdrawals (0.34 ft^3/s (0.01 m^3/s)) were made from one well (PFH; Fig. 2) operated by a private fish hatchery in the east-central part of the valley. Withdrawals at PFH, FH-4, and FH-5 wells (Fig. 2) were relatively constant during the period of study.

A large plume of volatile organic compounds (VOCs) is present within the MSGD aquifer and underlying bedrock (Harte and Mack 1992). The source of the plume has been identified as an area adjacent to the upper reaches of the Souhegan River where VOCs (mostly tetrachloro-

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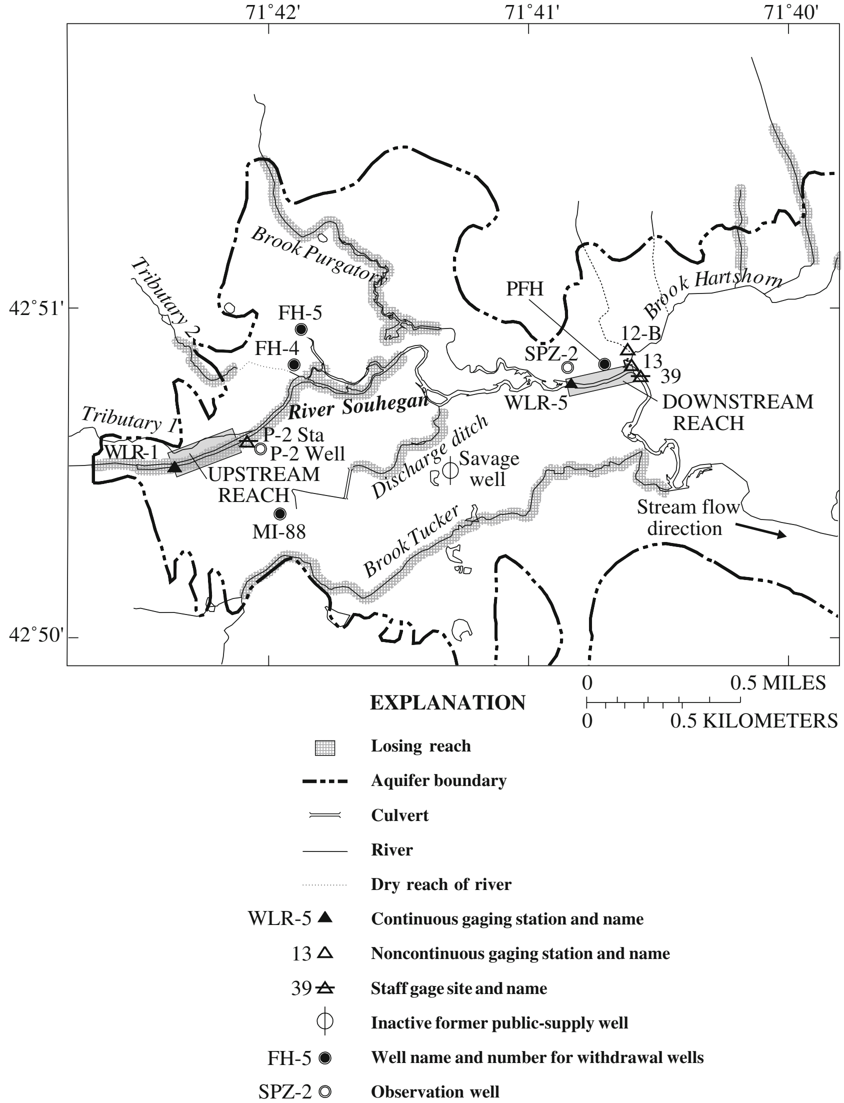

Fig. 2 Patterns of river leakage based on streamflow measurements during low-flow conditions and locations of upper and lower reaches

ethylene) were disposed into the subsurface prior to the early 1980s (HMM Assoc. Inc. 1991). In 1983, contaminants were detected at the former municipal water-supply well, called the Savage well (Fig. 1).

Approach

Discrete measurements of streamflow were made with current meters by methods adopted by the U.S. Geological Survey (USGS) as described in Rantz (1982a), and in the USGS Techniques of Water-Resources Investigations (U.S. Geological Survey 2007). These methods are consistent with the American Society for Testing and Materials standards.

Multiple coupled streamflow measurements were made to (1) identify changes in river-leakage patterns and rates of aquifer recharge and discharge; (2) identify patterns of gains and losses; and (3) provide information on potential variability of river leakage along two reaches of the Souhegan River. Multiple coupled streamflow measurements may allow for the inaccuracies inherent in the measurements to be randomly distributed, resulting in an equal distribution of computed river leakages that may be biased high or low. The average leakage value (or some other descriptor of the central tendency), in this instance, may be a good indicator of whether a reach is predominantly gaining or losing.

Discharge measurements were made during high-flow conditions in April 1994 and monthly from June 1994 through the summer of 1995. Monthly measurements provided a good range of streamflow conditions to measure.

Successive discharge measurements of coupled upstream and downstream stations took about 3 h to perform. Given that the time of travel of streamflow between stations was less than 10 min, it was important to examine trends in river stage to determine if conditions

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were stable. If river stage fluctuations exceeded 0.20 ft (0.0610 m) during the time of the measurements, the computed river-leakage rates were omitted from this analysis. For several sets of measurements that indicated only minor fluctuations, the effect of streamflow changes during the coupled measurements were eliminated by adjusting flow to a uniform stage by use of rating curves.

The selection of reaches for comprehensive coupled streamflow measurements was based on previous studies conducted on the Souhegan River. Harte and Mack (1992) identified areas with river-leakage losses and aquifer recharge areas in the upper reaches of the Souhegan River and several tributaries and variable river leakage (temporally dependent reach experiencing both gains and losses) in the central and eastern reaches of the Souhegan River (Fig. 2) using coupled streamflow measurements during periods of low flow. Harte and others (1999) and Brayton (2001) identified more detailed patterns of potential river leakage, particularly along reaches identified as variable, by utilizing surface-water and groundwater horizontal and vertical gradients determined from river stage, water levels in adjacent wells, and streambed piezometers. Harte and others (1997) used horizontal-head gradients between river stage and adjacent wells to show that the upper reach (site P-2; Fig. 2) lost water and the lower reach (site WLR-5; Fig. 2) gained water on one side of the river and lost water on the other side indicating flow-through conditions. Streambed piezometers installed at those locations showed a losing upper reach and a lower reach in which leakage was variable and neither gain nor loss was dominant (Brayton 2001).

Additional analyses were made on data from the upper reach to evaluate river leakage. An analytical model of river–aquifer interactions (Barlow and Moench 1998) was used to compute river leakage resulting from a 0.4 ft (0.12 m) rise in river stage observed during a period with no precipitation (Brayton and Harte 2001). The analytical model computed a maximum rate of leakage and river loss of 30 ft^3/s (0.85 m^3/s). Computed flow reversals and gains also resulted during a simulated decline. A numerical model of the aquifer (Harte et al. 1997) computed a maximum river loss, averaged over a 1-day period, exceeding 3 ft^3/s (0.9 m^3/s).

Based on the previous identification of losing and gaining river reaches presented above, two reaches were selected on the Souhegan River to obtain multiple coupled measurements of streamflow. Reaches included an upper, predominantly losing river reach (between river stations WLR-1 and P-2; Fig. 2) and a lower variable river reach (between river stations WLR-5 and 39; Fig. 2). Streamflow was monitored continuously at station WLR-1 at the upper reach. Streamflow at this station was used to correct for time changes in flow during the course of flow measurements. The coupled measurement stations were approximately 1,100 ft (335 m) apart for each reach. The end of the upper reach and the beginning of the lower reach are approximately 1 mile (1.6 km) apart. The upper reach was approximately 60 ft (18.3 m) wide with a cobble and sandy bed and a longitudinal river-stage gradient of 0.006 ft/ft (0.006 m/m). The lower reach was about 55 ft (16.8 m) wide and consists of a primarily sandy streambed bottom with a longitudinal rivers-stage gradient of 0.003 ft/ft (0.003 m/m).

The aerial extent and depth (volume) of the MSGD aquifer differs between the two reaches. At the upstream reach, the aquifer thickness, volume, and bulk transmissivity increases downstream. This spatial pattern facilitates streamflow loss because the aquifer head gradient is steeper than the river head gradient. At the downstream reach, aquifer thickness decreases downstream facilitating aquifer discharge and streamflow gains.

Results and discussion

River leakages were compared to a number of potential factors to identify important processes. These included: (1) the relative amount of river leakage compared to the average streamflow of the coupled measurements (called the percent difference of the streamflow differential), (2) changes in temporal trends of streamflow during measurements based on continuous streamgage data (for the upper reach only), and (3) a number of other environmental factors such as precipitation and water temperature. Evaluation of river leakage based on percent difference of streamflow indicates whether the amount of computed river leakage is large relative to streamflow and therefore is a reliable indicator of the true river leakage. Changes in trends in streamflow allow for evaluation of tendencies in river–aquifer interactions. Evaluation of other factors, such as water temperature, allow for identification of potential causal effects of river leakage.

Measured river leakage, the differential between a coupled upstream and downstream discrete discharge measurement station, varied by approximately 30 ft^3/s (0.85 m^3/s) at the upper reach and by approximately 26 ft^3/s (0.74 m^3/s) at the lower reach (Tables 1 and 2). Large losses and gains in river leakage were measured for both reaches. Thirteen of 19 measurements (68%) at the upper reach indicated losses in flow whereas ten of 13 (77%) for the lower reach indicated gains (Fig. 3a,b). The mean and median statistics reflect the tendency for the upper reach to lose flow and for the lower reach to gain flow (Table 2).

Correcting for temporal changes in streamflow during measurements at coupled stations, by use of a generated stage-discharge rating curve from the continuous streamgage station (WLR-1; Fig. 2) for the upper reach, results in generally small modifications to river leakage (Table 1). Typically, it took 3 h to measure discharge at two coupled stations. If time variations in streamflow during the measurement exceeded the actual gain/loss along the river reach, then the computed gain/loss values reflect more of a time series analysis of streamflow changes than gains or losses of flow along a reach at a specified time. The results of one measurement on February 27, 1995, changed significantly—from indicating a small gain (−1.4 ft^3/s (0.04 m^3/s)) to indicating a large gain (−32.3 ft^3/s

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Table 1 Streamflow measurements and river-leakage values, Souhegan River, Milford, NH, USA

Upstream reach														
Date	Station WLR-1				Station P-2				Measured differences			Rating differences		
	Measured streamflow (cfs)	Estimated accuracy (%)	Streamflow from stage-discharge rating (cfs)	Calculated rating error (%)	Measured streamflow (cfs)	Estimated accuracy (%)	Streamflow from stage-discharge rating (cfs)	Calculated rating error (%)	River leakage (cfs)	Percent of streamflow	Measured river leakage exceeded accuracy of measurement	River leakage (cfs)	Percent of streamflow	Rated river leakage exceeded accuracy of measurement
4/12/1994	569.0	8	569.00	0	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
4/14/1994	704.0	5	689.22	−2.1	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
6/16/1994	42.2	5	42.32	0.4	40.8	5	– –	– –	1.3	3.2	No	– –	– –	– –
7/20/1994	15.4	5	15.30	−0.5	15.9	5	14.25	11.58	−0.5	−3.3	No	1.05	7.13	No
8/24/1994	139.0	5	138.96	0	142.7	5	127.10	12.25	−3.7	−2.6	No	11.86	8.92	No
9/29/1994	236.4	5	232.16	−1.8	223.9	5	218.50	2.49	12.5	5.4	No	13.66	6.06	No
10/11/1994	88.3	5	83.71	−5.2	65.0	5	68.07	−4.52	23.3	30.4	Yes	15.64	20.61	Yes
11/22/1994	131.0	5	130.35	−0.5	116.9	8	116.80	0.07	14.1	11.4	Yes	13.55	10.96	Yes
12/19/1994	164.1	8	187.50	13.3	156.3	5	150.50	3.88	7.8	4.9	No	37.00	21.89	Yes
1/26/1995	297.0	5	282.12	−5	285.3	5	284.70	0.22	11.6	4.0	No	−2.58	−0.91	No
2/27/1995	133.0	5	198.05	39.3	134.3	8	230.30	−41.67	−1.4	−1.0	No	−32.25	−15.06	Yes
3/27/1995	196.4	5	185.98	−5.3	189.2	5	184.90	2.30	7.2	3.8	No	1.08	0.58	No
4/24/1995	140.5	5	136.86	−2.6	147.1	5	141.30	4.10	−6.6	−4.6	No	−4.44	−3.20	No
5/22/1995	129.3	5	128.22	−0.8	121.2	5	116.80	3.73	8.1	6.5	No	11.42	9.32	No
6/28/1995	17.8	5	17.52	−1.7	19.5	5	23.31	−16.19	−1.7	−9.2	No	−5.79	−28.34	Yes
8/3/1995	23.4	5	23.07	−1.3	22.5	5	30.56	−26.49	0.9	4.0	No	−7.49	−27.91	Yes
9/14/1995	5.1	8	5.90	14.6	4.6	8	5.05	−9.31	0.5	10.6	Yes	0.85	15.53	Yes
12/5/1995	171.0	5	165.60	−3.2	168.0	5	160.00	5.00	3.0	1.8	No	5.60	3.44	No
5/30/1996	131.0	5	134.00	2.3	134.0	5	139.50	−3.94	−3.0	−2.3	No	−5.50	−4.02	No
7/10/1996	27.0	5	27.60	2.1	25.9	5	26.00	−0.38	1.1	4.2	No	1.60	5.97	No
8/28/1996	10.9	5	11.00	0.9	10.6	5	10.87	−2.48	0.3	2.8	No	0.13	1.19	No
Downstream reach														
Date	Station WLR-5				Station #39				Measured differences			Corrected differences		
	Measured streamflow (cfs)	Estimated accuracy (%)	Streamflow from stage discharge rating (cfs)	Calculated rating error (%)	Measured streamflow (cfs)	Estimated accuracy (%)	Streamflow from stage-discharge rating (cfs)	Calculated rating error (%)	River leakage (cfs)	Percent of streamflow	Measured river leakage exceeded accuracy of measurement	River leakage (cfs)	Percent of streamflow	Measured river leakage exceeded accuracy of measurement
4/12/1994	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
4/14/1994	847.0	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
6/16/1994	46.7	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
7/20/1994	18.5	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
8/24/1994	150.5	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
9/29/1994	233.8	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
10/11/1994	77.1	5	– –	– –	71.2	5	– –	– –	5.9	8.0	No	7.5	10.1	Yes
11/22/1994	134.5	8	– –	– –	151.1	5	– –	– –	−16.6	−11.6	No	−13.7	−9.6	No
12/19/1994	188.9	5	– –	– –	196.1	5	– –	– –	−7.2	−3.7	No	−2.7	−1.4	No
1/26/1995	336.4	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
2/27/1995	160.7	8	– –	– –	168.8	8	– –	– –	−8.1	−4.9	No	−4.1	−2.5	No
3/27/1995	226.2	5	– –	– –	234.1	5	– –	– –	−7.9	−3.4	No	−5.2	−2.3	No
4/24/1995	166.5	5	– –	– –	177.8	5	– –	– –	−11.3	−6.5	No	−9.0	−5.2	No
5/22/1995	139.5	5	– –	– –	160.3	5	– –	– –	−20.8	−13.9	Yes	−18.7	−12.5	Yes
6/28/1995	21.5	5	– –	– –	23.0	5	– –	– –	−1.5	−6.7	No	−1.4	−6.3	No
8/3/1995	25.5	5	– –	– –	26.1	5	– –	– –	−0.6	−2.3	No	−0.5	−1.9	No
9/14/1995	6.7	8	– –	– –	6.9	5	– –	– –	−0.1	−2.1	No	0.1	0.9	No
12/5/1995	190.0	5	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
1/18/1996	161.0	8	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –	– –
5/30/1996	136.0	5	– –	– –	131.0	5	– –	– –	5.0	3.7	No	7.7	5.8	No
7/10/1996	30.0	5	– –	– –	29.8	5	– –	– –	0.2	0.7	No	0.4	1.5	No
8/28/1996	12.3	5	– –	– –	12.4	5	– –	– –	−0.1	−0.8	No	0.1	0.8	No

cfs, cubic feet per second; – –, no data; negative river leakage means gain; to convert cfs to cubic meters per second, multiply by 0.02832

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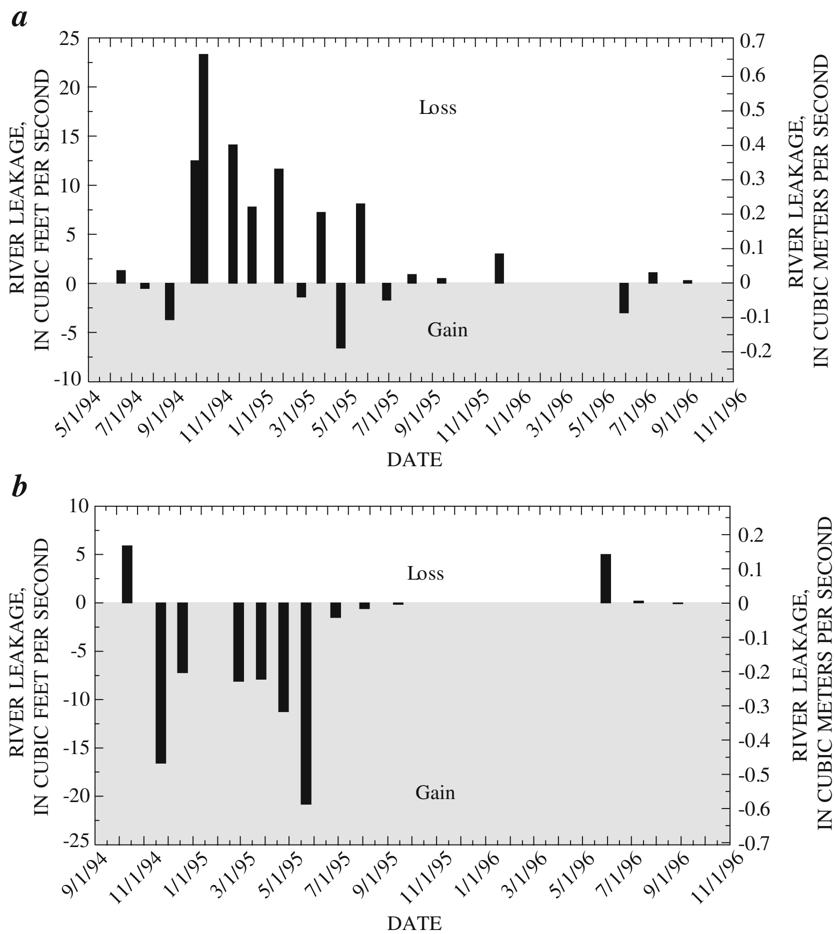

Fig. 3 River leakage for the upper (**a**) and lower (**b**) reaches

(0.91 m^3/s)) after applying a stage-discharge correction (Table 1)—but generally the differences were small. For the lower reach, correcting for unmeasured flow on a small tributary (site 13; Fig. 2) entering the lower reach, by use of flow correlations between measurements made during previous periods at the tributary and lower reach, results in small modifications to river leakage (Tables 1 and 2). Given the small differences between uncorrected and corrected river-leakage values, the original uncorrected river-leakage values were used in subsequent analysis for both the upper and lower reaches. Additionally, the small consistent withdrawals at PFH well (0.34 ft^3/s (0.01 m^3/s)) adjacent to the lower reach offset any uncorrected gains from the small tributary at site 13 (Fig. 2).

Accuracy of river-leakage rates is an issue for most measurement sets. Only three computed river-leakage rates from the upper reach exceeded the 10% error criteria

Table 2 Summary statistics for river-leakage values

Reach	Data set	Mean	Median	Number of readings	Standard deviation	Maximum loss	Maximum gain	
								Number of losses
Upstream	All measured river-leakage values	3.94	1.10	19	7.46	23.31	−6.58	13
	All rating estimated river-leakage values	3.08	1.07	18	13.95	37.00	−32.25	12
								Number of gains
Downstream	All measured river-leakage values	−3.94	−0.36	16	7.43	5.90	−20.82	10
	All corrected river-leakage values	−2.47	−0.24	16	6.79	7.73	−18.71	8

all units in cubic feet per second; negative river leakage means gain; to convert cfs to cubic meters per second, multiply by 0.02832

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for accuracy in streamflow (percent difference, Table 1). Thus 16 of the 19 coupled measurements for the upper reach were less than the precision or accuracy of the coupled measurements. However of the three measurements exceeding the error criteria, two measurements had the largest computed river-leakage values indicating large streamflow loss. Only two computed river-leakage rates from the lower reach exceeded 10% of streamflow (percent difference, Table 1) but these measurements also had the largest computed river-leakage values indicating streamflow gains.

The mean rate of river leakage for the upper reach was 3.94 ft^3/s (0.111 m^3/s); the positive river-leakage value indicated loss of streamflow, or potential aquifer recharge. The mean rate of river leakage for the lower reach was –4.85 ft^3/s (–0.137 m^3/s); the negative river-leakage value indicated gain of streamflow, or potential aquifer discharge. The variability in leakage was large; standard deviation was approximately 7.5 ft^3/s (0.212 m^3/s) for both reaches.

Runoff during the day of measurement appears to have a small effect on river leakage. As an indicator of runoff effects on computed river leakage, river leakage was compared to daily precipitation amounts recorded at a nearby (less than 3 miles (4.8 km) climatic station). For the upper reach, no large gains in flow were associated with appreciable precipitation (exceeding 0.3 inches (in.) (0.76 cm)) events, which might be expected if overland runoff was contributing to streamflow. Of the 3 days with more than 0.3 in. of rainfall, the upper reach showed no appreciable loss or gain in flow for two of the days and a large loss in flow of approximately 14 ft^3/s (0.396 m^3/s) for 1 day (11/22/1994, Table 1; Fig. 4a). For the lower reach, only one large gain was measured with precipitation amounts exceeding 0.3 in. (11/22/1994, Table 1; Fig. 4b).

The magnitude of streamflow showed no apparent effect on river leakage (Fig. 5). River leakage for the upper reach (Fig. 5a) and lower reach (Fig. 5b) showed a poor visual relation with streamflow amount. Although the tendency of a reach to lose or gain flow is apparently unaffected by the magnitude of flow, the magnitude of flow that a reach loses is nevertheless partly affected by the limits of total flow through a reach. This is an obvious, but important point, because the potential amount of available loss will increase with increasing streamflow. The smallest river-leakage values were recorded for streamflows less than 50 ft^3/s (1.32 m^3/s) (Fig. 5a and

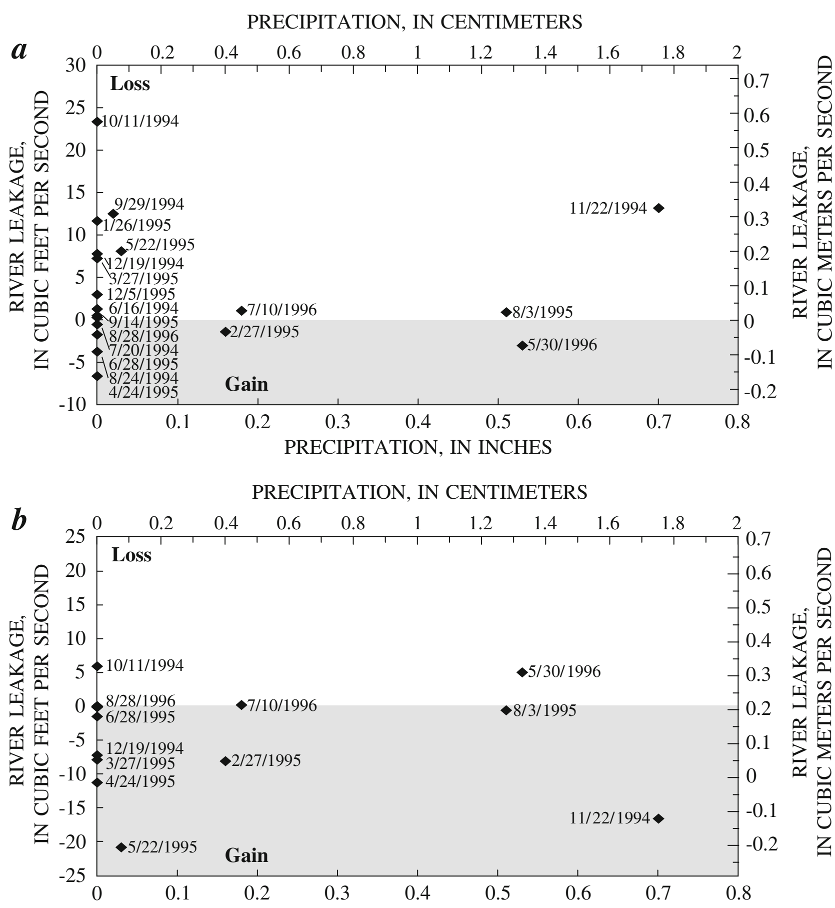

Fig. 4 Comparison of river leakage and daily precipitation amounts (identified by date of measurement) for the upper (**a**) and lower (**b**) reaches

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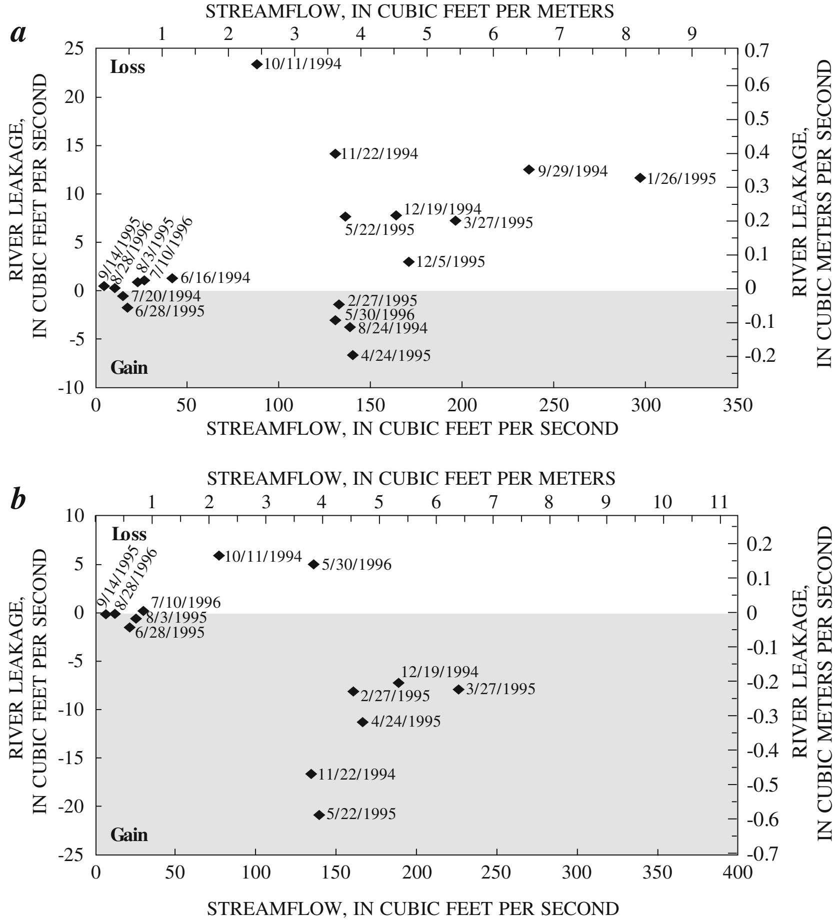

Fig. 5 Comparison of river leakage and streamflow for the upper (**a**) and lower (**b**) reaches

b). While the maximum potential loss occurs during the highest streamflow, the relation between the actual amount of loss and available streamflow is nonlinear. A good example of this is indicated by analysis of data for the upper reach for October 11, 1994, which had the highest streamflow loss (23.3 ft^3/s (0.66 m^3/s)) and also the highest percent of loss (30.4%) but not the highest total streamflow; the latter occurred on January 26, 1995, with a measured flow of 297 ft^3/s (8.41 m^3/s) at WLR1 and a loss of 11.6 ft^3/s (3.28 m^3/s), which represented 4% of total flow.

Streamflow duration as percent exceedance is a good indicator of general hydrologic conditions (wet or dry climatic periods). Streamflow durations for the days when coupled measurements were made on the Souhegan River were derived from durations computed from long-term streamflow measurement stations at nearby locations (Fig. 1). These stations had historical record of more than 50 years and their flow durations were computed from streamflow measurements made over a much longer period than that for the data available from the Souhegan River (less than 3 years). Figure 6 shows that the largest losses for the upper reach were measured for flow duration conditions less than 70% (higher flows). Conversely, the largest gains for the lower reach were measured for the same flow duration conditions.

Variability in river-leakage values decrease with increasing flow duration. River leakage for the upper reach shows that virtually no gains were measured when flow duration exceedance values were greater (lower flow) than 70% (Fig. 6), which suggests that for the higher flows and lower exceedance values, measured gains may reflect processes such as contribution of overland runoff and interflow to the reach.

Water temperatures correlate poorly (correlation coefficient of 0.13) with river leakage at the upper reach, suggesting that changes in water density and (or) the hydraulic conductivity of the streambed material are not critically important to the connectivity of surface water and groundwater systems in this environment. This may be attributable to the relatively coarse-grained sediments that are predominant in the streambed in the study reaches, as well as the effect of other factors on river leakage.

Evapotranspiration (ET) effects on river leakage were negligible (less than 0.02 ft^3/s) based on calculation of potential ET from methods described by Thornwhaite (1948) using mean air temperature. Furthermore, no diurnal fluctuations in water levels from wells adjacent to river reaches were observed indicating groundwater ET was negligible.

Surface- and groundwater horizontal hydraulic gradients from one location along the upper reach (P-2 river staff and

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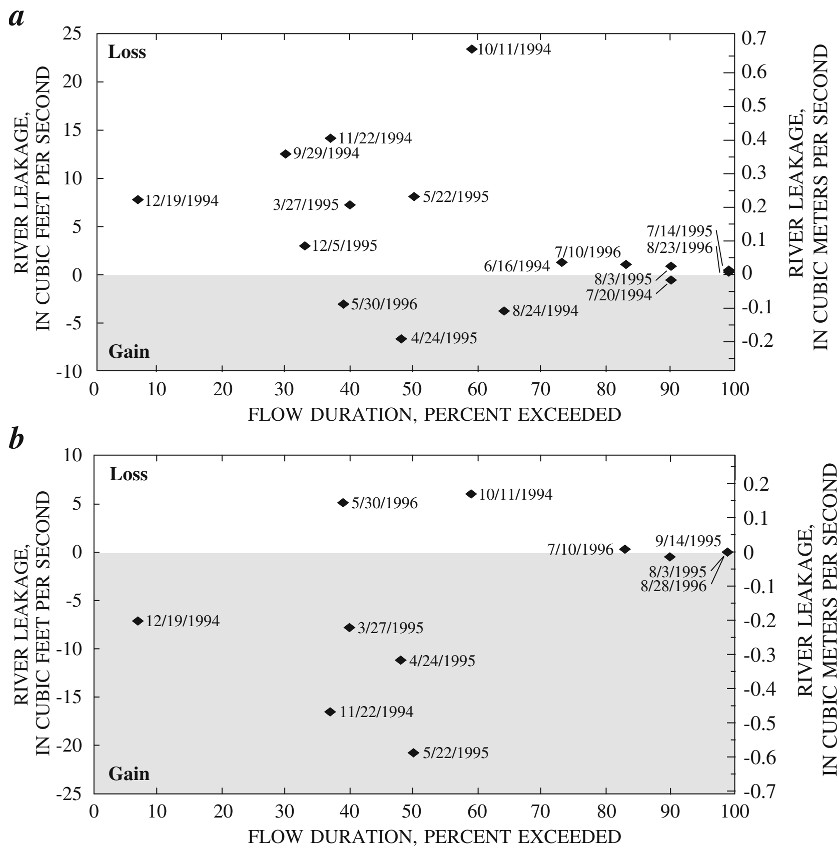

Fig. 6 Comparison of river leakage to streamflow duration for the upper (**a**) and lower (**b**) reaches

adjacent well; Fig. 2) and one location along the lower reach (WLR-5 river staff and adjacent well SPZ-2; Fig. 2) are generally poor indicators (correlation coefficient of 0.07 for the upper reach and 0.05 for the lower reach) of river leakage for their respective reaches (Fig. 7a, b). A possible reason for the poor relation is the scale differences between the two sets of measurements. Because the gradient measurements reflect a vertical slice at one location of the reach; it is a local measurement of gradients, whereas the measured river leakage integrates the bulk interaction along the entire reach. While many of the gradient measurements correlate poorly to river leakage, several gradient measurements do linearly correlate to river leakage as shown by gradient measurements that are in close proximity to the idealized gradient-leakage line (Fig. 7a, b). For the upstream reach, seven out of 19 measurements are within or near the 95% confidence interval for the idealized line. For the downstream reach, seven out of nine measurements are within or near the 95% confidence interval for the idealized line. For the upper reach, most (four of the five) low-flow measurements poorly correlate possibly due to streamflow measurement inaccuracies and small (less than 5) percent difference of river leakage to streamflow.

Temporal variation in gradients including periods when no river leakage was measured indicates a predominantly losing reach for the upper reach (Fig. 8). However, several (four) measurements show gradient reversals where groundwater levels exceed river stage and occur during fall-winter season.

An apparent factor in river leakage appears to be the rise or decline in river stage (Figs. 9 and 10) during a preceding period to the measurements. River stages respond quicker than aquifer groundwater levels to changes in hydrologic conditions. Therefore, for any given event, changes in aquifer heads will lag behind changes in river stages. The relative rate of response of each system appears to be an important control on measured river leakages. At the upper reach, gains in flow were measured primarily during periods of declining river stage as shown by a 1-day change in flow ((Fig. 9a) and a 3-day change in flow (Fig. 9b). Conversely, at the lower reach, streamflow loss was measured only during periods of rising river stage (Fig. 10a and b).

The conceptual model offered to explain river leakage is that during a declining river stage along a predominantly losing reach, the river stage may decline below aquifer heads along part of the reach and cause temporary flow reversals. Conversely, at a gaining reach during a rapid rise in river stage, the river stage may exceed aquifer heads along some part of the reach and streamflow may move into bank storage. The field results indicate that temporary bank storage occurs along both reaches. For both reaches, interflow above the water table could also contribute to streamflow gains and losses, and river-leakage rates that indicate streamflow loss may not be indicative of aquifer recharge rates.

Continuous stage and groundwater level measurements collected during the spring season at P-2 river staff and the

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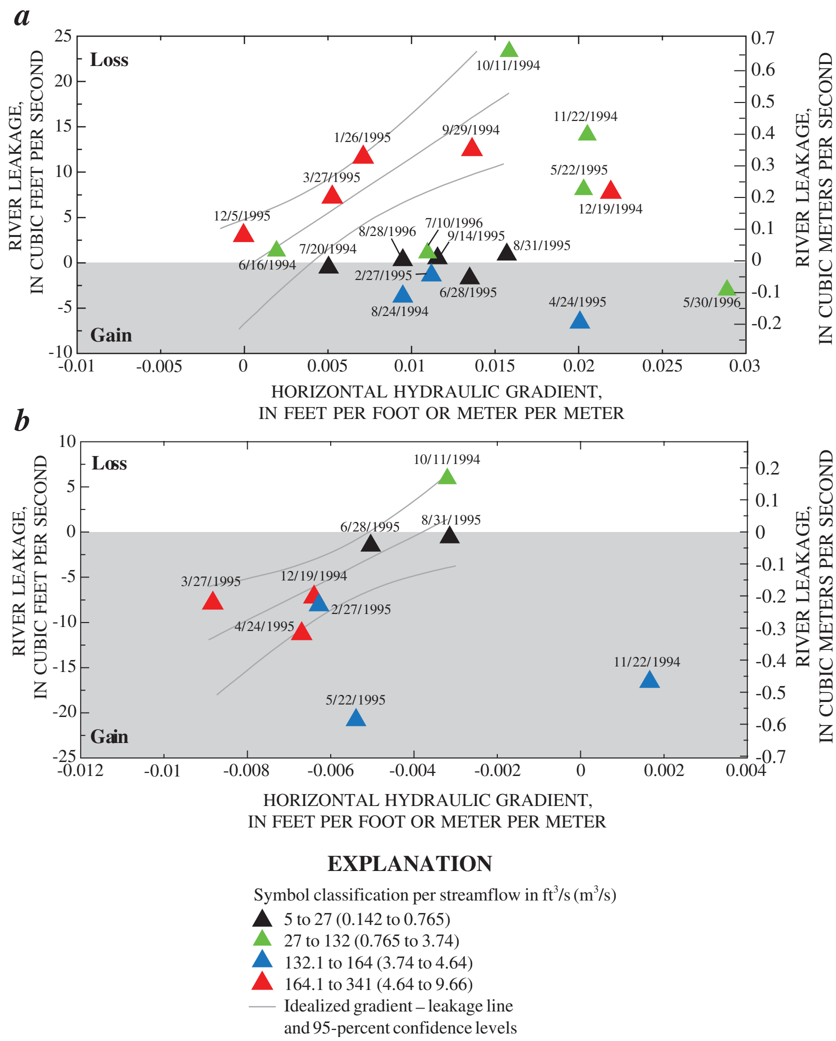

Fig. 7 Comparison of river leakage and horizontal hydraulic gradient between river stage and aquifer head for the upper reach (**a**) and lower reach (**b**)

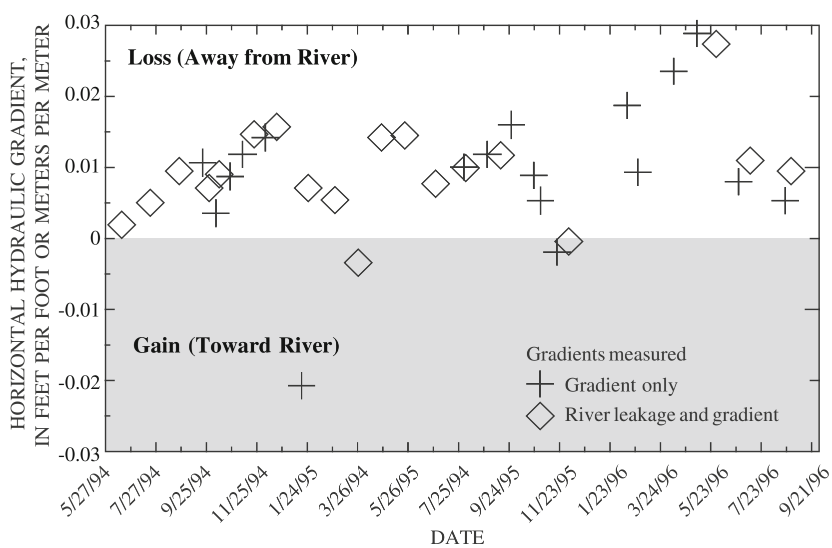

Fig. 8 Horizontal hydraulic gradient between river stage and aquifer head over time for the upper reach

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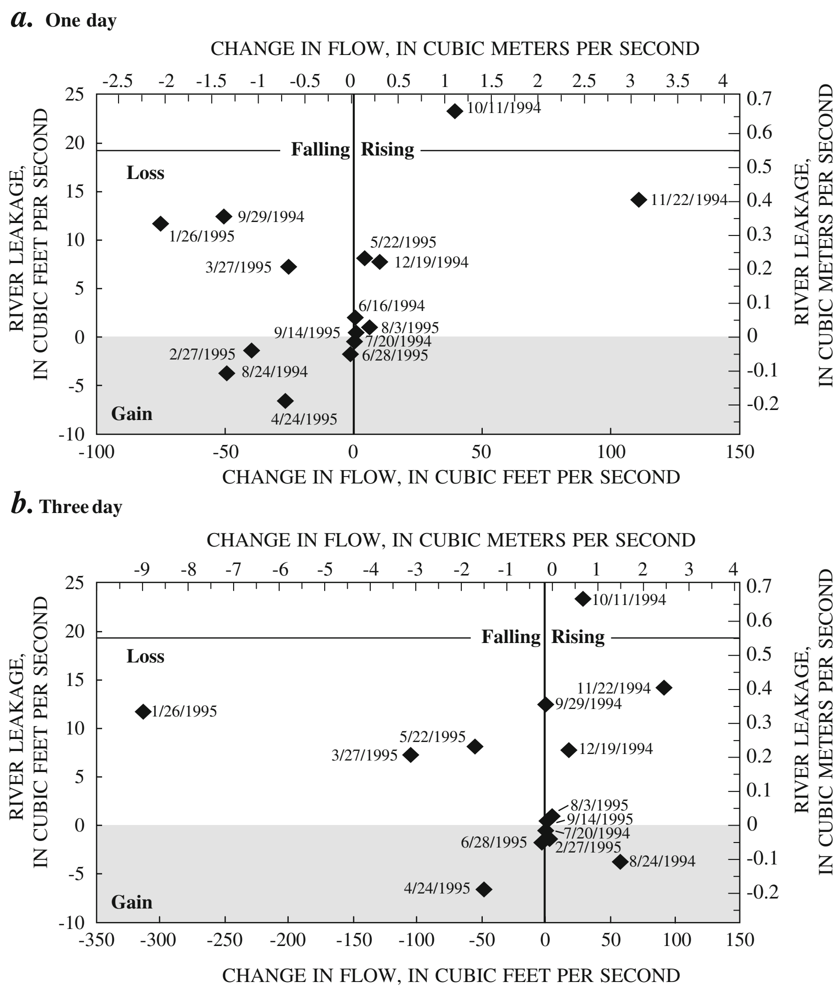

Fig. 9 Comparison of river leakage and change in streamflow for 1-day (**a**) and 3-day (**b**) periods for the upper reach

adjacent well along the upper reach (Fig. 2) indicate that the head difference between stage and groundwater level initially increases during the early rise of river stage (Fig. 11) from a storm event (March 21 to March 23). After the early rise and during the recession limb of the river stage fluctuation (March 24–26), the head difference between stage and groundwater level decreases to near zero. This corroborates manual measurements of gradient (Fig. 8) that indicate gradient reversals can occur and supports the hypothesis that declining limbs may experience leakage reversals.

Statistical summaries of river leakage were compiled for measurements in which both upstream and downstream measurements were made and where antecedent conditions in river stage were measured (Table 3). The summaries are grouped by antecedent river stage trends (1- and 3-day trends in river stage) and seasons (summer and non-summer).

At the upper reach, river leakages measured after an increase in streamflow from the preceding day had a mean loss of 8.68 ft^3/s (0.246 m^3/s) with a standard deviation of 7.93 ft^3/s (0.225 m^3/s) (Table 3). River leakages measured after a decrease in streamflow from the preceding day had a mean loss of only 1.59 ft^3/s (0.045 m^3/s) and a standard deviation of 6.03 ft^3/s (0.17 m^3/s). Although not statistically significant at the 95% confidence interval, the mean difference between leakage measurements during rising and falling stage (1-day trend) varies by a factor of five.

Superimposed on the short-term fluctuations of streamflow are apparent seasonal trends in river leakage. At the upper reach, river leakage showed a slight gain during the summer (mean leakage of −0.53 ft^3/s (0.015 m^3/s); Table 3); however, measurements made in the fall, winter, and spring (non-summer season) showed losses (mean leakage of 8.52 ft^3/s (0.24 m^3/s). At the lower reach, river leakage showed a slight gain during the summer (mean leakage of −0.74 ft^3/s (0.021 m^3/s); Table 3) and large gains during the other seasons. The difference in mean river leakage between summer and non-summer seasons is statistically significant at the 95% confidence interval.

Summary and conclusions

Measured river leakage, the differential between coupled upstream and downstream discrete discharge measurements, varied by approximately 30 ft^3/s (0.85 m^3/s)) at an upper, previously identified losing flow reach and by 26 ft^3/s

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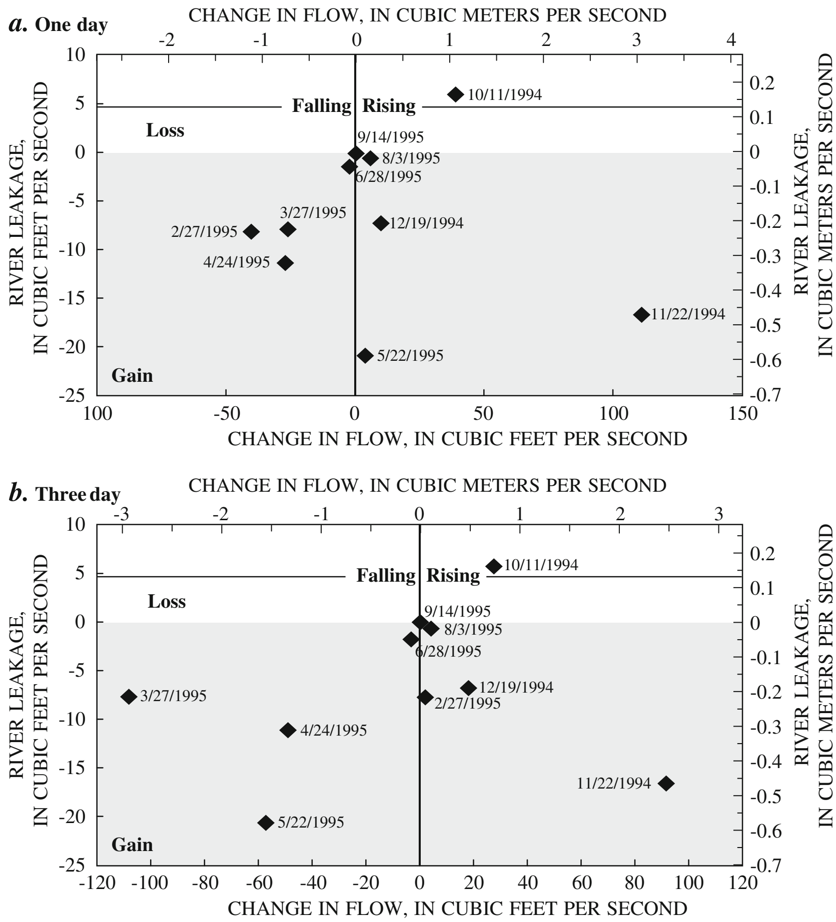

Fig. 10 Comparison of river leakage and change in streamflow for 1-day (**a**) and 3-day (**b**) periods for the lower reach

(0.736 m^3/s) at a lower, previously identified variable flow reach. Both reaches show variability in river leakage between gains (potential aquifer discharge) and losses (potential aquifer recharge) in flow, but results also show a general tendency of a reach toward being either a predominantly losing or gaining reach. At the upper reach,

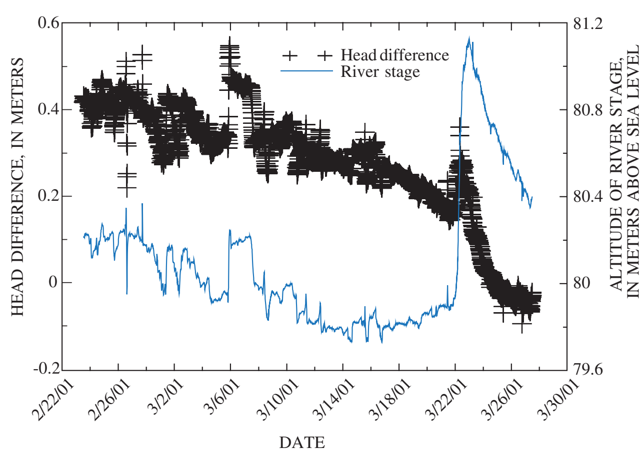

Fig. 11 Horizontal head difference between river stage and aquifer head and river stage altitude for the upper reach during a spring period

13 out of 19 coupled measurements indicated losses in flow, suggesting that the reach is primarily a losing reach. At the lower variable reach, ten out of 13 coupled measurements indicated flow gains, suggesting that the reach is primarily a gaining reach.

A significant factor in river leakage appears to be the rise or decline in antecedent trends in river stage. River stage responds quicker than adjacent aquifer groundwater levels to changes in hydrologic conditions. Therefore, for any given event, changes in aquifer heads will lag behind changes in river stages. The relative rate of response of each system appears to be an important control on measured river leakages. At the upper reach, gains were measured only during periods of declining river stage. Conversely, at the lower reach, streamflow loss was measured primarily during periods of rising river stage.

The final conceptual model offered is that during a period of declining river stage along a predominantly losing reach, the river stage may decline below aquifer heads along part of the reach and cause temporary and local flow reversals. Along a gaining river reach during a rapid rise in river stage, the river stage may exceed aquifer heads along some part of the reach and streamwater may move into bank storage. Bank storage occurs along both reaches and is an important process in understanding

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Table 3 Summary statistics for river-leakage values for selected periods

Antecedent or seasonal conditions	Upstream reach			Downstream reach		
	Number of measurements	Mean river leakage (cfs)	Standard deviation of river leakage (cfs)	Number of measurements	Mean river leakage (cfs)	Standard deviation of river leakage (cfs)
One-day rise in stage	7	8.68	7.93	– –	– –	– –
One-day fall in stage	8	1.59	6.03	– –	– –	– –
Three-day rise	7	7.06	8.88	– –	– –	– –
Three-day fall	8	3.01	6.16	– –	– –	– –
Summer season	6	−0.53	1.90	3	−0.74	0.69
Non-summer season	9	8.52	8.68	7	−9.42	8.48

cfs, cubic feet per second; – –, no data; negative river leakage means gain; to convert cfs to cubic meters per second, multiply by 0.02832

grains and losses. Multiple river-leakage measurements made during a single storm event would help validate this proposed conceptual model.

Collecting river-leakage information only during summer or low-flow periods may potentially underestimate rates of aquifer recharge and discharge. River leakage during the non-summer seasons showed a mean streamflow loss of 8.52 ft^3/s (0.24 m^3/s) at the upper reach. During the summer, streamflow loss was negligible along this reach. Seasonal trends in the relation of water-table surface to river stage may play a role in inducing additional flow loss in the fall because the water table is often lowest at this time in the study area.

Although some tendencies exist for the investigated reaches, several factors that can complicate the analysis of river leakage and streamflow gains and losses must be recognized. The first factor is inaccuracies in the computed stream discharge and thus in computations of river leakage. The computed net gain or loss between two coupled discrete discharge measurement stations is subject to a potential cumulative error of 10% for two measurements rated “good” (5% each). Therefore, confidence that computed gains/losses are realistic and not a function of measurement inaccuracies must be partly dependent on whether computed gains/losses exceed the 10% criteria for “good” measurements. Most river-leakage values were less than the qualitative accuracy assigned to the set of measurements. This reinforces the need to obtain multiple coupled measurements so that more reliable values of river leakage can be derived from statistical averages of the measurements rather than a single set of coupled measurements. The reliability of streamflow measurements in the quantification of river leakage and identification of patterns of gains and losses along reaches should also be validated against alternative methods including measurements of surface—and groundwater gradients. The second factor is time-dependent, river-aquifer responses. Different river-aquifer processes occur at different periods of the year due to climatic conditions and cause a non-linear system response. For example, some measurements were made during dry, low-flow conditions when groundwater recharge or discharge is the primary factor in gains or losses. During high flow or intermediate flow periods, processes such as overland runoff, interflow, and bank storage may affect streamflow gains or losses. Another time-dependent response to be considered is the seasonal position of the water table and the relative responses of river-stage and aquifer-head fluctuations that play an important role in river leakage.

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References

Atkins JB, Journey CA, Clarke JS (1996) Estimation of groundwater discharge to streams in the central Savannah River Basin of Georgia and South Carolina. US Geol Surv Water-Resour Invest Report 96–4179:35

Barlow PM, Moench AF (1998) Analytical solutions and computer programs for hydraulic interaction of stream-aquifer systems. US Geological Survey Open-File Report 98–415A:85

Batelaan O, De Smedt F, Triest L (2003) Regional groundwater discharge: phreatophyte mapping, groundwater modelling and impact analysis of land-use change. J Hydrol 275:86–108

Brayton MJ (2001) Combined use of streambed piezometers and water temperature measurements to characterize gaining and losing stream reaches. Unpublished Masters Research paper, University of New Hampshire, variously paginated

Brayton MJ, Harte PT (2001) Results of a monitoring program of continuous water levels, specific conductance, and water temperature at the OK Tool facility of the Savage Municipal Well Superfund site, Milford, New Hampshire. US Geological Survey Open-File Report 01–338:50

Callegary JB, Leenhouts JM, Paretti NV, Jones CA (2007) Rapid estimation of recharge potential in ephemeral stream channels using electromagnetic methods, and measurements of channel and vegetation characteristics. J Hydrol 344:17–31

Chen XH, Shu L (2006) Groundwater evapotranspiration captured by seasonally pumping wells in river valleys. J Hydrol 318:334–347

Constantz JE, Stewart AE, Niswonger RG (1999) The influence if water temperature on stream/groundwater exchanges: paper presented at Geological Society of America, 1999 annual meeting, Denver, CO

Dumouchelle DH (2001) Evaluation of ground-water/surface-water relations, Chapman Creek, West-Central Ohio by means of multiple methods. US Geol Surv Water-Resour Invest Report 01–4202:18

 DOI 10.1007/s10040-008-0359-1

Dunne T, Leopold LB (1978) Water in environmental planning. W.H. Freeman and Co, New York, p 818

Dysart JE, Rheaume SJ, Kontis AL (1999) Induced infiltration from the Rockaway River and water chemistry in a stratified-drift aquifer at Dover, New Jersey, with a section on modeling groundwater flow in the Rockaway River Valley. US Geol Surv Water-Resour Invest Report 96–4068:112

Fryar AE, Wallin EJ, Brown DL (2000) Spatial and temporal variability in seepage between a contaminated aquifer and tributaries to the Ohio River. Ground Water Monit Remediat 2–03:129–146

Harte PT, Mack TJ (1992) Geohydrology of, and Simulation of Groundwater Flow in the Milford-Souhegan Glacial-Drift Aquifer, Milford, New Hampshire. US Geological Survey Water Resources Investigations Report 91–4177:90

Harte PT, Flynn RJ, Kiah RG, Severance T, Coakley MF (1997) Information on hydrologic and physical properties of water to assess transient hydrology of the Milford-Souhegan Glacial-Drift aquifer, Milford, New Hampshire. US Geol Surv Open-File Report 97–414:96

Harte PT, Flynn RJ, Mack TJ (1999) Construction and calibration of numerical groundwater flow models of the Milford-Souhegan Glacial Drift aquifer, Milford, New Hampshire. US Geol Surv Open-File Report 99–462:76

HMM Associates Inc (1991) Remedial investigation, Savage well site, Milford, New Hampshire. Concord, Mass., no. 2176 HAZ/4814:800

Kontis AL, Randall AD, Mazzaferro DL (2004) Regional hydrology and simulation of flow of stratified-drift aquifers in the glaciated northeastern United States. US Geol Surv Water-Supp Paper 1415-C:156

Lee KK, Risely JC (2002) Estimates of ground-water recharge, baseflow, and stream reach gains and losses in the Willamette River Basin, Oregon. US Geol Surv Water-Resour Invest Report 01–4215:53

Meyboom P (1964) Three observations on streamflow depletion by phreatophytes. J Hydrol 2:248–261

Rantz SE (1982a) Measurement and computation of streamflow-methods of measurement. US Geological Survey Water-Supply Paper 2175–1:1–284

Rantz SE (1982b) Measurement and computation of streamflow-computation of discharge. US Geol Surv Water-Supp Paper 2175–2:285–631

Rosenberry DO, Winter TC (1997) Dynamics of water-table fluctuations in an upland between two prairie-pothole wetlands in North Dakota. J Hydrol 191:266–289

Rushton KR, Tomlinson LM (1979) Possible mechanisms for leakage between aquifers and rivers. J Hydrol 40:49–65

Rutledge AT (1993) Computer programs for describing the recession of groundwater discharge and for estimating mean groundwater recharge and discharge from streamflow records. US Geol Surv Water-Resour Invest Report 93–412:45

Sauer VB, Meyer RW (1992) Determination of error in individual discharge measurements. US Geol Surv Open-file Report 92–144:21

Simonds FW, Sinclair KA (2002) Surface water–ground water interactions along the lower Dungeness River and vertical hydraulic conductivity of streambed sediments, Clallam County, Washington, September 1999- July 2001. US Geol Surv Water-Resour Invest Report 02–4161:60

Stevens HH, Ficke JF, Smoot GF (1975) Water temperature-influential factors, field measurement, and data presentation. US Geol Surv Tech Water- Resour Invest D1–1:1–65

Thornthwaite CW (1948) An approach toward a rational classification of climate. Geogr Rev 38:55–94

US Geological Survey (2007) Techniques of Water-Resources Investigations (TWRI). USGS TWRI Book 3, Chapter A1 through A19 and Book 8, Chapters A2 and B2 accessed online at http://pubs.usgs.gov/twri/, accessed 2007

Wilcoxon F (1945) Individual comparisons by ranking methods. Biometrics 1:80–83

Williams-Sether T (2004) River gain and loss studies for the Red River of the North Basin, North Dakota and Minnesota. US Geol Surv Open-File Report 2004–1076:21

Winter TC, Harvey JW, Franke OL, Alley WM (1998) Ground water and surface water a single resource. US Geol Surv Circ 1139:79

 DOI 10.1007/s10040-008-0359-1