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SEASONAL ISOTOPE HYDROLOGY OF THREE APPALACHIAN FOREST CATCHMENTS

D. R. DEWALLE[1], P. J. EDWARDS[2], B. R. SWISTOCK,[1] R. ARAVENA[3] AND R. J. DRIMMIE[3]

[1] *School of Forest Resources and Environmental Resources Research Institute, The Pennsylvania State University, University Park, PA 16802, USA*

[2] *USDA Forest Service, Northeastern Forest Experiment Station, Parsons, WV 26287, USA*

[3] *Waterloo Center for Groundwater Research and Department of Earth Sciences, University of Waterloo, Waterloo, ON, N2L 3G1 Canada*

ABSTRACT

Seasonal oxygen-18 variations in precipitation, throughfall, soil water, spring flow and stream baseflow were analysed to compare the hydrology of two forested basins in West Virginia (WV) (34 and 39 ha) and one in Pennsylvania (PA) (1134 ha). Precipitation and throughfall were measured with funnel/bottle samplers, soil water with ceramic-cup suction lysimeters and spring flow/baseflows by grab and automatic sampling during the period March 1989 to March 1990. Isotopic damping depths, or depths required to reduce the amplitude of subsurface oxygen-18 fluctuations to 37% of the surface amplitude, were generally similar for soil water on the larger PA basin, and baseflows and headwater spring flows on the smaller WV basins. Computed annual isotopic damping depths for these water sources averaged 49 cm using soil depth as the flow path length. The equivalent annual mean hydraulic diffusivity for the soil flow paths was 21 cm^2 d^{-1}. Mean transit times, based upon an assumed exponential distribution of transit times, ranged from 0·2 y for soil water at a depth of 30 cm on the larger catchment, to 1·1–1·3 y for most spring flows and 1·4–1·6 y for baseflows on the smaller catchments. Baseflow on the larger PA basin and flow of one spring on a small WV basin showed no detectable seasonal fluctuations in oxygen-18, indicating flow emanated from sources with mean transit times greater than about 5 y. Based upon this soil flow path approach, it was concluded that seasonal oxygen-18 variations can be used to infer mean annual isotopic damping depths and diffusivities for soil depths up to approximately 170 cm. © 1997 John Wiley & Sons, Ltd.

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INTRODUCTION

Environmental isotopes such as oxygen-18 (^{18}O) and deuterium (^{2}H) have been used commonly to trace stream flow sources during runoff events (e.g. Sklash *et al.*, 1976, DeWalle *et al.*, 1988). Few studies, however, have explored the use of these isotopes for understanding the seasonal dynamics of soil water, stream flow and spring flow. This latter application is based upon a distinct seasonal ^{18}O and ^{2}H pattern observed in precipitation (Deines *et al.*, 1990), which can be approximated by sine waves and traced in various subsurface water sources. The seasonal isotopic signal of precipitation has been used to evaluate the mean transit time for stream baseflow (Lindstrom and Rodhe, 1986; Turner *et al.*, 1987), soil water (Stewart and McDonnell, 1991) and spring flow (Ramspacher *et al.*, 1992). Maloszewski *et al.* (1983) described a model for determining subsurface water transit times based upon sine-wave analysis of seasonal isotope variations in precipitation and subsurface sources. Sine-wave theory used for analysis of soil temperature fluctuations

* Correspondence to: D. R. DeWalle.

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(Sellers, 1972) can also be extended to compute isotopic damping depths and soil hydraulic diffusivities to further describe subsurface hydrology on catchments.

In this paper we use seasonal oxygen-18 variations in precipitation, throughfall, soil water, stream flow and spring flow to compare the hydrology of Appalachian forest catchments of different sizes. Small catchments included in the analysis are the main control watershed and an adjacent treated watershed on the Fernow Experimental Forest in West Virginia. These watersheds are part of a paired-catchment experiment being used to determine the long-term effects of ammonium sulfate applications (Adams *et al.*, 1993) where similar hydrological behavior is assumed. Isotopic analysis has been previously conducted to learn about the storm flow hydrology of these basins (Aravena *et al.*, 1991). Seasonal variations in isotopic data from these two small experimental catchments are compared with seasonal isotopic data from a much larger Pennsylvania watershed (DeWalle *et al*, 1993). Specifically, sine-wave functions fitted to seasonal isotope data for inputs and outputs were used to determine if amplitudes, phase lags, isotopic damping depths and estimated mean transit times varied between the two small Fernow watersheds and between the Fernow watersheds and the larger Benner Run basin in Pennsylvania.

METHODS

Study sites

Data were collected on forested catchments at two unglaciated, Appalachian Mountain locations: the Fernow Experimental Forest in north-central West Virginia (39°05′N, 79°49′W) near Parsons and the Benner Run catchment in north-central Pennsylvania (40°56′N, 78°01′W) near Black Moshannon State Park. Two small Fernow catchments (Figure 1), WS3 and WS4, with areas of 34 and 39 ha, respectively, were compared with the larger Benner Run catchment (1134 ha, Figure 2). WS3 and WS4 slopes average 27 and 20%, respectively, while slopes on Benner Run are approximately 10–15%.

Bedrock at Benner Run is dominated by sandstones and shales from the Burgoon sandstone, Mauch Chunk formation and Pottsville group (DeWalle *et al.*, 1993). At Fernow, bedrock is composed of sandstone and shale from the Hampshire formation (Adams *et al.*, 1994). Soils on the Benner Run catchment are primarily stony, sandy loams or loams of the Hazleton and Clymer series. Fernow WS3 and WS4 soils are stony, sandy loams of the Calvin and Berks series. Soils on both Fernow watersheds are about 1 m deep and directly overlie bedrock, while some soils in the lowlands of the Benner Run catchment overlie compacted colluvium and alluvium up to 4 m deep.

Average annual precipitation of 1400 mm at Fernow exceeds the 1100 mm average annual precipitation at Benner Run. Precipitation is distributed uniformly throughout the year at both locations. Although snowfall is common in the region, intermittent melting generally occurs during winter and no significant seasonal snowpack accumulation or pronounced snowmelt season occurred during the period analysed in this paper. Mean annual air temperatures are about 7·3 and 10°C at Benner Run and Fernow, respectively.

Vegetative cover at both sites is mixed hardwood forests. Benner Run trees are generally comprised of large pole- to sawtimber-sized trees with two regrowing clear-cuts in the headwaters. WS4 forest is mature and WS3 forest is regrowing after a complete clear-cut in 1970–1972.

Field measurements

Samples of soil water, throughfall and stream flow were collected at Benner Run, and precipitation, spring flow and stream flow were sampled at Fernow. The timing of data collection roughly coincided so that data could be compared between sites for a similar climatic regime. Sampling at Fernow was described by Aravena *et al.* (1991) and at Benner Run by DeWalle *et al.* (1993).

Benner Run sampling was conducted from March 1989 to March 1990. Composite biweekly bulk throughfall samples were collected with three plastic funnel/bottle sets near the stream gauging station (Figure 2). Funnels mounted near the ground conveyed water to the bottles through plastic tubing, which

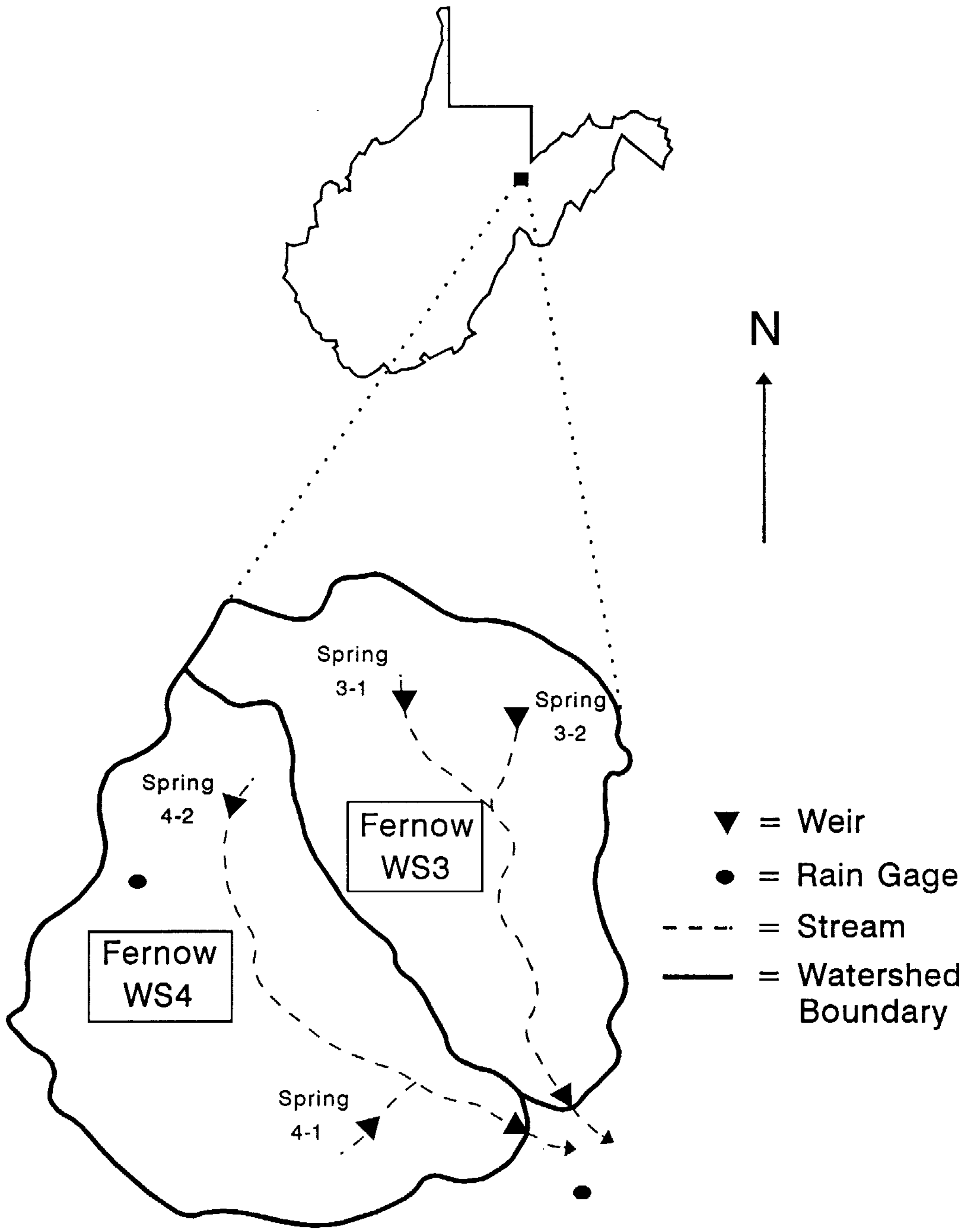

Figure 1. Map of Fernow watersheds WS3 and WS4 showing location of precipitation, baseflow and spring flow sampling sites

was looped to form a water trap and to prevent evaporation of the sample prior to collection. Stream baseflow samples were collected biweekly by grab sampling at the gauging station. Soil water samples were also collected biweekly using three ceramic-cup suction lysimeters installed in mineral soil (30 cm depth) on a broad floodplain and essentially level terrain near the gauging station (Figure 2).

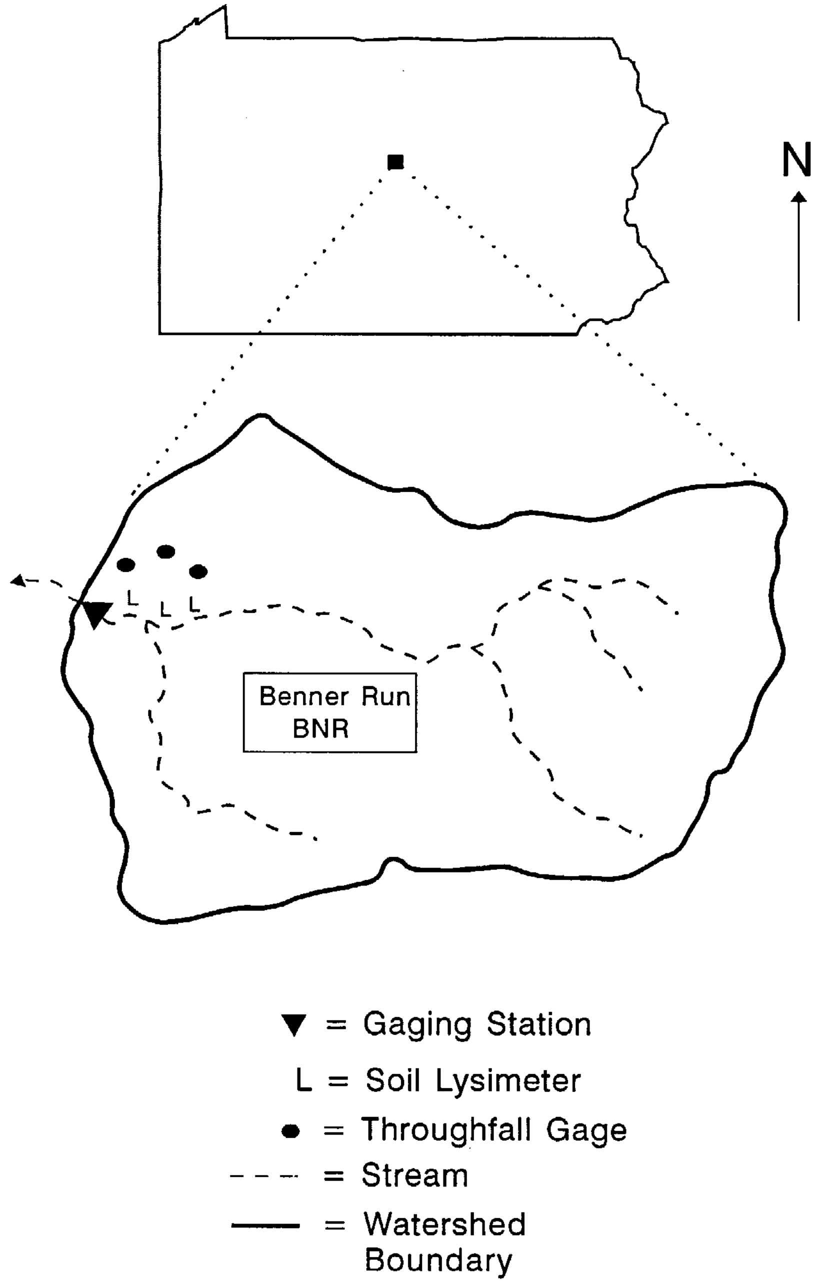

Figure 2. Map of Benner Run watershed showing baseflow, soil water and throughfall sampling sites

Stream flow, spring flow, and precipitation sampling at Fernow was conducted from late autumn 1988 to late summer 1990. Precipitation samples were collected using funnel/bottle collectors at a lower elevation site near the WS4 gauging station (Figure 1). Stream baseflow samples were collected several times weekly at the weirs on WS4 and WS3. Two intermittent headwater springs on each basin (WS3-1, WS3-2, WS4-1 and WS4-2) were also sampled weekly when flow was observed (Figure 1).

Oxygen-18 analysis

Oxygen-18 was analysed with the standard equilibration method (Epstein and Mayeda, 1953) and a VG 903 mass spectrometer at the Environmental Isotope Laboratory, University of Waterloo, Canada. Isotope concentrations were expressed in δ units (‰, parts per mille) defined as:

$$\delta^{18}O = [(R_x - 1)/R_s] \times 1000 \quad (1)$$

where R_x and R_s are the $^{18}O/^{16}O$ ratios for the sample and a standard (SMOW or standard mean ocean water), respectively. Analytical reproducibility of $\delta^{18}O$ determinations was approximately 0·1‰.

Data analysis

Regression analysis was used to determine seasonal oxygen-18 trends in throughfall, precipitation, soil water, spring flow and stream baseflow for the concurrent period of March 1989 to March 1990 at Benner Run and Fernow. Periodic regression analysis (Bliss, 1970) was used to fit a seasonal sine-wave model to oxygen-18 fluctuations as:

$$\delta^{18}O = X + A[\cos(ct - \theta)] \quad (2)$$

where $\delta^{18}O$ is the predicted oxygen-18 level in ‰, X is the annual mean $\delta^{18}O$ value in ‰, A is the $\delta^{18}O$ annual amplitude in ‰, c is the radial frequency of annual fluctuations or 0·017214 rad d^{-1}, t is time in days after 21 March 1989, and θ is the phase lag or time of annual peak oxygen-18 in radians.

Damping and lagging of the seasonal fluctuations with increasing depth or flow path length in the subsurface soil and rock were used to compute an 'isotopic damping depth' (d_h) analogous to the damping depth computed for soil temperature fluctuations based upon sine-wave analysis (Sellers, 1972). In this analogy, it is assumed that vertical fluxes dominate flows in the soil and along slopes to springs and the mouths of the small, steep basins. Weyman (1973) found that soil water flow along slopes was largely vertical except at the boundary of soil with bedrock. Isotopic damping depth was defined as the depth of soil needed to reduce the amplitude of annual oxygen-18 fluctuations to e^{-1} or 37% of the surface value found in precipitation or throughfall. d_h is computed as:

$$\text{amplitude change } d_h = [\ln(A_{z2}/A_{z1})/(z2 - z1)]^{-1} \quad (3)$$

$$\text{phase shift } d_h = [(ct_{maxz2} - ct_{maxz1})/(z2 - z1)]^{-1} \quad (4)$$

where d_h is in cm, A_{z2} is the amplitude at depth $z2$ in ‰, A_{z1} is the amplitude at depth $z1 = 0$ in ‰, t_{maxz2} is the time of maximum oxygen-18 at depth $z2$ in days and t_{maxz1} is the time of maximum oxygen-18 at depth $z1 = 0$ in days. In this analysis precipitation or throughfall fluctuations represent the surface or $z1 = 0$ values. Soil water sampling at Benner Run was conducted at 30 cm depth. At Fernow an average soil depth of 100 cm was assumed for the flow path of headwater spring flows and baseflows (Adams *et al.*, 1994).

Isotopic damping depth can also be equated to an annual average hydraulic diffusivity for water movement in the soil based on an analogy with unsteady heat flow as:

$$D_w = cd_h^2/2 \quad (5)$$

where D_w is the annual average hydraulic diffusivity for water movement in cm^2 d^{-1}.

Mean transit times for water to move through the subsurface system can also be computed based upon a simple steady-state, well-mixed model, in which precipitation is assumed to mix immediately with all water in the soil reservoir (Stewart and McDonnell, 1991). Sine waves are fitted to input and output water tracer concentrations, and an exponential distribution of transit times is assumed. Mean transit time (T) or the mean age of water leaving the system in days is computed as:

$$T = c^{-1}[A_{z2}/A_{z1})^{-2} - 1]^{0 \cdot 5} \tag{6}$$

(see model 1, Maloszewski *et al.*, 1983) where all terms have been defined previously. Application of a steady-state model to water flow in soil, spring flows and baseflows is obviously a first approximation; however, Maloszewski and Zuber (1993) indicate that the well-mixed model has been frequently applied to unconfined aquifers such as the soil in this study.

RESULTS AND DISCUSSION

Seasonal cycles at Benner Run

Seasonal sine-wave regression models fit to the data at Benner Run indicated a damping and lagging of the seasonal oxygen-18 signal from throughfall to soil water to stream baseflow (Figures 3 and 4, Table I). Throughfall seasonal $\delta^{18}O$ amplitude was 3·41‰ with a maximum occurring on 3 August 1989. Throughfall oxygen-18 was highly variable and only crudely described by a sine-wave cycle ($R^2 = 39\%$, Figure 3) during the year. In soil water, the oxygen-18 seasonal cycle was damped and smoothed and better described by a sine wave ($R^2 = 89\%$, Figure 4). Soil water $\delta^{18}O$ amplitude at 30 cm depth was reduced by 40% to 2·06‰ and the peak oxygen-18 did not occur until 24 September, 52 days after the throughfall peak. Stream baseflow oxygen-18 showed no statistically significant seasonal variation on Benner Run and remained essentially constant at $-10 \cdot 12$‰ during the entire period (Figure 4). The minimum detectable amplitude in $\delta^{18}O$ probably is about 0·1‰ using the regression approach.

Mean annual $\delta^{18}O$ in throughfall and soil water (i.e. intercepts in the equations in Table I) were $-9 \cdot 22$ and $-8 \cdot 80$‰, respectively. These values were not significantly different. Mean baseflow $\delta^{18}O$ of $-10 \cdot 12$‰ was significantly less than that in soil water. Similar mean annual deep groundwater $\delta^{18}O$ in limestone aquifers in

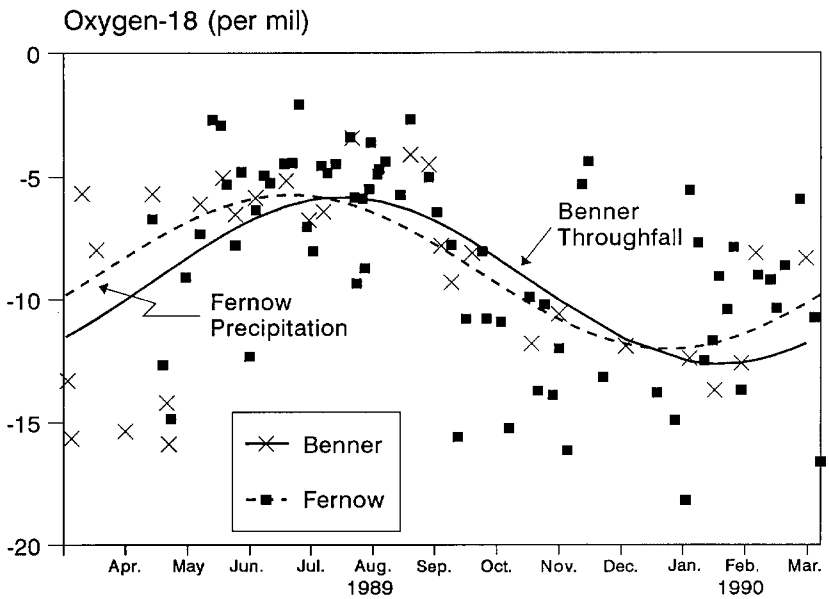

Figure 3. Seasonal oxygen-18 variations in throughfall at Benner Run and precipitation at Fernow; raw data and sine-wave curves fitted to data (March 1989–March 1990)

Table I. Annual oxygen-18 ($\delta^{18}O$) fluctuations in precipitation, throughfall, soil water, spring flows and stream baseflow for the Benner Run and Fernow watersheds (21 March 1989–20 March 1990)

Component	n	Mean (‰)	Amplitude (‰)	Phase lag (d)	R^2	RMSE*
Benner Run						
Throughfall	29	−9·22	3·41	135	39	3·11
Soil water	31	−8·80	2·06	187	89	0·58
Baseflow	29	−10·12	ns†	ns	ns	ns
Fernow						
Precip.	70	−8·87	3·15	110	35	3·26
Baseflow						
WS3	71	−9·16	0·31	224	52	0·226
WS4	71	−9·13	0·36	216	62	0·211
Headwater spring flow						
WS3-1	17	−9·32	0·44	207	60	0·271
WS3-2	15	−9·10	0·39	252	42	0·302
WS4-1	17	−9·81	ns	ns	ns	ns
WS4-2	14	−8·98	0·37	225	59	0·229

* Root mean square error of regression.
† Regression non-significant.

central Pennsylvania of −9·33 and −9·34‰ were found by Deines *et al.* (1990). More depleted oxygen-18 in baseflow water may reflect varying oxygen-18 levels in groundwater recharge in previous years, or the commonly observed situation in which groundwater recharge primarily occurs from winter–spring precipitation, which has more depleted oxygen-18 levels. Deines *et al.* (1990) estimated that about 55–65% of precipitation in this region does not contribute to deep groundwater recharge on the basis of oxygen-18 analysis of precipitation and groundwater samples.

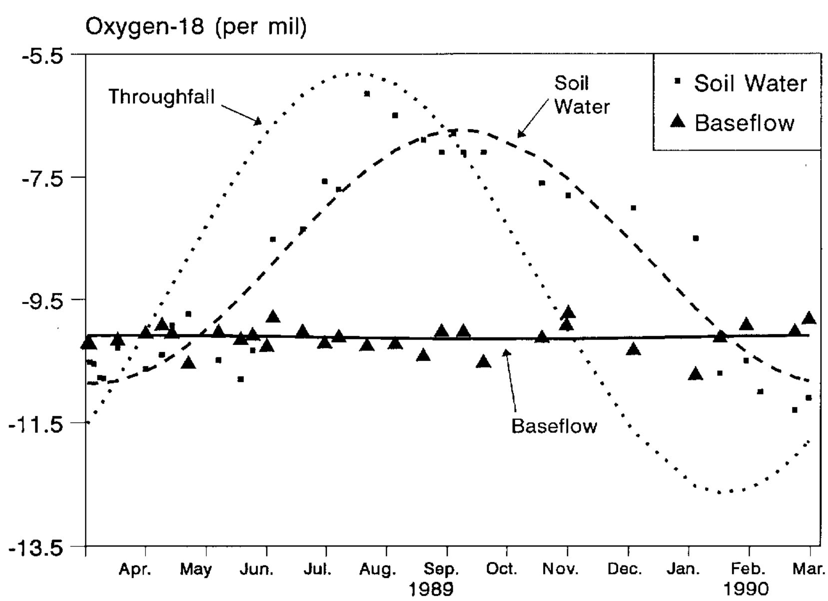

Figure 4. Seasonal oxygen-18 variations in soil water and stream baseflow on Benner Run catchment (March 1989–March 1990). Sine waves fitted to throughfall, soil water and baseflow data (non-significant) are shown for reference

Lower seasonal amplitudes in soil water fluctuations than in throughfall indicated that infiltrated water is mixed with water from previous events stored in the soil. The lagged seasonal cycle of soil water behind the throughfall cycle shows that sampled soil water moved slowly. Suction lysimeters primarily sample slow-moving, micropore soil water rather than rapidly moving, macropore soil water. The non-significant amplitude in baseflow fluctuations indicated a mixing of recharge water with a large volume of stored groundwater from previous years.

Precipitation or throughfall infiltrating the soil can mix with and displace significant amounts of stored soil water into streams during events (DeWalle *et al.*, 1988; Swistock *et al.*, 1989). Figure 4 suggests that evaluation of soil water contributions using oxygen-18 during storm flow events will depend on the time of year. Seasonal oxygen-18 differences between throughfall, soil water and baseflow source waters for stream flow are greatest in late summer. In contrast, events in early spring would not be expected to show clear source contributions because oxygen-18 levels in the three source waters are essentially the same.

Seasonal cycles at Fernow

Seasonal oxygen-18 variations in precipitation at Fernow for March 1989–March 1990 were described by a sine wave with similar accuracy as for Benner Run ($R^2 = 35\%$, Table I, Figure 3). The amplitude of seasonal precipitation $\delta^{18}O$ variations was 3·15‰ at Fernow, slightly lower than the 3·41‰ amplitude for throughfall fluctuations at Benner Run. Throughfall is known to have slightly enriched oxygen-18 levels compared with open precipitation owing to canopy storage and fractionation during canopy evaporation in some storms (DeWalle and Swistock, 1994), but the effects of these processes on the amplitude of annual fluctuations are not known. Peak oxygen-18 in precipitation at Fernow occurred on July 9, about 25 days before peak throughfall at Benner Run. The mean annual precipitation $\delta^{18}O$ at Fernow of −8·87‰ was not significantly different from the mean of −9·22‰ in throughfall at Benner Run.

Seasonal variations in oxygen-18 levels in stream baseflow on WS3 and WS4 were quite similar. Baseflow sine waves on both basins were damped and delayed compared precipitation variations (Table I, Figure 5). Baseflow $\delta^{18}O$ amplitudes on WS3 and WS4 were only 0·31 and 0·36‰, showing a 90 and 89‰ reduction in amplitude compared with precipitation. Peak baseflow oxygen-18 on WS3 and WS4 lagged that in

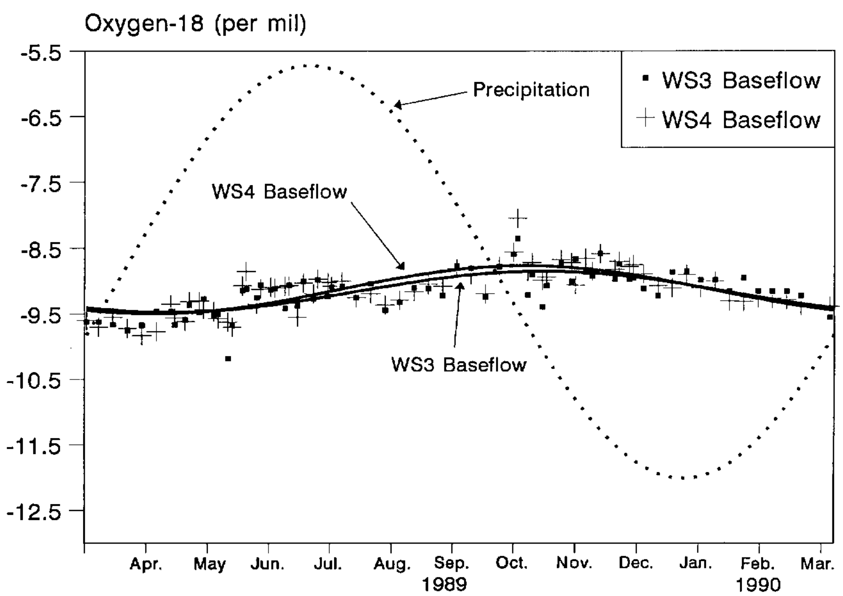

Figure 5. Seasonal oxygen-18 variations in baseflow on Fernow catchments WS3 and WS4 (March 1989–March 1990). Sine waves fitted to precipitation and baseflow data are shown for reference

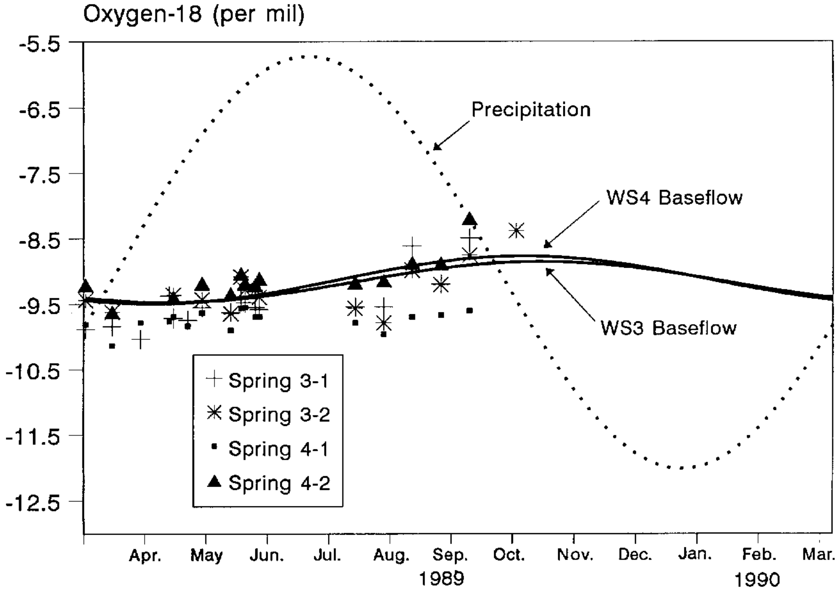

Figure 6. Seasonal oxygen-18 variations in intermittent headwater springs on Fernow catchments WS3 and WS4 (March 1989–March 1990). Sine waves fitted to precipitation and baseflow data are shown for reference

precipitation by 114 and 106 days, respectively, based on sine-wave models. WS4 baseflow showed slightly less amplitude damping (1% less) and slightly shorter lag (8 days shorter) relative to precipitation than WS3; however, overall seasonal oxygen-18 dynamics in baseflow were nearly identical on these two watersheds.

Although headwater springs sampled on WS3 and WS4 were not perennial, oxygen-18 seasonal dynamics of spring flow were generally similar to those for baseflow. On WS3, oxygen-18 of both WS3-1 and WS3-2 spring flows overlapped data for baseflow (Figure 6 and Table I). Oxygen-18 in spring WS4-2 also appeared similar to that in baseflow, but spring WS4-1 showed less seasonal oxygen-18 fluctuation than WS4 baseflow or WS4-2 (Figure 6 and Table I). As expected from Figures 5 and 6, amplitudes and phase angles for seasonal oxygen-18 fluctuations in spring flow were similar to those for baseflow (Table I). No significant seasonal variation was detected for spring WS4-1.

The headwater springs of the two Fernow basins produced water that had already undergone essentially the same oxygen-18 damping and lagging as baseflow water. This suggests that headwater spring flows and baseflow are generated by similar processes, probably from water flowing through the soil profile and downslope on bedrock surfaces. Lack of seasonal fluctuations in spring WS4-1 suggests flow results from a better mixed subsurface reservoir.

Mean annual oxygen-18 levels in baseflow and spring flows from both Fernow basins showed slightly greater oxygen-18 depletion than in precipitation. Mean annual $\delta^{18}O$ in baseflow and the three similar spring flows at Fernow ranged from $-8{\cdot}98$ to $-9{\cdot}32$‰, while that in precipitation was $-8{\cdot}87$‰. Spring WS4-1, which showed no significant seasonal fluctuations, indicated the greatest mean annual oxygen-18 depletion at Fernow ($\delta^{18}O = -9{\cdot}81$‰). Depletion in spring WS4-1 was similar to that for Benner Run baseflow ($\delta^{18}O = -10{\cdot}12$‰, Table I). Depletion observed in spring WS4-1 at Fernow and baseflow at Benner Run probably reflects a water contribution from a well-mixed reservoir that is recharged preferentially by winter and spring precipitation depleted in oxygen-18.

Annual isotopic damping depths and hydraulic diffusivities

Isotopic damping depths and hydraulic diffusivities obtained for soil at 30 cm depth at Benner Run and for baseflow and spring flow at Fernow generally were similar (Table II). At Benner Run, isotopic damping

Table II. Computed annual isotopic damping depths and hydraulic diffusivities for subsurface flow at Benner Run and Fernow catchments

Site/source	Damping depth (cm)		Diffusivity ($cm^2\ d^{-1}$)		Transit time (y) [Eq. (6)]
	Amplitude change [Eq. (3)]	Phase shift [Eq. (4)]	Amplitude change [Eq. (3) + (5)]	Phase shift [Eq. (4) + (5)]	
Benner Run					
Soil water @ 30 cm	59	34	30	10	0·2
Baseflow	ns*	ns	ns	ns	> 5·0†
Fernow					
Baseflow					
WS3	43	51	16	22	1·6
WS4	46	55	18	26	1·4
Headwater spring flow					
WS3-1	51	60	22	31	1·1
WS3-2	49	41	21	14	1·3
WS4-1	ns	ns	ns	ns	> 5·0†
WS4-2	47	50	19	22	1·3
Mean	49	49	21	21	

* Non-significant annual fluctuations.
† Computed for minimum detectable $\delta^{18}O$ amplitude of 0·1‰.

depths for the upper soil profile (0–30 cm depth) were 59 cm computed from the amplitude shift and 34 cm computed from the time lag for maximum oxygen-18 occurrence (Table II). Equivalent annual diffusivities for water movement at 30 cm soil depth at Benner Run were 30 and 10 $cm^2\ d^{-1}$, respectively. Mean isotopic damping depths for baseflow on WS3 and WS4 and the three springs with significant fluctuations were 47 cm based on amplitude shifts and 51 cm based on lag times (Table II) for an assumed soil depth of 1 m. These isotopic damping depths were equivalent to mean annual hydraulic diffusivities of 19 and 23 $cm^2\ d^{-1}$, respectively, and are within the same order of magnitude as at Benner Run.

The general agreement of computed soil damping depths and diffusivities between Benner Run soil water and Fernow baseflow and spring flow is surprising. Isotope fluctuations at Fernow represent variations in baseflow and spring flow, which undoubtedly includes flow through the soil and over bedrock surfaces, in contrast to percolation through a known depth of soil at Benner Run. Computing damping depths and diffusivities using soil depth as the path length at Fernow essentially assumes that the soil primarily controls baseflow and spring flow dynamics at Fernow. This seems a reasonable first approach.

Hydraulic diffusivities computed for soil at Benner Run, and baseflow and spring flow fluctuations at Fernow, fall in the lower moisture content range of expected soil hydraulic diffusivities. Hydraulic diffusivities in soil generally range from about 1–10 $cm^2\ d^{-1}$ at low moisture contents to a maximum value of about $10^4\ cm^2\ d^{-1}$ (Hillel, 1982) at high moisture contents. Diffusivities obtained using the method in this paper represent annual average conditions, and they seem to reflect water movement during periods when soils are at lower moisture contents.

Owing to the nature of sampling, damping depths and diffusivities computed in this study probably represent the slow unsteady flow of water in soil micropores rather than rapid flow through soil macropores. Soil suction lysimeters on Benner Run are expected to sample primarily micropore water, and the timing of baseflow and spring flow sampling was designed largely to exclude storm flow periods. Thus, rapid macropore flow in these forest soils probably is not reflected in the derived hydraulic soil parameters.

Damping depth multiplied by 4·61 is often taken as the depth where fluctuations reach 1% of the surface value (Sellers, 1972); thus, in this study the 1% damping depth would be approximately

$4{\cdot}61 \times 49$ cm $= 226$ cm. At this depth, annual oxygen-18 amplitudes would be 0·0315 or 0·0341‰ depending on the site. An amplitude of this magnitude is below the probable minimum 0·1‰ detection limit for oxygen-18 measurements. Thus, for soils encountered in this study, this method of deriving an isotopic damping depth or diffusivity using oxygen-18 fluctuations would be applicable for soil depths up to about 170 cm.

Mean transit times

Mean transit times were computed for soil water at Benner Run and baseflow and spring flow on both Fernow basins (Table II) using the well-mixed model (equation 6, Maloszewski *et al*., 1983). The mean transit time for soil water at the 30 cm depth at Benner Run was 76·5 d (0·2 y). Baseflow in WS3 and WS4 at Fernow showed mean transit times of 1·6 and 1·4 y, respectively. Except for spring WS4-1, Fernow springs showed slightly lower mean transit times, about 1·1–1·3 y, than stream baseflow. Benner Run baseflow and WS4-1 spring flow showed no significant seasonal fluctuations. The minimum detectable oxygen-18 amplitude of about 0.1‰ (the oxygen-18 analytical reproducibility) is roughly equivalent to a mean transit time of 5 y. This implies that the water supplying Benner Run baseflow and Fernow WS4-1 spring flow had at least a 5-year mean transit time.

Fernow springs were expected to show lower mean transit times than baseflow since they emanated in the headwaters and appeared to have smaller contributing areas than baseflows at the mouth of the catchments. However, mean transit times for spring flow and baseflow varied by only 0·3–0·5 y on WS3 and 0.1 y on WS-4 (excluding spring WS4-1). Similar transit times indicate that springs supply a significant proportion of baseflow on these small catchments.

Lack of seasonal baseflow oxygen-18 fluctuations at Benner Run is probably mainly related to the larger basin area than at Fernow. The larger basin area at Benner Run can increase the flow path length for subsurface water to reach the basin mouth, which can increase the opportunity for subsurface mixing and storage. In addition, slopes on the Benner Run catchment are less steep and soils in the valley bottom are deeper than at Fernow.

Other studies on small catchments have shown varying mean subsurface transit times of up to about 1 y. Stewart and McDonnell (1991), using the well-mixed model, calculated mean water transit times of 12, 15, 48–65 and 100 days for soil depths of 20, 30, 40 and 80 cm, respectively, when studying interannual deuterium variations on the Maimai M8 watershed in New Zealand. They also found mean transit times increased from ridge tops to valley bottoms, reflecting mixing with progressively greater amounts of older stored water. In the current study, our soil water data were collected at shallow depth on relatively level terrain and probably do not include contributions of water from upslope. Using a modified version of the PULSE model, Lindstrom and Rodhe (1986) computed mean transit times of 212 and 365 days, respectively, using oxygen-18 fluctuations in stream baseflow water in two small Swedish basins (4 and 180 ha). Turner *et al*. (1987) found time lags between rainfall and stream flow ranging from 20 to about 50 days depending on antecedent moisture conditions on an 80-ha eucalypt forest basin in Western Australia. All of these small catchment studies showed relatively rapid response to rainfall.

CONCLUSIONS AND MANAGEMENT IMPLICATIONS

As is often assumed in paired watershed studies, variations in oxygen-18 indicate that the two adjacent Fernow watersheds, WS3 and WS4, have nearly identical baseflow and spring flow seasonal dynamics. Computed mean transit times or observed time lags between surface and subsurface oxygen-18 fluctuations, suggest that watershed treatments or disturbances should affect Fernow baseflow and most Fernow spring flows within one to two years. Adams *et al*. (1993) have reported that nitrate exported in stream flow from Fernow WS-3 increased during the first year of watershed fertilization with ammonium sulfate. Rapid change should also occur in baseflow or spring flow following cessation of a treatment or disturbance on the Fernow watersheds.

Lack of detectable seasonal fluctuations in baseflow oxygen-18 on Benner Run basin indicate that greater groundwater storage and mixing occurred on this larger basin than on the smaller Fernow watersheds. Detection limits for oxygen-18 suggest mean transit times for Benner Run baseflow exceed about 5 years. Evidence of watershed disturbance or treatment effects in stream water would be greatly delayed on Benner Run.

Seasonal oxygen-18 fluctuations detectable in subsurface waters can be used with an estimate of flow path length to derive annual isotopic damping depths and hydraulic diffusivities assuming a vertical subsurface flow system. With flow path length set equal to lysimeter sampling depth for soil water and total soil depth for spring flows and baseflows, soil at Benner Run and subsurface flow systems controlling spring flows and baseflows at Fernow showed remarkably similar isotopic damping depths and hydraulic diffusivities.

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