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$\delta^{18}O$, δD and ^{3}H measurements constrain groundwater recharge patterns in an upland fractured bedrock aquifer, Vermont, USA

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Abstract

Stable isotope ratios, measured in groundwater samples, reflect the effects of evapotranspiration on bedrock recharge and flow patterns in a 10.5 km^2 upland mountainous watershed in northwestern Vermont. Precipitation and groundwater samples ($n = 619$) were collected weekly over a 1.5 year period. Precipitation $\delta^{18}O$ ranges yearly over ~25 per mil (‰), and decreases 2.5‰ for every 1000 m of elevation gain, reflecting seasonal and altitudinal temperature changes. δD values are well correlated with $\delta^{18}O$ and plot close to the Meteoric Water Line, indicating no evaporation effects. In the colder months (late November to early April), groundwater $\delta^{18}O$ composition varies by as much as 4.3‰ in response to precipitation events, indicating rapid local recharge to the bedrock. In the warmer months (late April to early November), variation in groundwater $\delta^{18}O$ is smaller (within 0.4‰) at the lower elevations in the watershed, reflecting a reduction of infiltration to the bedrock. However, at higher elevations (above 800 m asl), groundwater $\delta^{18}O$ continues to respond to precipitation events due to the sparse vegetative cover and colder temperatures that result in lower evapotranspiration rates. Tritium concentrations in groundwater range from 6.7 to 26.7 TU, indicating that groundwater residence times may vary from less than 1 year to in excess of 30 years.

Keywords: Bedrock; Ground-water; Isotope-geochemistry; Recharge; Stable isotopes; Vermont

1. Introduction

In New England, upland bedrock aquifers are rich sources of clean drinking water. To predict water supply sustainability or to address contamination problems in a particular upland basin, a comprehensive understanding of groundwater recharge and flow is required. Of particular importance are the locations of significant recharge areas, as well as the amount and timing of recharge to bedrock (Maloszewski and Zuber, 1982; Alley, 1984; Harte and Winter, 1993).

The stable isotopic composition of water reflects the temperature to which it was exposed during condensation (Friedman and O'Neil, 1977; Hoefs, 1987). We used stable isotope measurements (oxygen and hydrogen) in precipitation and groundwater to characterize patterns of groundwater recharge and flow in the upland, fractured phyllite and schist aquifer of the Browns River, northwestern Vermont (Fig. 1). Large seasonal temperature changes (>50°C annually) and high relief ($\Delta = 1080$ m) in this setting

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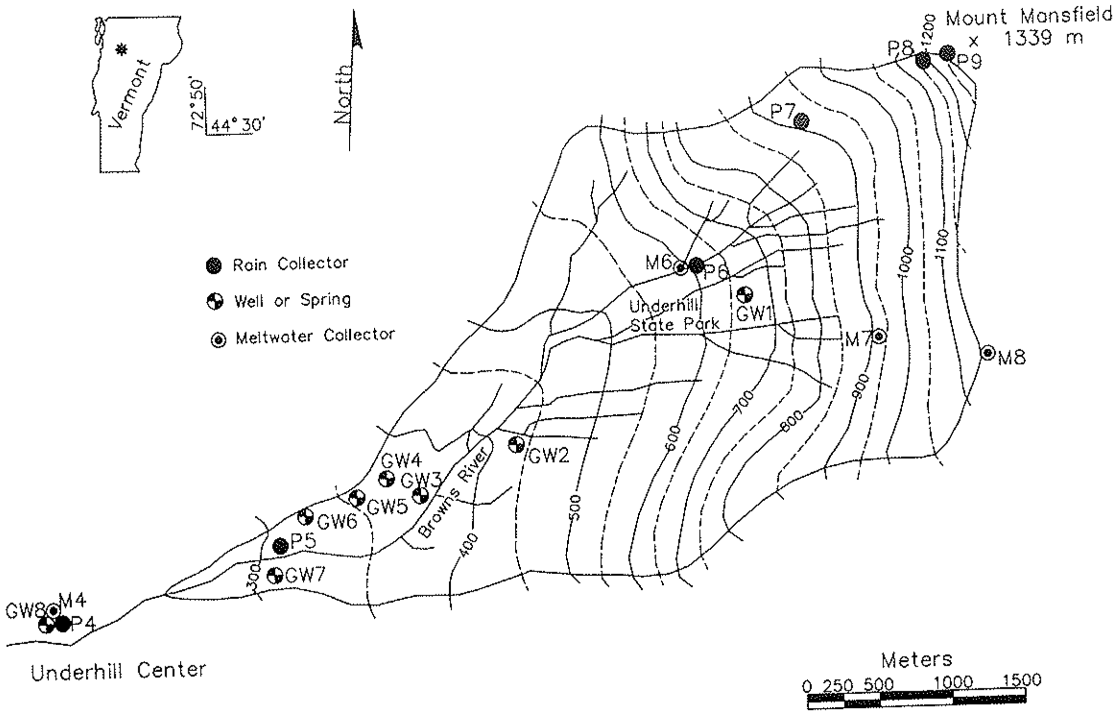

Fig. 1. Study area and sampling locations. Sampling devices were installed at the locations shown for meltwater and rain collection. Sampling location P1 was located outside the study area. Contours are in meters asl.

(δD) ranged from −52 to −155‰. We found that seasonal changes in evapotranspiration rates affect the $\delta^{18}O$ composition of groundwater at some locations. Temporal and spatial patterns in $\delta^{18}O$ and δD were used to estimate the location of significant recharge areas in the basin. Radiogenic isotope (tritium) measurements in groundwater were also used to obtain a preliminary interpretation of relative residence times.

2. Location and geology of the study area

This study was focused on the upland watershed (10.5 km^2) of the Browns River (44°30′N, 72°50′E), bordered on the east by the north–south trending ridge of Mt. Mansfield, the highest of the Green Mountains (Fig. 1). Elevations in this watershed range from

from 250 cm at the highest elevations to 80 cm in the lower portion of the basin (Wilmot and Scherbatskoy, 1994). The dominant tree type changes from a mix of large deciduous (such as oak and maple) and coniferous (such as fir and spruce) trees in the valley, to sparsely distributed and often scrubby conifers on the mountain slopes, and finally to arctic and alpine tundra vegetation at the highest elevations on Mount Mansfield (Burns and Otis, 1916; Connor, 1994). Soils are generally thin sandy to stony loams at the upper elevations. Below 350 m asl, the topography flattens, and more sandy soils are present (USDA, 1974; 1981). The upper portion of the watershed has a thin cover of till or colluvium; at lower elevations (below about 350 m asl), deposits of glaciolacustrine sediments and till of up to 150 m in thickness exist (Stewart, 1961; Connally, 1968; Stewart and MacClintock, 1969).

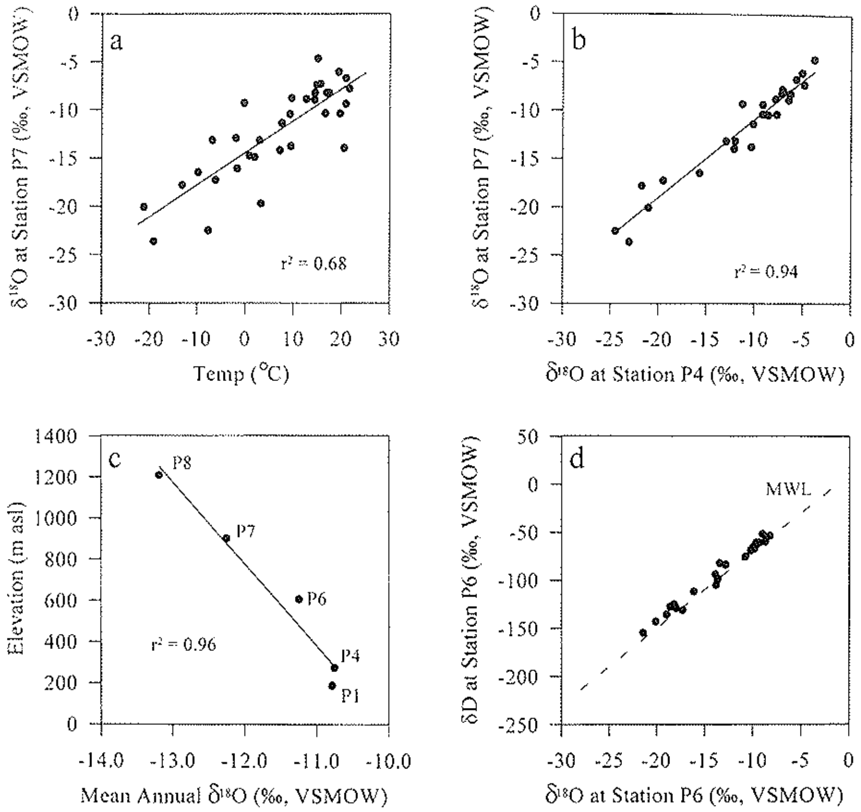

Fig. 3. Precipitation data. (a) Precipitation $\delta^{18}O$ correlation with temperature. $\delta^{18}O$ measurements are from station P7 (900 m asl). Temperature from Proctor Maple Research Center (600 m asl), Underhill (Cummings, 1997, pers. comm.). Equation of linear best fit is: $Y = 0.3X - 14.4$. (b) Precipitation $\delta^{18}O$ correlation between sites. $\delta^{18}O$ measurements are from stations P7 (900 m asl) and P4 (270 m asl). Equation of linear best fit is: $Y = 0.8X - 2.9$. (c) The relationship of mean annual precipitation $\delta^{18}O$ and elevation. Equation of linear best fit is: $Y = -398X - 4003$. (d) Precipitation $\delta^{18}O$ versus δD. $\delta^{18}O$ and δD measurements are from station P6 (600 m asl). Equation of Meteoric Water Line (dashed) is: $\delta D = 8\delta^{18}O + 10$ (Craig, 1961).

enriched and analyzed at the University of Waterloo (Heemskerk et al., 1993).

4. Results of precipitation analysis

Measured $\delta^{18}O$ values in rain and newly fallen snow varied over a range of 25‰ (−3 to −28‰) at most locations during the sampling period (Fig. 2). Positive correlation ($r^2 = 0.47$–0.70) was observed between precipitation $\delta^{18}O$ values and seasonal air

with increasing elevation (−2.5‰ per 1000 m, $r^2 = 0.96$) as a result of the colder temperatures at higher elevations (Fig. 3c). Although minimal fractionation by rainout may occur in the case of a localized cloudburst, neither the spatial density of rain collectors nor the frequency of collection were adequate to quantify any such effects. In this setting, the effects of temperature and elevation dominate any other factors in controlling the $\delta^{18}O$ composition of precipitation.

Deuterium values in precipitation are linearly and positively correlated to $\delta^{18}O$ values (Fig. 3d). The

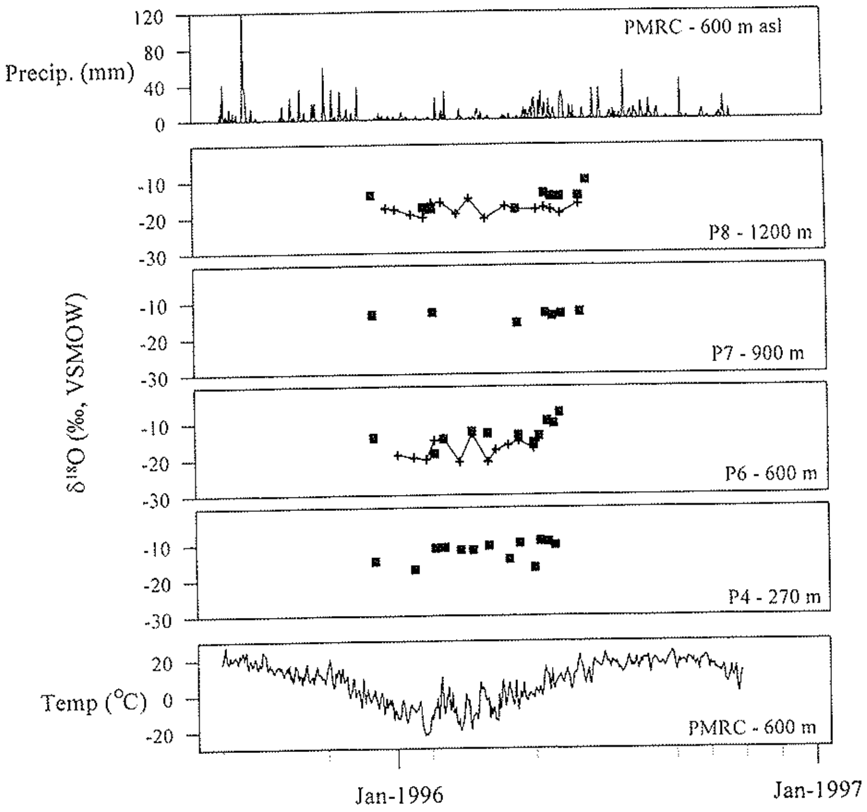

Fig. 4. Meltwater and snowpack $\delta^{18}O$ records. (■) meltwater samples. (+) cumulative snowpack samples. Daily temperature and precipitation records from measurements taken at Proctor Maple Research Center (600 m asl), Underhill (Cummings, 1997, pers. comm.).

The line described by this equation fits closely to the Global Meteoric Water Line (MWL) (Craig, 1961):

$$\delta D = 8 \cdot \delta^{18}O + 10‰, \qquad (2)$$

indicating little or no evaporation prior to or after collection of the water samples.

5. Results of meltwater analysis

The $\delta^{18}O$ composition of meltwater from the snow pack is influenced by the $\delta^{18}O$ composition of the snow pack and by mixing with rain (Fig. 4). Melting events occurred from late November to late May at

probably a result of the relatively depleted melting snow mixing with relatively enriched rainfall (Friedman et al., 1991; Sommerfield et al., 1991; Shanley et al., 1994; IAHS, 1995). During the period of highest melting rates, from late May to early June, the $\delta^{18}O$ value of meltwater increases sharply, reflecting the input of large amounts of less-depleted rainfall. δD was not measured in meltwater samples.

6. Results of groundwater analysis

The $\delta^{18}O$ composition of groundwater varies seasonally in wells and springs. The mean yearly

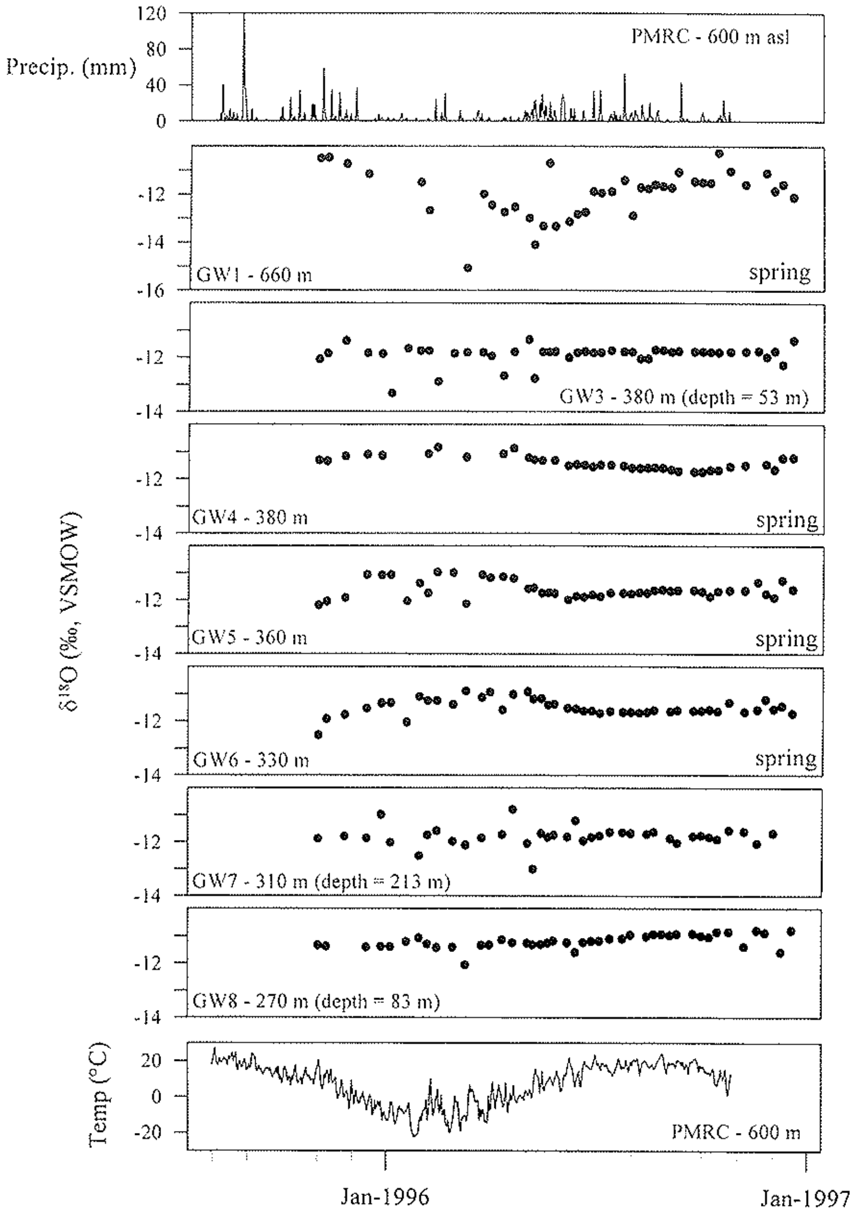

Fig. 5. Groundwater $\delta^{18}O$ records. Complete records were not established for GW2 and GW7b. Daily temperature and precipitation records from measurements taken at Proctor Maple Research Center (600 m asl), Underhill (Cummings, 1997, pers. comm.).

for GW1 mimics but appears to lag the seasonal temperature curve. The least depleted $\delta^{18}O$ values

reflects the rate of groundwater infiltration and flow through the bedrock in the area recharging GW1. The

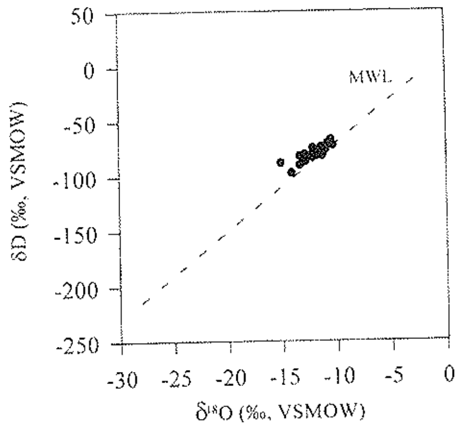

Fig. 6. Groundwater $\delta^{18}O$ versus δD. $\delta^{18}O$ and δD measurements are from wells GW1 (660 m asl), GW3 (380 m asl), and GW5 (360 m asl). Equation of Meteoric Water Line (dashed) is: $\delta D = 8\delta^{18}O + 10$ (Craig, 1961).

without a high degree of mixing prior to reaching GW1. Measured groundwater δD and $\delta^{18}O$ values plot very close to the precipitation values (discussed earlier) and the MWL (Fig. 6), indicating that groundwater has not been modified isotopically during or since recharge.

In most of the wells, with the exception of GW1, the groundwater exhibits a nearly constant $\delta^{18}O$ signature ($1\sigma \leq 0.2$‰) from mid-April to mid-November, indicating a thoroughly mixed groundwater system not subject to $\delta^{18}O$ alteration by individual rain events. More substantial fluctuations ($1\sigma = 0.2$–0.5‰) in the $\delta^{18}O$ composition of groundwater in these wells occur during the colder months (late November to early April). The $\delta^{18}O$ values for GW1 vary over a larger range ($1\sigma = 0.9$‰) throughout the year. However, the variance in $\delta^{18}O$ at GW1 is somewhat lower during the summer.

The arithmetic mean annual $\delta^{18}O$ values differ slightly among the sampling sites, ranging from −12.1 to −11.2‰ (Table 1). The spring located at the highest elevation (GW1) exhibits the most depleted mean annual $\delta^{18}O$ composition and the well lowest in elevation (GW8) exhibits the least depleted mean annual $\delta^{18}O$. However, no consistent relationship ($r^2 < 0.1$) between sampling site elevation and the mean annual $\delta^{18}O$ value was observed. In addition, no significant correlation ($r^2 < 0.1$) was observed between mean annual $\delta^{18}O$ values and the physical characteristics of the wells, such as well depth or yield. Unlike precipitation, the $\delta^{18}O$ values in groundwater do not correlate well with air temperature changes ($r^2 < 0.1$). This is due to the lag in time that occurs between recharge and arrival of groundwater at a well or spring and to mixing that occurs during travel within the bedrock. Groundwater $\delta^{18}O$ records are only slightly correlated between wells ($r^2 \leq 0.3$), again due to variations in groundwater travel time and mixing. Based on these findings, it appears that the $\delta^{18}O$ composition of groundwater in each well responds uniquely to seasonal changes. Furthermore, it appears that the degree of mixing between local recharge and deep groundwater is temporally and spatially variable and dependent upon the nature and interconnection of bedrock fractures at each location.

Table 1
Well characteristics, mean annual $\delta^{18}O$ values, estimated recharge elevations and tritium concentrations for groundwater sampling locations

Sample location	Depth (m)	Sample elevation (m asl)	Mean annual $\delta^{18}O$ (‰)	Estimated mean recharge elevation (m asl)	June 1996 tritium concentration (TU)
GW1	Spring	670	−12.1	815	15.8 (±1.3)
GW2	Spring	440	n/a[a]	n/a	15.9 (±1.3)
GW3	53	380	−11.9	735	26.7 (±1.9)
GW4	Spring	380	−11.4	540	14.6 (±1.2)
GW5	Spring	360	−11.6	605	15.9 (±1.2)
	Spring	330	−11.5	555	15.2 (±1.2)

Tritium concentrations in groundwater may be used as indicators of groundwater age by comparison with historical records of elevated tritium levels in precipitation, which resulted from atmospheric thermonuclear testing in the early 1950s to late 1970s (Schlosser et al., 1988; Busenberg and Plummer, 1993). Tritium values ranged widely among the groundwater sampling locations in this study. All the samples contained tritium concentrations of 6.7 TU or greater, indicating that most of the water was recharged within the last 40–50 years (i.e. since the start of atmospheric thermonuclear testing). However, considering that substantial mixing may occur within the groundwater system, the lowest value measured (6.7 TU at GW7) may indicate recharge prior to the early 1950s of some groundwater, which has since mixed with younger water. The highest value measured (26.7 TU at GW3) may indicate a recharge date of about 1965, based on matching with historical tritium curves and consideration of tritium decay rates (Schlosser et al., 1988). Most of the samples analyzed exhibited tritium concentrations of between 14 and 17 TU, which generally matches the composition of modern precipitation (Table 1).

7. Discussion of recharge

Measurements of the $\delta^{18}O$ composition of groundwater, specifically the stable summer and erratic winter values, suggest that recharge to bedrock from precipitation events occurs in significant amounts only during the colder months when evapotranspiration is very low. Supporting our observations are 50 years (1910–1960) of precipitation, soil moisture, and potential evapotranspiration measurements compiled for Burlington, Vermont (C.W. Thornthwaite Associates, Laboratory of Climatology, 1964). These data indicate that evapotranspiration (E_p) is negligible from the months of December to March. E_p increases in late March or early April to a peak in July, then decreases to a negligible amount in November. Although precipitation is relatively high during the warmer months, soil moisture remains

We believe that the steady $\delta^{18}O$ signature observed in most of the wells and springs from late April to early November is governed by well-mixed groundwater flowing in deeper, interconnected fractures. In the colder months when evapotranspiration does not prevent recharge, infiltration occurring near the wells during each storm or melting event results in alteration of the $\delta^{18}O$ composition of the sampled groundwater (Fig. 5). During the warmer months, groundwater recharge in the valley, where most of the wells are located, is virtually shut off by vegetative extraction of water from the vadose zone in overburden soils. This results in a reduction of the influence of individual rain events on the groundwater $\delta^{18}O$ and δD signatures. Vegetative transpiration and evaporation in this region, especially in the valley forests during the warm months, are capable of transferring up to 70% of the yearly precipitation back to the atmosphere, thereby removing it from the recharge flux (Saxton and McGuinness, 1982). Such effects are important in the behavior of many hydrologic systems (McGuinness and Harold, 1962; Knisel et al., 1969; Parmele, 1972; Woolhiser, 1973). Several studies in the northern forests of the United States have shown that significant infiltration through soil is unlikely to occur during the summer due to high evapotranspiration (Likens et al., 1977; Dugan and Peckenpaugh, 1985; Likens et al., 1985). In contrast to recharge processes in the lower portion of the basin, the watershed near GW1 has a sparse forest of smaller trees and is typically exposed to lower temperatures, allowing recharge to occur throughout the year.

8. Delineation of recharge areas

Assuming that groundwater in a well or a spring is well mixed, the arithmetic mean annual $\delta^{18}O$ value of each groundwater sampling location provides an approximation of the spatially and temporally weighted mean elevation of upland recharge to the well or spring (Table 1). Based on a best-fit linear interpretation of the relationship for mean annual $\delta^{18}O$ of precipitation vs. elevation (Fig. 3c), the

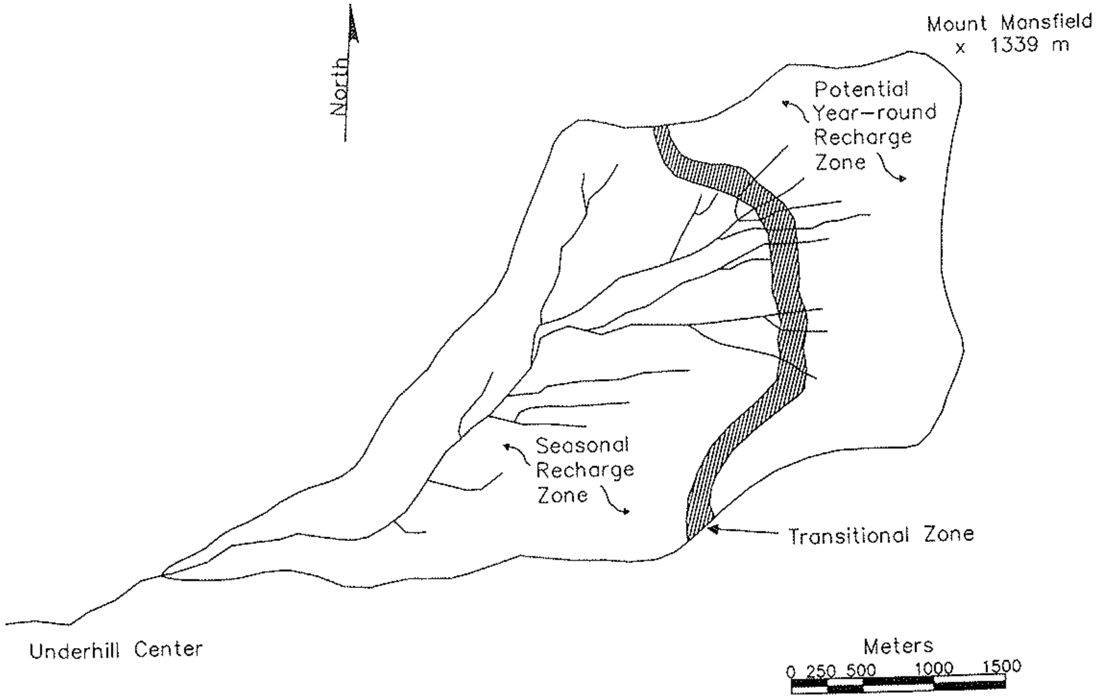

Fig. 7. Delineation of recharge areas. Potential year-round recharge zone (800–1330 m asl), seasonal recharge zone (250–750 m asl), and transitional zone (hatch pattern) (750–800 m asl).

where Y is the spatially and temporally weighted mean recharge elevation (meters asl) and X the mean annual $\delta^{18}O$ in groundwater (‰).

Several factors may complicate the use of a mean annual $\delta^{18}O$ value as an indicator of recharge elevations. Groundwater flowing from the highest portion of the well or spring's recharge area (lowest mean $\delta^{18}O$ values) may become mixed with water recharged at lower elevations (higher mean $\delta^{18}O$ values) along the flowpath. The relative proportions of recharge contributed from different elevations within the recharge area determine the mean $\delta^{18}O$ value recorded at the well. Furthermore, the groundwater recharge appears to be dominated by cold season (lower $\delta^{18}O$) infiltration. Therefore, the above equation may overestimate recharge elevations.

The watershed can be separated into two general

by the interpreted recharge elevations of wells and springs used in the study. The portion of the watershed above 800 m asl is interpreted to be the year-round recharge zone for a well or spring like GW-1, which has an interpreted recharge elevation of 815 m asl. The portion of the watershed below 750 m asl is interpreted to be the seasonal recharge zone. The seasonal recharge zone encompasses the interpreted recharge zones of all the other groundwater sampling locations, which exhibit $\delta^{18}O$ records indicative of seasonal recharge and have interpreted recharge elevations of 450–735 m asl. The portion of the watershed between the two recharge zones (750–800 m asl) is considered to be a transitional zone between the year-round and seasonal recharge zones. A more spatially dense sampling program could identify the precise boundaries of each zone.

overlap zone) is 3.8 km^2, which represents about 36% of the entire basin area. However, this area supplies recharge to the groundwater system at least sporadically on a year-round basis while recharge only occurs in the lower basin for 5 months out of the year. Furthermore, the average annual precipitation in the year-round recharge zone (up to 250 cm per year) is at least twice that of the rest of the basin. Based on these general relationships, the following equations can be developed to describe recharge to groundwater as a function of infiltration rates and the duration of the recharge season at a particular location in the watershed:

$$R_{\text{upper}} \approx (0.36A)\cdot i \tag{4}$$

$$R_{\text{lower}} \approx (0.64A)\cdot(i/2)\cdot(5/12) \tag{5}$$

where R_{upper} is the recharge to bedrock in year-round recharge zone (L^3/T), R_{lower} the recharge to bedrock in seasonal recharge zone (L^3/T), A the total basin area (L^2) and i is the maximum infiltration rate to soils (function of precipitation rates); assume that infiltration in the lower portion of the watershed is about half that in the upper basin, based on relative precipitation rates (L/T).

The percent of total recharge in the basin that occurs in the potential year-round recharge zone can be estimated as follows:

$$\begin{aligned} R_{\text{upper}} &\approx \{(0.36A)\cdot i/[(0.36A)\cdot i \\ &\quad + (0.64A)\cdot(i/2)\cdot(5/12)]\}\cdot 100\% \\ &\approx 73\% \end{aligned} \tag{6}$$

Therefore, the potential year-round recharge zone, although covering only 36% of the total watershed area, may account for 73% of the annual recharge to the bedrock aquifer, suggesting that the upper elevation of this watershed is a significant source of recharge worthy of special protection against any potential contaminating activities.

9. Summary and conclusions

bedrock groundwater system is controlled by the effects of evaporation and plant uptake, which are functions of elevation. Most of the recharge to bedrock groundwater occurs during the colder months of the year (late November to early April) when evapotranspiration rates are low. However, recharge is more likely to occur in significant amounts throughout the year at the higher elevations (above about 800 m asl) where evaporation and transpiration rates are relatively low even during the warmer months.

In terms of water supply evaluation, the upper portion of the watershed (above 800 m asl) may be considered a potential year-round recharge zone in which a relatively high percentage of the precipitation that falls each year recharges groundwater. Below 750 m asl, only the precipitation falling (or melting) during the colder months of the year is recharged. Based on a simple water balance analysis, the potential year-round recharge zone provides about 73% of the annual recharge to the basin. This study provides strong additional evidence for the importance of New England's upland areas as sources of significant recharge to bedrock aquifers.

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Thompson, 1991). The rock exhibits a strong bedding schistosity and folding oriented north–south along the axis of the Green Mountain Anticlinorium (Christman, 1959). Brittle fractures exist in the form of steeply dipping joints as well as sheeting joints observed in cliff faces. The spacing of joints varies widely from 1 or 2 cm to greater than 1 m at most outcrops. Weathering along foliation surfaces is seen in outcrop on the exposed ridges above treeline, but is not evident with depth in relatively recent cuts or in less exposed areas at lower elevations. Fracture orientation measurements taken at outcrops throughout the basin indicate the presence of three major fracture sets: (1) north–south; (2) N80°E; and (3) N30°E. However, no single fracture set is predominant, suggesting that there is a high potential for interconnection of fractures. Considering the added effect of the sheeting joints and weathered foliations (at the upper elevations), it is likely that recharging groundwater experiences some degree of mixing with groundwater recharged during earlier precipitation or snowmelt events during transport in the bedrock. Such mixing has the potential to homogenize the isotopic signature of groundwater that resides in the bedrock aquifer for several months or longer. The highly metamorphosed nature of the phyllite and schist results in low matrix porosity. Therefore, for the purpose of this study, we have assumed that groundwater flow is fracture-dominated and that the effect of matrix diffusion is minimal.

Urbanization in the basin is limited, consisting of about 12 year-round homes, three summer residences, and a recreational area (Underhill State Park) that operates only in the summer months. The small town of Underhill Center lies just downgradient from the study basin (Fig. 1). Residents in the basin draw groundwater from drilled bedrock wells or bedrock springs for their water supply. Bedrock well yields range from less than 6×10^{-5} m^3/s (about 1 gpm) to about 6×10^{-4} m^3/s (10 gpm). This area was chosen for study because it is typical of many upland areas in New England that may face new challenges in managing water resources as residential populations increase.

weekly from July 1995 to December 1996 at six locations (P4–P9) representative of the elevation distribution in the upland basin (Fig. 1) and at three locations (P1–P3) slightly downgradient from the basin (not shown in figure). Rain samples were collected through a funnel into 1 liter glass bottles, which were insulated to minimize evaporation. Snowmelt samples ($n = 40$) were collected at four locations (M4, M6, M7 and M8) in lysimeters set at the ground surface. Snow samples ($n = 54$) were obtained at four locations in the basin (P4, P6, P7 and P8) and one downgradient location (P1) by coring the recently fallen portion of the snow pack, then allowing the samples to melt in sealed containers. Cumulative snow pack samples ($n = 29$) were also obtained by coring the entire snow pack at stations P6 and P8.

Untreated groundwater samples ($n = 311$) were collected weekly from seven wells (Fig. 1) from October 1995 to December 1996. Two additional wells (GW2 and GW7b, which is located at the same residence as GW7) were not included in the weekly sampling program, but were sampled on several occasions during the study. Samples were collected directly from an outside or inside tap after system purging. Purging time was determined by stabilization of temperature, specific conductance and pH in a separate study (Autery, pers. Comm, 1996). Three of the wells, GW3, GW7, and GW8, are deep uncased bedrock wells with depths of 53, 213, and 83 m, respectively. GW8 is located in the town of Underhill Center, downgradient from the study area. The remaining groundwater sampling points are developed bedrock springs. The high-elevation spring, GW1, is used as a water supply for Underhill State Park on Mount Mansfield.

All samples ($n = 619$) were analyzed for δ^{18}O composition at the University of Vermont Environmental Stable Isotope Laboratory. Samples were equilibrated with CO_2 gas and extracted using standard cryogenic separation techniques (Taylor, 1973; Wong et al., 1987; Socki et al., 1992). Measurements were performed using a VG SIRA Series II Isotope Ratio Mass Spectrometer, and normalized to values obtained for standards. The standard deviation of

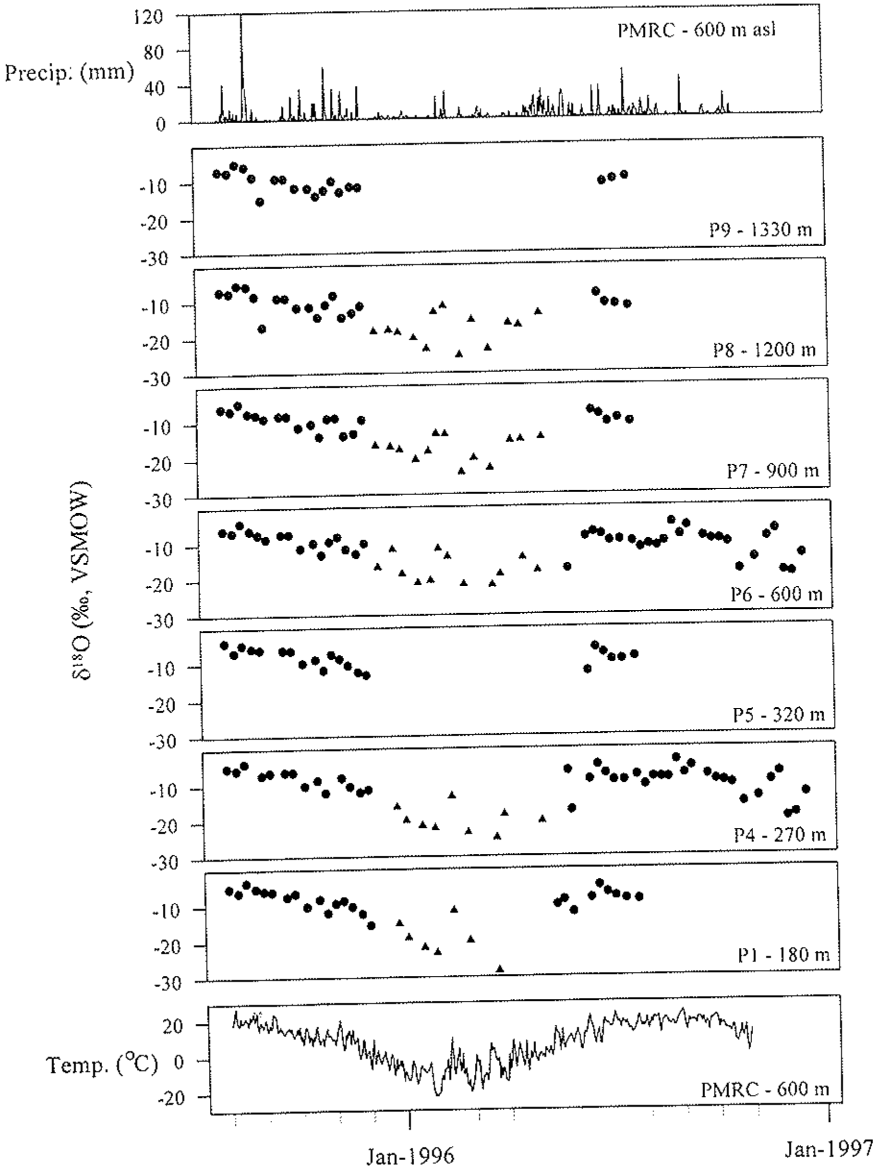

Fig. 2. Precipitation δ^{18}O records. (●) rain samples. (▲) snow samples. Snow samples were not collected at stations P5 and P9. Daily temperature and precipitation records from measurements taken at Proctor Maple Research Center (600 m asl), Underhill (Cummings, 1997, pers. comm.).

locations GW1, GW3, and GW5, and 25

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