WATER RESOURCES RESEARCH, VOL. 33, NO. 12, PAGES 2637–2653, DECEMBER 1997

Subsurface flow paths in a steep, unchanneled catchment

Suzanne Prestrud Anderson,[1] William E. Dietrich,[2] David R. Montgomery,[3]
Raymond Torres,[4] Mark E. Conrad,[5] and Keith Loague[6]

Abstract. Tracer studies during catchment-scale sprinkler experiments illuminate the pathways of subsurface flow in a small, steep catchment in the Oregon Coast Range. Bromide point injections into saturated materials showed rapid flow in bedrock to the catchment outlet. Bedrock flow returned to the colluvium, sustaining shallow subsurface flow there. The bromide peak velocity of $\sim 10^{-3}$ m s^{-1} exceeded the saturated hydraulic conductivity of intact bedrock. This, and the peak shapes, verify that fractures provide important avenues for saturated flow in the catchment. Deuterium added to the sprinkler water moved through the vadose zone as plug flow controlled by rainfall rate and water content. Ninety-two percent of the labeled water remained in the vadose zone after 3 days (~140 mm) of sprinkling. Preferential flow of new water was not observed during either low-intensity irrigation or natural storms; however, labeled preevent water was mobile in shallow colluvium during a storm following our spiking experiment. In response to rainfall, waters from the deeper bedrock pathway, which have traveled through the catchment, exfiltrate into the colluvium mantle and mix with relatively young vadose zone water, derived locally, creating an area of subsurface saturation near the channel head. This effectively becomes a subsurface variable source area, which, depending on its size and the delivery of water from the vadose zone, dictates the apportioning of old and new water in the runoff and, correspondingly, the runoff chemistry. The slow movement of water through the vadose zone allows for chemical modification and limits the amount of new water in the runoff. Moreover, it suggests that travel time of new rain water does not control the timing of runoff generation.

1. Introduction

The mechanisms of subsurface storm flow have been debated since *Hursh* [1936] observed that overland flow was not the source of storm runoff in forested catchments. Despite progress in the intervening years, *Hornberger and Boyer* [1995] concluded in a recent review that "there is much to be learned about complex flow paths within catchments" (p. 954) and that "hydrological science is in greater need of more and better experimentation" (p. 954). Understanding the flow of water in the subsurface of a catchment is an important prerequisite for understanding the source of solutes in streams [e.g., *Nielsen et al*., 1986; *Trudgill*, 1988; *Brusseau and Rao*, 1990; *Williams et al*., 1990; *Wilson et al*., 1991; *Litaor*, 1992; *Luxmoore and Ferrand*, 1993; *Anderson et al*., 1997; *Buttle and Peters*, 1997]. Subsurface pathways define the earth materials that waters will see, and these pathways set transit times, which prescribe the time available for chemical reactions. Movement of water in the subsurface also controls the production of storm runoff in humid environments [*Freeze*, 1974] and therefore must be understood to model runoff generation.

The goal of this paper is to describe flow paths operating in the CB1 catchment, near Coos Bay in the Oregon Coast Range. Subsurface flow is the sole source of storm discharge in the CB1 catchment, as is common in humid, soil-mantled, vegetated, headwater regions [*Dunne et al*., 1975; *Freeze*, 1974; *Cheng*, 1988; *Turton et al*., 1992; *Bonell*, 1993; *Mulholland*, 1993; *Allan and Roulet*, 1994; *Eshleman et al*., 1994; *Peters et al*., 1995; *Sidle et al*., 1995]. Detailed monitoring and whole-catchment sprinkler experiments over several years at CB1 have yielded considerable insight into the hydrology of the site. *Montgomery et al*. [1997] demonstrated that saturated water flow through bedrock is important at CB1. The near-surface development of exfoliation fractures in the underlying bedrock of this site has created a complex, locally highly conductive hydrologic environment capable of conveying nearly all storm runoff. The importance of bedrock flow at CB1 is in accord with the few other studies of headwater catchments in which bedrock flow paths have been considered [*Wilson and Dietrich*, 1987; *Herwitz*, 1993]. From tensiometric response (R. Torres et al., The influence of the unsaturated zone on the hydrologic response of a small catchment, submitted to *Water Resources Research*, 1997, hereinafter referred to as submitted manuscript) showed that the vadose zone sets the timing of runoff development at CB1. Owing to the steep characteristic curve for these soils, even relatively light rainfall tends to drive pressure heads to zero and create vertical flow through the vadose zone driven largely by the elevation head. The importance of these flow paths on runoff chemistry was suggested by concentration-discharge relationships at CB1 [*Anderson et al*., 1997].

[1]Department of Earth Sciences and Institute of Tectonics, University of California, Santa Cruz.
[2]Department of Geology and Geophysics, University of California, Berkeley.
[3]Department of Geological Sciences, University of Washington, Seattle.
[4]Department of Geological Sciences, University of South Carolina, Columbia.
[5]Berkeley Center for Isotope Geochemistry, Lawrence Berkeley Laboratory, Berkeley, California.
[6]Department of Geological and Environmental Sciences, Stanford University, Stanford, California.

Paper number 97WR02595.
0043-1397/97/97WR-02595$09.00

Experiments with different sprinkler water chemistry yielded similar runoff chemistry, indicating that transit through the vadose zone buffers water chemistry and plays an important role in setting runoff composition. In this paper we present data from tracer experiments conducted during our last sprinkling experiments at the site. The sprinkler experiment was designed to produce a long period of steady flow conditions during which the tracer studies of saturated and vadose zone flow could be conducted, so that the hypothesized flow paths based on head gradients and outflow chemistry could be directly illuminated.

One objective of the tracer experiments at CB1 was to measure subsurface water flow velocities through parts of the catchment under controlled conditions. Hydraulic conductivities in the forest soils of the Coast Range are high [*Harr*, 1977; *Hammermeister et al*., 1982; *Montgomery et al*., 1997], and slopes in headwater catchments are steep (mean at CB1 is 43°), yet we have found the time required to develop full runoff response is not short (R. Torres et al., submitted manuscript). This is at odds with the predictions of simple models of runoff generation [e.g., *Humphrey*, 1982; *Iida*, 1984; *Robinson and Sivapalan*, 1996], in which the time required to produce full runoff response is controlled by the travel time of water from the furthest reaches of the catchment.

A second, related objective of the tracer experiments was to explore the transit time for water through the catchment and residence time of water in the catchment. The study of runoff generation for many years focused on identifying subsurface flow mechanisms fast enough to produce storm discharge [*Dunne*, 1978]. The variable source area conceptual model identified dynamic regions near streams where rapid saturation overland flow paths (both return flow and direct precipitation onto saturated areas) could be established quickly during storms to produce the runoff response. The use of isotopic tracers led to recognition that old or "preevent" water often dominates storm runoff in humid catchments with high infiltration capacities [*Sklash*, 1990; *Bonell*, 1993; *Buttle*, 1994], implying relatively long residence times (greater than one storm event) for much of the water in catchments. This led to a shift in emphasis to identifying means of displacing old water from catchments by subsurface storm flow [*Kennedy et al*., 1986; *Pearce et al*., 1986; *Sklash et al*., 1986; *McDonnell*, 1990; *Ogunkoya and Jenkins*, 1991; *Wilson et al*., 1991]. Two whole-catchment applications of isotopic tracers in the Lake Gårdsjön watershed illustrate how long residence times of water can be. Within 1 year, only 40% of the tritium applied below the rooting zone in two catchments (3.6 and 2.8 ha) was displaced [*Nyström*, 1985], while a step change in the $\delta^{18}O$ of water used to irrigate a smaller (0.6 ha) covered catchment propagated through to the runoff in 7.5 months [*Rodhe et al*., 1996]. The latter study concluded that "water of different ages mixes near the outlet and in the runoff, giving the same effect in the runoff as if there were complete mixing in the catchment as a whole" [*Rodhe et al*., 1996, p. 3510]. This observation justifies the common use of mixing models to describe runoff components but highlights how this practice obscures the actual routes traveled by water through catchments.

The two tracer experiments at the CB1 catchment are unique in that they began during steady flow conditions, when tracer travel times and fluxes can be interpreted with respect to established head gradients. Our monitoring of the tracers continued through an additional short sprinkler run conducted 8 days later and a brief storm, which allowed us to see how contributions of vadose zone and saturated flow paths varied under transient flow conditions. One experiment was a catchment-wide application of deuterium in the rainwater. In effect, this experiment is similar to the step change in precipitation composition of *Rodhe et al*. [1996], although our experiment design is focused on the short-term fate of the tracer. The other tracer experiment consisted of bromide injections into saturated colluvium and bedrock to examine flow in the saturated zone. For comparison with these sprinkler experiments, we were fortunate to sample a 5-day natural storm at another time that was amenable to isotope hydrograph separation.

We begin with the hydrograph separation for a natural storm. As is often the case in monitoring natural events, it was not possible to identify unequivocally the pathways contributing to storm runoff. The difficulties stem both from insufficient sample collection prior to the storm and from the short duration of the event. We then present the tracer experiments during the sprinkling experiment, first detailing movement of water parcels through the saturated zone and then examining flow in the vadose zone. These experiments provide critical direct evidence of hypothesized flow paths and reveal that runoff is controlled by a dynamic subsurface region just upslope from the channel head, where there is an interplay of exfiltrating bedrock-derived flow and vertical unsaturated flow through the colluvium mantle. This area acts in a manner similar to the variable source area that generates saturation overland flow, and plays a comparably important role in runoff dynamics and hydrogeochemistry.

2. Field Site

The CB1 study catchment is an 860 m^2, steep (43°), unchanneled valley (Figure 1) in the Oregon Coast Range near Coos Bay. The hillslopes in the Coast Range are corrugated into a nose-hollow-nose topography, with thickened accumulations of colluvium in the hollows [*Dietrich et al*., 1986]. The CB1 catchment, comprising one such corrugation, is small enough to manipulate and monitor and yet comprises the entire source area for a first-order ephemeral stream.

Organic-rich, sandy colluvium, with an average bulk density of 720 ± 20 kg m^{-3} ($n = 10$) and porosity of 0.71 ± 0.03 ($n = 12$), thickens from a few tens of centimeters on the edges of the catchment to a maximum of 2 m in the axis of the catchment. Soil developed in the colluvium is classified as a Haplumbrept [*Haagen*, 1989]. The colluvium can transmit substantial amounts of water as unsaturated flow [*Yee and Harr*, 1977; R. Torres et al., submitted manuscript]. A second-growth forest of Douglas fir (*Pseudotsuga menzeisii*) was logged from the site in early 1987. One application of herbicide preceded replanting in 1988, a year before our hydrologic monitoring of the site began.

The Eocene volcaniclastic sandstone bedrock (Flournoy formation [*Baldwin*, 1974]) dips 8°–17° south into the slope. On the basis of a 35-m-long core obtained at the ridge crest, and on additional shallower cores elsewhere, the weathered bedrock was divided into three layers (see Figure 2 for interpretation of this stratigraphy at the lower portion of the slope). Saprolite, defined as oxidized bedrock that can be penetrated with a hand auger, is up to 0.5 m thick in the upper part of the catchment but is absent on the lower third of the slope. Pervasively oxidized bedrock thins from 4 m upslope to 0 at the channel head. Fractured, partially oxidized bedrock thins from 4 to 3 m downslope. We consider the rock below this fractured layer to be unweathered, though it is cut by a few open fractures with oxidation staining to a depth of 24 m at the ridge crest. Oxidation staining was absent below 24 m, which also corresponds to the annual maximum depth of the water table in the well created by drilling the core. These layer thicknesses are known to within a few tenths of meters at the ridge crest, where they are constrained by the core, for which there was 93% recovery between 3 and 35 m depth. Elsewhere, the thickness of the weathered rock layers below saprolite have uncertainties of perhaps 50%.

Runoff generation in the catchment is entirely produced from subsurface stormflow [*Montgomery et al.*, 1997]. Water flows vertically down through the vadose zone during rainfall events (R. Torres et al., submitted manuscript) and reaches the subsurface saturated zone in either the colluvium or bedrock. The water table does not reach the soil surface anywhere on CB1. The pattern of saturation within the colluvium (Figure 1) implies that most water travels through the bedrock during its transit through the catchment. Indeed, we have measured infiltrating and exfiltrating gradients across the bedrock-colluvium interface during natural storms and sprinkling experiments [*Montgomery et al.*, 1997].

3. Experimental Design and Monitoring

3.1. Instrumentation

The CB1 catchment is instrumented with time domain reflectometry (TDR) wave guides, piezometers, tensiometers, rain gauges, lysimeters, and two weirs at the catchment outlet. The hydrologic instrumentation was arrayed along rows across the catchment, localized by suspended wooden platforms designed to prevent trampling of the soil.

3.1.1. Input and output measurements. A total of 148 wedge rain gauges were arrayed across the catchment and read twice daily during sprinkling experiments. Storms were monitored with three tipping bucket rain gauges at top, mid, and bottom slope locations. Rainwater was collected for chemical analysis in a high-density polyethylene (HDPE) bottle about 0.5 m off the ground surface, well above local vegetation.

Two weirs, equipped with flumes with V notches and stage height recorders, measured water discharge from the catchment. Soil-water flow was funneled into the flume at the upper weir, located at the channel head, by 3-m-long plastic-lined plywood walls sealed to the bedrock. The lip of the flume was cemented to bedrock. Outflow from the upper weir was routed around the lower weir, located 15 m down channel, to separate unambiguously measurements at the two locations. The rating curves for both weirs were developed with manual measurements during water sample collection. Samples were collected at the weirs at 6-hour intervals during sprinkling experiments and less frequently during natural storms.

3.1.2. Soil water sampling. Soil water samples were collected from 34 lysimeters (Table 1), which are arrayed in six nests of 2–8 instruments, and in scattered isolated installations (Figure 1). The lysimeter nests at rows 5, 6, 9B, and 11B all lie along the long axis of the catchment, where the deepest colluvium occurs and the thickest saturation layer developed in previous experiments [*Montgomery et al.*, 1997].

Three lysimeter designs were used in an attempt to sample both readily mobile soil water (zero-tension lysimeters) and less mobile water held in tension (plate [*Driscoll et al.*, 1985] and cup-style suction lysimeters). To minimize soil disruption, we installed lysimeters in either auger holes or narrow hand-dug horizontal slots, rather than trenches. All were installed in May 1991, 9 months before the first samples were collected and 1 year before the sprinkling experiments described below.

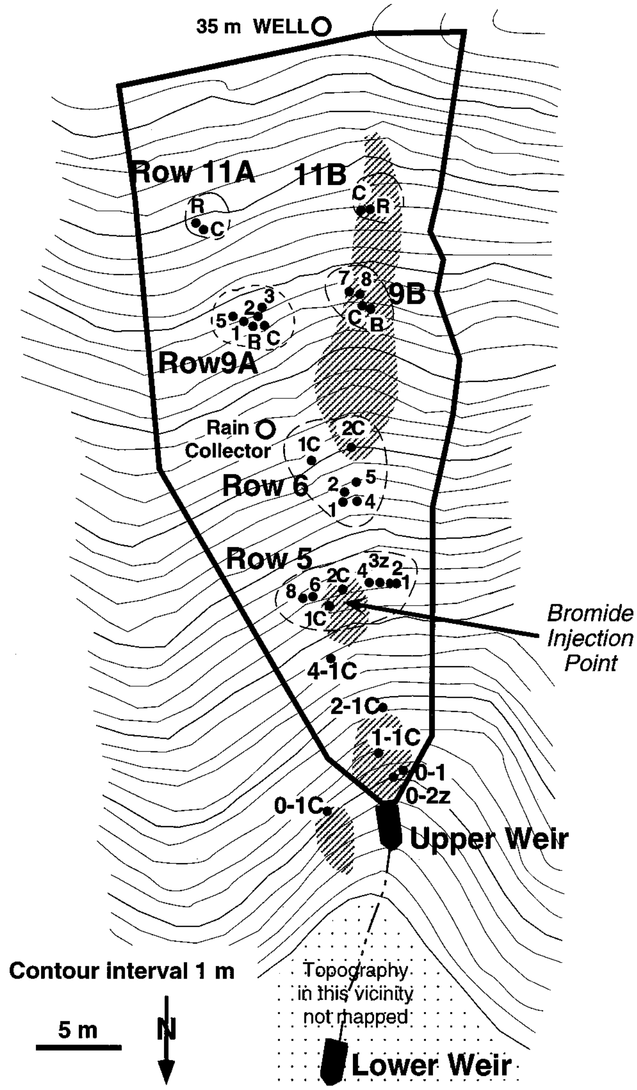

Figure 1. Lysimeter locations in the CB1 catchment. The heavy line outlines the drainage area to the upper weir at the channel head. Dots show lysimeter locations and numbers (see Table 1). Dashed lines enclose nests of lysimeters used to draw profiles in subsequent figures. Hatching shows the extent of subsurface saturation within the colluvium during sprinkling experiment 1 [*Montgomery et al.*, 1997] and is probably representative of the extent of saturation during sprinkling experiment 3. Topography was not fully mapped between the upper and lower weirs.

Suction to the tension lysimeters was regulated at the vacuum pump to 8.5 kPa. This suction will sample most readily water from mesopores [*Luxmoore*, 1981], which constitute most of the pore volume in the CB1 colluvium (R. Torres et al., submitted manuscript). The temporary head field it establishes around the lysimeters should, however, produce a gradient that drives water flow in all pore sizes. Sufficient sample (50–500 mL) could be obtained from most lysimeters in 6–20 hours. During the sprinkling experiments, samples were collected daily after 8–10 hours of pumping. Three lysimeters (0-1, 1-1C, and 2C on row 6) tended to be more efficient in sample col-

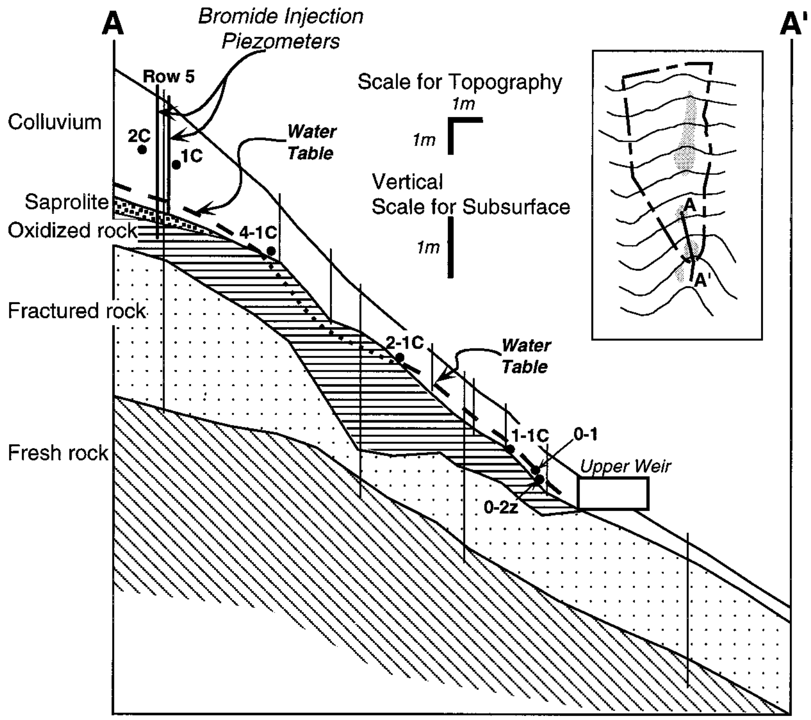

Figure 2. Profile from row 5 to the upper weir (along line shown on inset map) showing topography, geology, and locations of instruments. Lysimeters shown with dots. Piezometers (vertical lines) were used to define the geology shown and to locate the water table in the colluvium, shown for the quasisteady period of experiment 3. Note that lysimeters 4-1C and 2-1C are in unsaturated colluvium at the bedrock interface, downslope of the bromide injection points. The topography is without vertical exaggeration, but the thicknesses of subsurface layers are expanded by 2 times in order to present detail.

lection; these were not opened to vacuum until 1–2 hours before sample collection time. One zero-tension sampler (0-2z) flowed continuously during the experiments and was therefore sampled at 6-hour intervals; the other (3z on row 5) yielded samples only when a wetting front was passing through the colluvium. We found that soil water chemistry, including the tracers discussed here, did not differ in an obvious way between zero-tension and tension samplers. Further details regarding sampling methods and equipment are given by *Anderson* [1995].

3.2. Sprinkling Experiments and Tracer Applications

Four whole-catchment sprinkling experiments have been conducted in CB1. Tracer experiments during the last two of these will be discussed in this paper. Experiment 3 consisted of a 7-day rain application at 1.65 ± 0.20 mm hr^{-1} in May and June 1992, for a total of 275 ± 33 mm of rain (Figure 3a). All of the water in experiment 3 was demineralized as part of a geochemical experiment before entering the sprinkler feed lines [*Anderson*, 1995; *Anderson et al*., 1997]. Experiment 4 was a 21-hour sprinkler run with untreated water conducted 8 days after experiment 3 and which was followed 18 hours later by a natural storm in which 15 mm of rain fell in 3 hours. Our primary focus is on tracer studies during experiment 3, but interesting behavior was observed by continuing to monitor through experiment 4.

Discharge at the two weirs leveled off after three days of sprinkling (Figure 3a), except for a daily oscillation due to evapotranspiration and wind-induced variations in the rainfall rate. During the 4-day period of quasi-steady discharge, rainfall was nearly balanced by runoff: The sum of the discharge at the two weirs equaled 80% of the rainfall less the estimated evapotranspiration (T. Giambelluca, personal communication, 1992). The remaining 20% is attributed to both deep leakage and errors in measurements. Very little water (at most 0.3 mm hr^{-1}), therefore, entered a deep groundwater system.

Two types of experiments were designed to trace vadose zone and saturated flow. The vadose zone was probed with a deuterium spike in the rainwater. Five kilograms of 99.8 at. % deuterium oxide was fed into the sprinkler line with a chemical dosing pump over a 2-day period (Figure 3b). Flow through the sprinkler pipes was turbulent, insuring that the tracer was well mixed by the time it reached the sprinkler nozzles. The dosing pump rate probably varied owing to changes in back-pressure and increased significantly after 1 day of application, resulting in a two-stepped spike with additional, smaller temporal variations.

We calculate, from mass balance, the fraction of old (non-spiked) water contributing to the runoff:

$$\frac{Q_{\text{old}}}{Q_{\text{runoff}}} = \frac{C_{\text{runoff}} - C_{\text{new}}}{C_{\text{old}} - C_{\text{new}}}. \tag{1}$$

where Qs are discharge and Cs are δD compositions for the water reservoir indicated with the subscripts. A similar expression may be derived for the fraction of new (spiked) water. This mixing model hydrograph separation depends on two key assumptions well met in the tracer deuterium spiking experiment: that runoff derives from only two types of water and that these can be identified and characterized.

Samples for deuterium analysis were stored up to 17 months at 4°C in filled, sealed polypropylene vials before analysis. The water was converted to hydrogen gas by zinc reduction [*Ven-*

nemann and O'Neil, 1993] and analyzed on a Prism Series II isotope ratio mass spectrometer. Deuterium analyses are reported as δD values relative to Vienna Standard Mean Ocean Water (VSMOW), where

$$\delta D = \left(\frac{(D/H)_{\mathrm{sample}} - (D/H)_{\mathrm{VSMOW}}}{(D/H)_{\mathrm{VSMOW}}} \right) \times 1000‰ \qquad (2)$$

and D/H is the ratio of deuterium to hydrogen atoms. Replicate analyses of 67 samples differed by 2.0 ± 1.4‰, yielding an analytical error of ±1‰.

Sodium bromide injections were used to trace water flow in saturated materials. A measured quantity of sodium bromide was mixed with a minimum of water (0.5 kg in about 2 L), and the resulting solution was poured into a piezometer. The two piezometers chosen for injection were seated in saturated colluvium and in the underlying bedrock at row 5, located in the hollow axis 19 m upslope from the upper weir (Figures 1 and 2). The injections were spaced 1 day apart, with the second injection occurring more than 20 hours after the peak arrival had passed the weirs from the first injection. Manual water sampling frequency at the two weirs was increased following injections. The samples were analyzed on a Dionex 2000i ion chromatograph with an analytical precision of 0.2 mg L^{-1}.

Table 1. Lysimeter Depths and Types

Lysimeter Name	Design	Location*	Depth, m
0-1C	ceramic cup	C/B	0.47
0-2z	zero-tension	C	0.52
0-1	plate	C	0.44
1-1C	ceramic cup	C/B	0.57
2-1C	ceramic cup	C/B	0.64
4-1C	ceramic cup	C/B	1.04
	Row 5		
1	plate	C	0.73
2	plate	C	0.21
3z	zero-tension	C	0.35
4	plate	C	0.53
6	plate	C	0.85
8	plate	C	0.26
1C	ceramic cup	C	0.87
2C	ceramic cup	C/B	1.12
	Row 6		
1	plate	C	0.56
2	plate	C	0.16
4	plate	C	0.45
5	plate	C	0.38
1C	ceramic cup	C/B	1.56
2C	ceramic cup	C/B	1.45
	Row 9A		
1	plate	C	0.34
2	plate	C	0.49
3	plate	C	0.16
5	plate	C	0.26
C	ceramic cup	C/B	1.10
R	ceramic cup	S	1.40
	Row 9B		
7	plate	C	0.54
8	plate	C	0.29
C	ceramic cup	C/B	0.64
R	ceramic cup	S	1.29
	Row 11A		
C	ceramic cup	C/B	1.27
R	ceramic cup	S	1.36
	Row 11B		
C	ceramic cup	C/B	1.61
R	ceramic cup	S	2.00

*Lysimeter inlet location: C, in colluvium; C/B, at colluvium/bedrock interface; S, in saprolite (weathered rock penetrable with a hand auger).

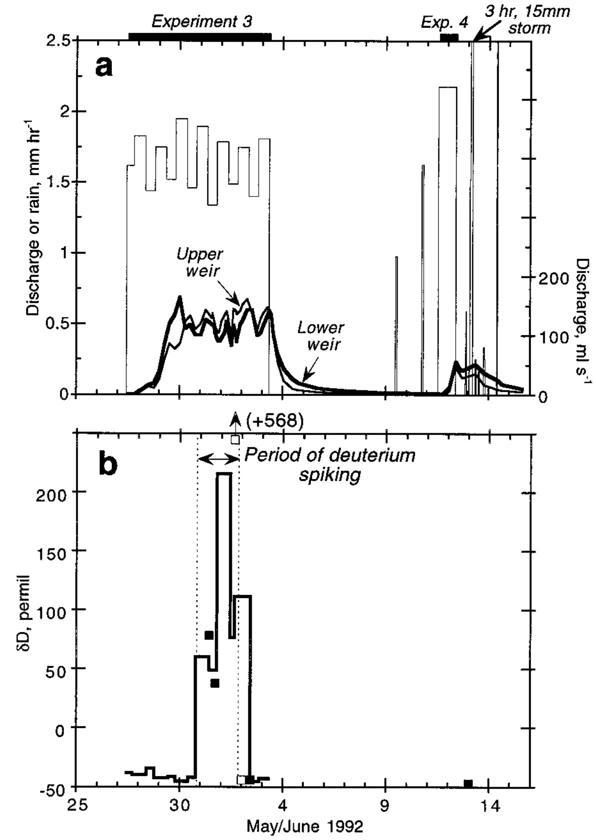

Figure 3. (a) Average rainfall intensity and upper and lower weir hydrographs during sprinkling experiments 3 and 4 and a brief natural storm on June 13. Discharge records shown were measured by hand. Specific discharge is based on a catchment area of 860 m^2. Total discharge from the catchment is the sum from both weirs. Note that the rainfall intensity during the June 13 is off scale; it peaked at 8.5 mm hr^{-1}. (b) The δD of rainfall. The period of addition of labeled water is indicated with the dashed vertical lines. Thick line shows δD of rain collected in time-integrated samples. The filled squares show δD of instantaneously collected samples. Open squares show the calculated +568‰ (spiked) and measured −44‰ (untreated) components of the rain sample collected over the interval spanning the cessation of spiking. The analytical error of ±1‰ for δD measurements is smaller than the line widths and symbols.

3.3. Winter Storm

A storm from February 16 to 22, 1992, brought a total of 144 mm of rain to the study site (Figure 4a). The storm peaked with 50 mm of rain on February 20. Although this 24-hour total has a 1-year recurrence interval at the North Bend rain gauge, located 15 km to the SE, it yielded the highest discharge measured at the site between December 1989 and December 1995 and initiated a debris flow in an adjacent catchment. Importantly, the δD of the rain during the most intense rainfall on February 19 and 20 was sufficiently distinct from base flow to permit isotope hydrograph separation.

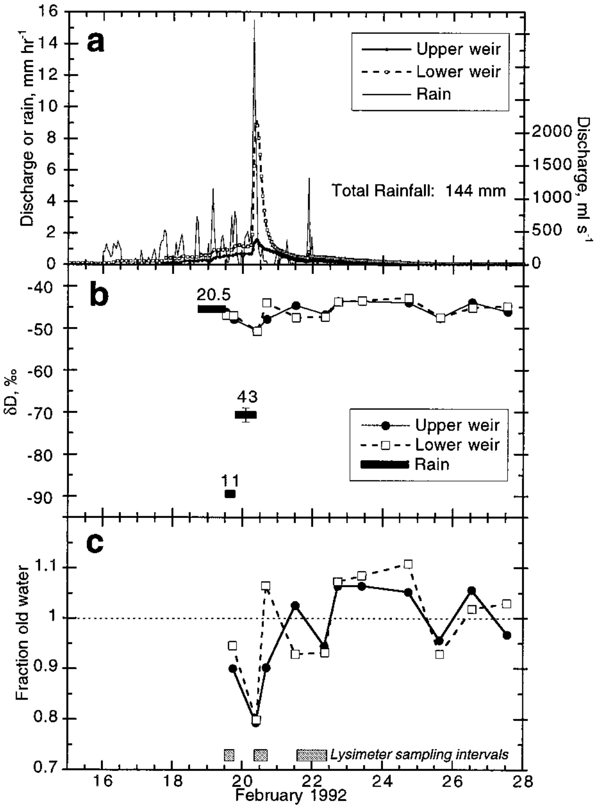

Figure 4. (a) Record of 10 min rainfall intensity and of discharge at the upper and lower weirs during the natural storm of February 16–20, 1992. (b) The δD of the rain and runoff at the weirs. The horizontal lines show the time intervals over which the rain samples were collected; numbers above indicate the mm of rain in the sample. An error bar shows the results of a replicate analysis. (c) Fraction of old water in the runoff at the weirs (equation (1)), calculated using the incremental-intensity mean rain δD [*McDonnell et al.*, 1990] for new water. Old water was defined as the average composition of all the runoff in the record except for the samples collected on February 20. Shaded bars show when lysimeter samples plotted in Figure 5 were collected.

Water sampling at the weirs and lysimeters began on February 19, after 65 mm of rain had fallen but when the total discharge was still low. Samples were collected at the weirs twice daily on February 19 and 20 and once daily thereafter. The lysimeter vacuum system was being assembled for the first time at the onset of this storm; sampling was therefore incomplete on February 19, but full suites were collected subsequently (Figures 4 and 5).

4. Results

4.1. Isotope Hydrology During the Winter Storm

Rainwater during the peak rainfall of the February 16–22 storm had δD values 25–45‰ lower than runoff before or after the discharge peak (Figure 4b). Runoff samples collected less than an hour after peak discharge on February 20 had δD values only 3–4‰ lower than runoff the previous day and 19–38‰ greater than the rainfall. Within 7 hours, runoff at the weirs had recovered to prestorm δD values of −46‰, and hovered around this composition for the next week.

The slight decline in δD values at the weirs during peak discharge can be attributed to a mixture of 20% new water and 80% old water (Figure 4c). Alternatively, the variation in δD of runoff may reflect a greater contribution to runoff from shallower parts of the soil during peak discharge. Soil water profiles showed persistent gradients of increasing δD with depth, but most samples had δD values lower than runoff at the weir, and higher than the most intense rainfall (Figure 5). The isotopic composition of base flow runoff at the weirs resembled the deepest soil waters; the lower δD of peak flow may have resulted from a shift in the locus of runoff sources toward shallower depths. This interpretation does not require any new water contribution to the peak runoff. Although the data are scanty, the response of soil water to the low-δD rainfall was slight, and primarily confined to the shallowest (<0.3 m) lysimeters. The δD of soil water below 0.3 m is sufficiently low to explain the shift in the runoff.

It is interesting to note that the lysimeters sampling water from saturated colluvium near the upper weir (filled symbols in Figure 5) tended to have higher δD values than vadose zone waters at comparable depths elsewhere in the catchment. This area of saturation near the weir exhibits exfiltrating head gra-

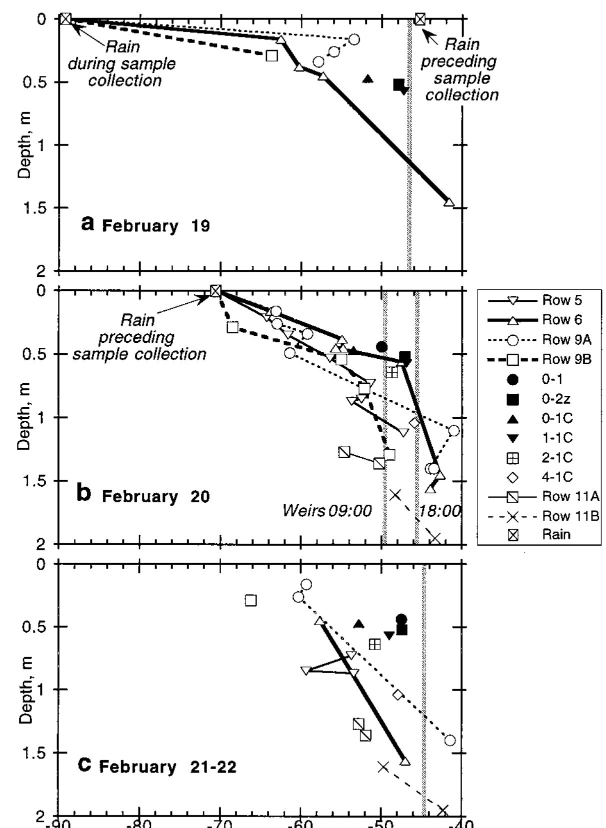

Figure 5. The δD of water from lysimeters on (a) February 19, (b) February 20, and (c) February 21 and 22, 1992, collected during the intervals indicated on Figure 4c. Filled symbols indicate samplers located in frequently saturated colluvium near the channel head. Data points at depth 0 are values for rainfall either during or just preceding lysimeter sample collection. Vertical gray lines indicate the δD of the runoff during lysimeter sample collection. On February 20 two weir samples were collected: at 09:00, ~30 min before peak discharge, and 9 hours later. The lysimeter samples were collected in the intervening time.

dients under some conditions [*Montgomery et al.*, 1997] and is therefore a zone where shallow percolating soil water can mix with deeper exfiltrating water.

The lysimeter isotope data show that during this winter storm, which produced the highest recorded discharge in 5 years and caused a debris flow nearby, new rainwater apparently did not percolate to soil depths greater than 0.3 m within 2 days. The runoff composition shifted slightly toward the low δD of the most intense rainfall at the peak discharge and then returned to its prestorm value. Two explanations for this shift are plausible: (1) New rainwater traveled quickly via preferential flow paths, not sampled by our lysimeters, through the vadose zone, into the saturated zone, and out through the weir to influence the runoff δD. Cessation of preferential flow after the rain stopped would then explain the return to prestorm runoff composition. (2) Alternatively, nonpreferential flow of lower-δD water from the soil during peak discharge could be responsible for the shift. This mechanism is allowed by the structure of the soil water δD composition, wherein δD values decreased towards the soil surface. If the high-intensity period in the storm increased shallow soil water contributions to runoff, this could have produced the shift in the runoff composition without involving new water. Indeed, a prediction of this mechanism is that runoff will shift towards the composition of soil water during storms rather than towards the composition of the rain. Conditions allowed a test of this prediction in our deuterium-spiked rainfall experiment.

4.2. Tracer Experiment: Flow in the Saturated Zone

Two bromide injections were conducted during sprinkler experiment 3 to trace saturated flow. The first injection was into a piezometer with its inlet at the bedrock/colluvium interface at 1.9 m depth. A hydrostatic to slightly infiltrating head gradient existed in the saturated soil at this location.

Bromide arrived at the upper weir in 16 min, and the peak concentration was observed 55 min after the injection (Figure 6a). The average saturated zone flow velocity along the slope-parallel distance calculated from the peak arrival time is 6×10^{-3} m s^{-1}. Bromide also appeared at the lower weir. Because soil water flow was funneled into the upper weir flume by plastic-lined wing walls sealed to bedrock and because flow from the upper weir was diverted in a pipe around the lower weir, the only route for bromide to reach the lower weir was through the bedrock. At the lower weir the first arrival and peak arrival times were 38 and 180 minutes, respectively, yielding an average flow velocity of 4×10^{-3} m s^{-1}. If, however, subsurface flow emerged in the vicinity of the channel head but below the upper weir, the velocity to the lower weir may have been as low as 2×10^{-3} m s^{-1}, assuming that surface flow between the exfiltration point and the lower weir takes no time. This, therefore, is a conservative lower bound for the average velocity of flow through bedrock.

The magnitude of the bromide peak differed by a factor of more than 4 between the weirs (~110 mg L^{-1} at the upper weir versus 24 mg L^{-1} Br at the lower weir). Because discharge at the upper weir slightly exceeded that at the lower weir, the mass flux peak at the upper weir was 5 times higher than at the lower weir. After 24 hours a total of 0.25 kg of bromide (64%) had been recovered, of which two thirds passed through the upper weir.

The second bromide injection, a day after the first, was into a piezometer drilled 0.87 m into the weathered bedrock. This piezometer is 0.65 m from the piezometer used for the earlier injection. The head gradient was hydrostatic to slightly infiltrating.

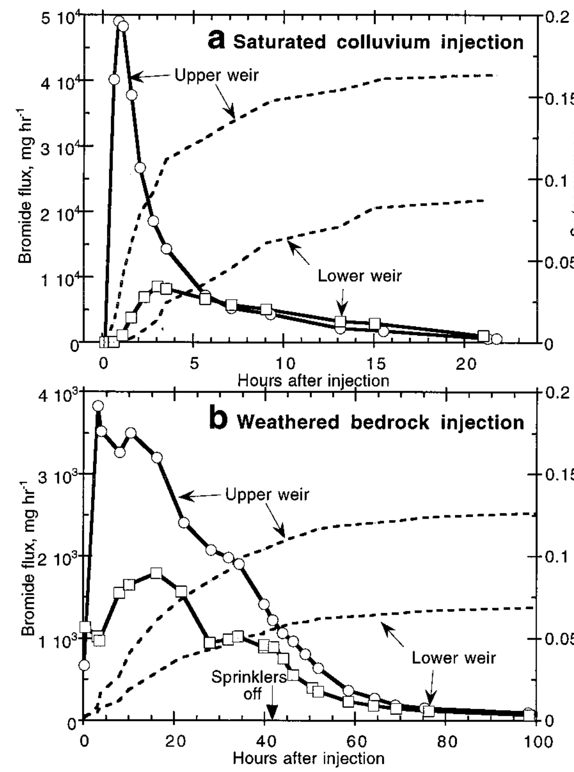

Figure 6. Instantaneous mass flux and total mass flux at the weirs following bromide injections. (a) Twenty-four-hour period following injection into saturated colluvium. (b) One-hundred-hour period following injection into weathered bedrock in a piezometer located 0.65 m from the one used in the earlier experiment. Note the difference in scales between Figures 6a and 6b.

The difference between injection into weathered bedrock and colluvium was immediately obvious: the solution backed up in the piezometer pipe and drained slowly out. This impromptu falling-head test yielded a bedrock saturated hydraulic conductivity of 2×10^{-7} m s^{-1} (Hvorslev method, in work by *Freeze and Cherry* [1979]), comparable to previous falling head tests in weathered bedrock at CB1 [*Montgomery et al.*, 1997]. Despite this low conductivity, bromide appeared at both weirs relatively soon after injection (Figure 6b). The peak arrival at the upper weir was 3 hours after injection began, while the peak at the lower weir took 10 hours. These times yield velocities of 2×10^{-3} m s^{-1} and 1×10^{-3} m s^{-1} to the upper and lower weirs, respectively, which are comparable to the lower bound for flow velocity through bedrock derived from the first injection experiment. The concentrations and mass fluxes for the bedrock injection were considerably lower than those for the saturated colluvium injection, reflecting the slow injection and greater dispersion for flow through bedrock. After 100 hours, only 50% of the bromide injected had been recovered at the two weirs. The lower recovery is in part attributable to lower discharge rates, as the sprinklers were turned off about 40 hours after the injection. Again, nearly two thirds of the recovered bromide passed through the upper weir.

The asymmetrical shape of the bromide peaks with a long tail on the right (Figure 6) is characteristic of transport that occurs via several pathways having different velocities [*Sposito*

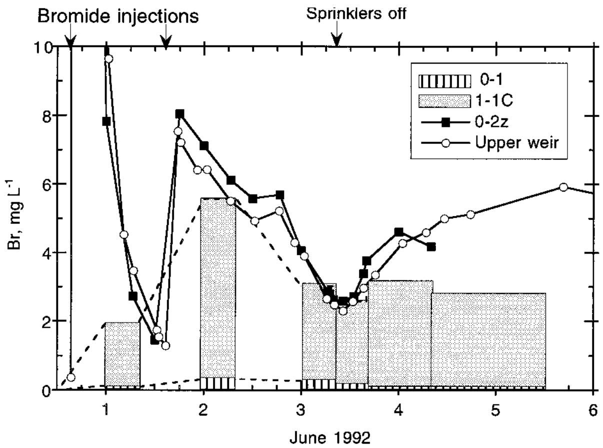

Figure 7. Bromide concentrations in samples from lysimeters located in the saturated colluvium near the upper weir following the bromide injections. Samples from 0-2z and the upper weir represent instantaneous concentrations in the soil water and runoff, respectively, while samples from 0-1 and 1-1C were collected over time intervals indicated by the width of the bars. Times of bromide injections are indicated at the top of the figure. The peak concentrations at the upper weir and 0-2z following the first injection are off scale.

and Jury, 1988]. Flow and dispersion through a porous medium characterized by only a single velocity would give rise to a symmetrical peak; the long tail is due to transport along a slower pathway. The broader curve shape at the lower weir (Figure 6a), and for the saturated bedrock injection (Figure 6b), suggests either multiple flow paths, or two pathways, one fast, one slow, of comparable importance in the total flux.

The low bedrock saturated hydraulic conductivity ($\sim 10^{-5}$ to 10^{-7} m s^{-1}) compared with the high velocity of the bromide peak ($\sim 10^{-3}$ m s^{-1}) to the weirs, and the shapes of the bromide flux curves (Figure 6), suggest that most bedrock flow is conducted along open fractures such as those observed in the upper third of the 35 m core obtained at the ridge crest. The bulk conductivity of the fractured bedrock is therefore best characterized with the velocity of the bromide peak, which averages over a large volume of rock, rather than from falling-head tests in piezometers. The conductivity should vary spatially with fracture density and tightness and therefore should decrease with depth in the bedrock. This could in part explain the lower velocities measured in the second injection (although the noninstantaneous injection also played a role). The similarity of the partitioning of bromide total mass flux in both experiments suggests that both injections followed similar flow paths once the fracture system was entered. The unequal bromide partitioning between the two weirs, despite nearly equal water discharges, probably reflects the importance of recharge from further upslope to the lower weir.

Bromide appeared in some of the soil lysimeters located between the injection points and the weirs. The pattern of bromide appearance in these lysimeters demonstrates a flow path crossing between bedrock and colluvium. Bromide appeared only in those lysimeters located within the patch of saturation above the upper weir (samplers 1-1C, 0-1, and 0-2z; see Figure 2). Significantly, no bromide was detected in lysimeters 2-1C and 4-1C despite their locations at the colluvium/bedrock interface downslope of the injection point. No bromide appeared in lysimeter 0-1C, located in the patch of saturation outside the wing wall (Figure 1), indicating that lateral dispersion of the tracer was not large.

These observations show that the hydrologic connection between colluvium and bedrock under saturated conditions is strong. We found in earlier monitoring and sprinkling experiments that infiltrating and exfiltrating gradients across this interface exist [*Montgomery et al.*, 1997]. The bromide data show that water emerges from the bedrock into the saturated colluvium above the weir even when our piezometers do not show exfiltrating gradients and, moreover, that this flow is rapid.

Mixing within the saturated zone is not uniform, as is evident from the bromide concentrations there (Figure 7). Sampler 0-2z tracked closely the concentrations measured at the upper weir. Sampler 1-1C showed concentrations slightly lower than the weir, and sampler 0-1, the shallowest in the saturated colluvium, had very low bromide concentrations. The concentration gradients implied by Figure 7 show that the saturated colluvium above the upper weir cannot be characterized as a homogeneous water body; rather, it is a zone of strong mixing of water from different flow pathways.

4.3. Tracer Experiment: Flow in the Vadose Zone

Vadose zone flow was traced with deuterium-spiked rain. The deuterium spike increased the δD of the rainwater during sprinkler experiment 3 by at least 100‰. Untreated sprinkler water δD averaged −41‰ before the spiking; the volume-weighted mean spiked rain δD was +155‰ over the 2-day application. On the first day the volume-weighted mean δD of the rain was +54‰. The mean stepped up on the second day to +255‰, with fluctuations from +76‰ to possibly as high as +559‰ (Figure 3b). Some ambiguity exists regarding the latter value. It is based on a precipitation sample with a δD value of +111‰ that spanned the final 4.75 hours of deuterium-spiked rain as well as 14 hours of untreated rain. The calculated value of +559‰ is the spiked component for a 4.75:14 mixture with untreated sprinkler water with a δD of −41‰.

4.3.1. Lysimeter observations. Under the steady conditions of the sprinkling experiment, deuterium-labeled water moved through the vadose zone in the soil in a plug-like manner. Two days of deuterium-spiked rain on May 31 and June 1 caused two days of peak δD values in the shallowest lysimeters at row 5 on June 2 and 3 but did not produce substantial changes in deeper lysimeters until June 4 (Figure 8). Profiles of δD in all the lysimeters (Figure 9) show that this behavior was repeated elsewhere in the catchment. An exception to the general pattern was seen in the row 9A profile, where the deuterium peak at shallow depths was smaller and occurred a day later than elsewhere. Nowhere is there evidence of the rapid arrival of labeled water at depth that one might expect from preferential or bypass flow.

Only the shallowest lysimeters (<0.5 m) showed significant increases in δD value (Figure 9), and only two of these passed through a maximum before the termination of sprinkling. Nonetheless, if we assume that the highest δD value obtained in these shallow lysimeters marks the passage of the peak from spiking, we can calculate the velocity of spiked water to these depths. As most of the maximum δD values are greater than the mean of the rain on the first day of spiking (+54‰), we use the beginning of the second step in the rainwater δD to calculate the travel time, Δt, to the lysimeter depth, d. The average linear velocity, v, [*Freeze and Cherry*, 1979] calculated in this way is reasonably well predicted by

$$v \equiv \frac{d}{\Delta t} = \frac{I}{\theta_v}, \quad (3)$$

where I is the rainfall intensity and θ_v is the volumetric water content (Table 2 and Figure 10). We use θ_v of 0.34 ± 0.06 (n = 16) measured with TDR in the top 0.3 m of soil during steady conditions of the experiment, and rainfall intensity measured in the three wedge gauges closest to each lysimeter.

The success of (3) in predicting observed velocities for the deuterium peak ($d/\Delta t$) suggests that the vadose zone water is largely mobile. If a substantial fraction of soil water was immobile during steady flow conditions, then the measured water content would produce an underestimate of velocity in (3).

Although the velocity appears to be well represented with a plug-flow model, the peak δD does not. With one exception, the maximum observed δD values were smaller than the overall volume-weighted mean δD of the spiked rain (Table 2), and all were considerably lower than the δD of the second day step in the spike. Two mechanisms could be operating to produce the peak reduction. First, the spike was attenuated by dispersion as it moved through the soil, as is shown by the several-day rise of soil water δD in response to an instantaneous step in rainwater δD (Figure 8). This implies that dispersion mixes new and old water in the vadose zone. It is worth noting here that the lysimeter samples integrate soil water compositions over the 8- to 10-hour interval that the vacuum pump operated each day. The sample integration time is smaller, however, than the duration of the two steps of spiked rain application and therefore should not strongly interfere with observation of the peak soil water δD. Second, it is also possible that the deuterium peak did not arrive before the sprinklers were turned off, and true water velocities are therefore lower than suggested by (3) and Figure 10. Many lysimeters reached their maximum δD values just before the sprinklers were turned off (Figure 9e), supporting this argument.

4.3.2. Soil water remobilization. A few samples emerged passively (without suction) in plate lysimeters on June 13, 10 days after experiment 3 (Table 3). We believe the water flowed into our samplers during a 3-hour, 15 mm storm in the early morning hours of June 13, but it is possible that some of the water appeared during experiment 4 on June 11 and 12. All of the samples were from less than 0.45 m depth. Their δD values were comparable to the final soil water from experiment 3, and wholly distinct from the −47‰ rain water (Figure 9h).

The soil volumetric water content during the storm was probably between 0.2 and 0.34, the drained and experiment 3 steady state water contents, respectively (R. Torres et al., submitted manuscript). Simple displacement by plug flow of this soil water by 60 mm of rain (experiment 4 plus storm total) would bring new water to a depth of 0.17–0.30 m, which is indeed the depth interval from which most of the water samples were produced. Water depth-equivalents, obtained by dividing the sample volume by the cross-sectional area of the sampler, decline with depth in the soil and are intriguingly close, for shallow depths, to the 15 mm rainfall total from the storm (Table 3).

Where did this water come from? As the samples were collected without suction, they represent water that was mobile and had sufficient potential to enter our lysimeters. The high δD values of these samples demonstrates that mobile water in the shallow soil was old, rather than new, or event, water.

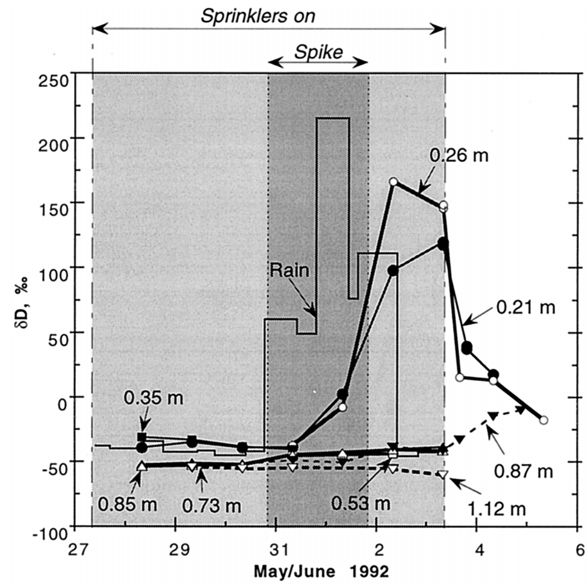

Figure 8. The δD through time of the precipitation and water from the row 5 nest of lysimeters (see Figure 1 for location). Dark shading shows the time of application of deuterium-spiked rain water; δD of the rainwater in time-integrating samples is indicated with thin stepped line. Light shading shows duration of sprinkling. Depths of each lysimeter are indicated on the plot. The water table at this location was 1.7 m below the surface from May 30 through early June 3, 1992.

Although the hydrometric conditions that allowed water to enter our samplers appeared to occur during the higher rain intensities of the storm (rather than low intensity experiment 4), the water so mobilized came out of storage, rather than being preferential flow of new water. Furthermore, the similarity in δD of the June 13 samples to the last soil water samples collected on June 4 implies that soil water flow virtually halted after June 4. Although the δD of one lysimeter fell from +81‰ to +30‰, the remainder showed an average decline in δD of 4.1‰ between June 4 and June 13. It appears therefore that water in the shallow soil was relatively immobile following the rapid drainage at the end of experiment 3 and was remobilized during the subsequent storm.

4.3.3. Runoff and runoff sources. Given the clear evidence of high deuterium concentrations in the upper part of the soil column at the conclusion of the sprinkling experiment, the low deuterium content of runoff at the weirs (Figure 11) is no surprise. The two-component mixing model, using average runoff compositions before the spiking experiment as the old, or nonspiked, component and the volume-weighted mean spiked rainfall composition of +155‰ as the new, or spiked, component yields a minimum of 10% spiked water at the upper weir and 5% spiked water at the lower weir at the end of experiment 3 (Figure 11). Slightly larger contributions from the spike to the runoff appeared during experiment 4, but it is important to note that the spike is then no longer new.

The first shift in δD at the upper weir began approximately 28 hours after the start of the spiking and increased steadily thereafter. Lysimeter samples from the time of first arrival at the upper weir show the labeled water reaching depths of up to 0.4 m in the soil. As the unsaturated zone was also roughly 0.4 m thick just upslope from the upper weir during experiment 3 (Figure 2), it is likely that percolation of the spike through

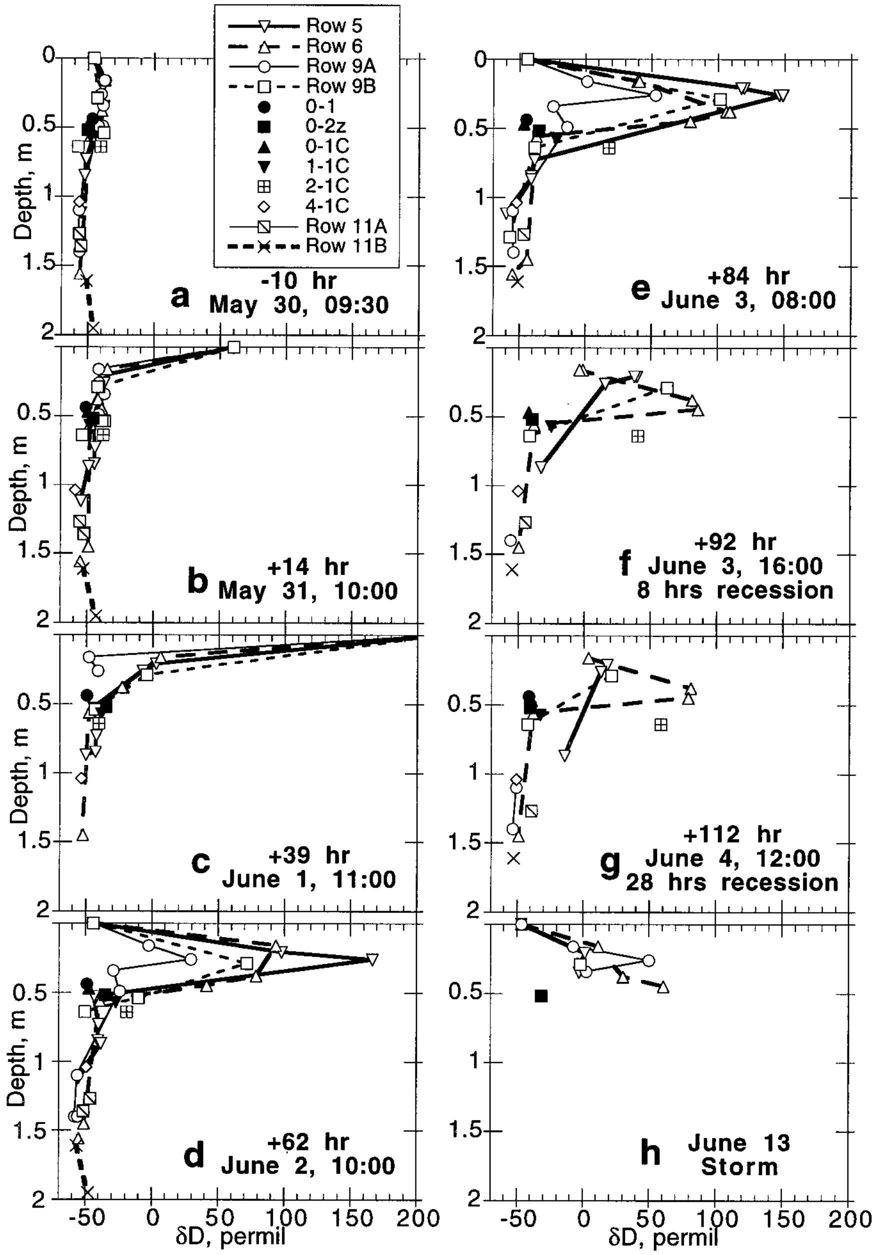

Figure 9. Profiles of δD in the soil at different times during experiments 3 and 4. The δD value at the surface is that of the rainfall during the time suction was applied. The hours relative to the beginning of the deuterium application and the date and time of sample collection are indicated on each panel. The sprinklers were turned off after the June 3, 08:00 sample collection (Figure 9e). Suction was applied to lysimeters beginning at midnight for all collections, except the June 4 collection (Figure 9g), when the suction began at 19:00 on June 3, and the June 13 samples (Figure 9h), which were collected without applied suction.

the vadose zone to the saturated colluvium near the weir was sufficient to cause the δD shift in the runoff. Because of the greater thickness of the vadose zone elsewhere in the catchment, spiked water had not reached the patches of saturated colluvium higher on the slope, and therefore the increase in δD at the weir reflects only contributions from the saturated area immediately upslope of the upper weir.

This mechanism of local saturated area control on runoff is further revealed in Figure 12. The four panels show lysimeter and upper weir δD values through the course of the sprinkling experiment and for about 2 days after. Four distinct responses are apparent: (1) shallow lysimeters show a significant rise in δD during the experiment followed by sharp declines due to the incursion of unlabeled water from the last day of sprinkling (Figure 12a); (2) some intermediate depth lysimeters also rose rapidly, to not as great a value, but after the end of irrigation their concentrations did not decline (Figure 12b); (3) deep lysimeters in the colluvium either responded after the experiment or not at all (Figure 12c); and (4) saturated zone lysimeters near the channel head rose a small amount during irrigation and declined slightly at the end of the experiment, and this pattern matched that of the upper weir (Figure 12d). Together these data show that the spiked water did not reach deeply in the subsurface flow field. The runoff signature is the

Table 2. Velocities of Deuterium Peak to Lysimeters

Lysimeter	Peak δD, ‰	Depth, m	Δt, Observed, Hour	Velocity of Peak,* Observed, mm hr^{-1}	Local Rainfall, mm hr^{-1}	Velocity of Peak,† Calculated, mm hr^{-1}
				Row 5		
2	118	0.21	56	3.8 ± 0.3	2.26 ± 0.41	6.6 ± 1.7
8	166	0.26	33	7.9 ± 1.4	2.26 ± 0.41	6.6 ± 1.7
				Row 6		
2	93	0.16	33	4.8 ± 0.9	2.24 ± 0.31	6.6 ± 1.5
5	107	0.38	56	6.8 ± 0.5	2.24 ± 0.31	6.6 ± 1.5
4	85	0.45	64	7.0 ± 0.5	2.24 ± 0.31	6.6 ± 1.5
				Row 9A		
3	1	0.16	80	2.0 ± 0.1	1.69 ± 0.23	5.0 ± 1.1
5	53	0.26	80	3.3 ± 0.2	1.69 ± 0.23	5.0 ± 1.1
				Row 9B		
8	101	0.29	56	5.2 ± 0.4	1.56 ± 0.24	4.6 ± 1.1

*Average linear velocity determined from sample depth and travel time.
†Calculated with equation (3) and $\theta_v = 0.34 \pm 0.06$.

same as that of the saturated zone around the upper weir, and this signature differs from vadose zone waters upslope, implying that local mixing in the saturated zone is responsible for the runoff signature. This mixing is a combination of the high velocity return flow of unspiked deep bedrock water to the colluvium and the vertical percolation of spiked vadose zone water, with the bedrock flow dominating.

Changes in the proportions of these flow paths under non-steady conditions can be seen in the variations of runoff δD after sprinkling experiment 3. The δD declined at the upper weir during the falling limb of experiment 3, and then surged with the increased discharge during experiment 4 (Figure 11). The falling limb behavior of the weir was mirrored by declines in δD of several lysimeters in saturated colluvium near the weir and again contrasts with the intermediate depth vadose zone lysimeters (Figure 12c). This suggests that the vertical flux of spiked water from the vadose zone decreased faster during the falling limb than did flux of water exfiltrating from the bedrock and that base flow between the experiments was fed largely by unlabeled water exfiltrating from bedrock. More of the deuterium-labeled water stored in the soil was pushed out when the sprinklers were turned on again, producing the rapid rise in δD at the upper weir during experiment 4.

The response to the deuterium spike at the lower weir was of lower amplitude and possibly delayed relative to the upper weir (Figure 11). The δD of runoff at the lower weir began to rise along with the upper weir after 28 hours of spiking but did

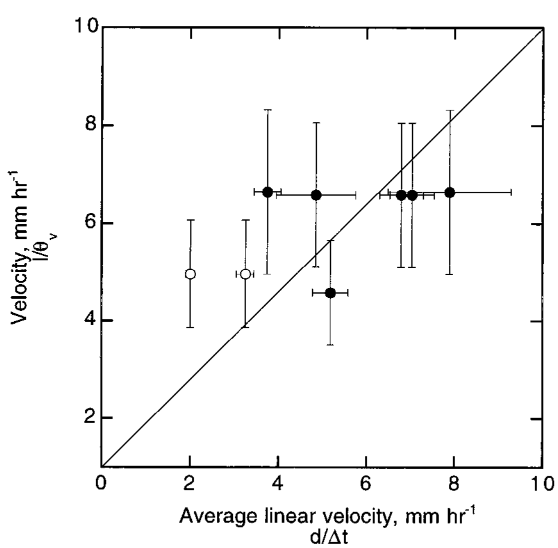

Figure 10. Average linear velocity of the deuterium peak to lysimeters in which a significant δD rise was observed, calculated from travel time versus that calculated from precipitation intensity and volumetric water content of the soil. Solid data points are lysimeters with peak δD values greater than the rain δD on the first day of spiking; for these, Δt is defined as time since the step in rain δD values (20:00 on May 31). Open data points are lysimeters in which maximum δD values did not exceed the spiked rain δD. For these two lysimeters Δt was defined as time since the beginning of spiking. Horizontal error bars arise from uncertainty due to coarse sampling interval (24 hours) and long times integrated by each sample (8–11 hours); they are calculated from the range of times represented by each sample collection. Vertical error bars reflect uncertainties in rainfall rate and volumetric water content.

Table 3. Soil Water Samples Collected on June 13

Lysimeter	Depth, m	Volume, mL	Water Depth Equivalent,* mm	δD, ‰
0-2z	0.52	···	···	−31.6
		Row 5		
2	0.21	100	10	1.5
3	0.35	50	2.5	−2.9
		Row 6		
2	0.16	110	11	11.7
5	0.38	100	10	30.5
4	0.45	50	5	61.0
		Row 9A		
3	0.16	25	2.5	−7.1
5	0.26	50	5	49.8
1	0.34	25	2.5	2.6
		Row 9B		
8	0.29	400	40†	−1.6

All samples collected without vacuum.
*Water depth equivalent is equal to sample volume/sampler area.
†May include some water from experiment 4.

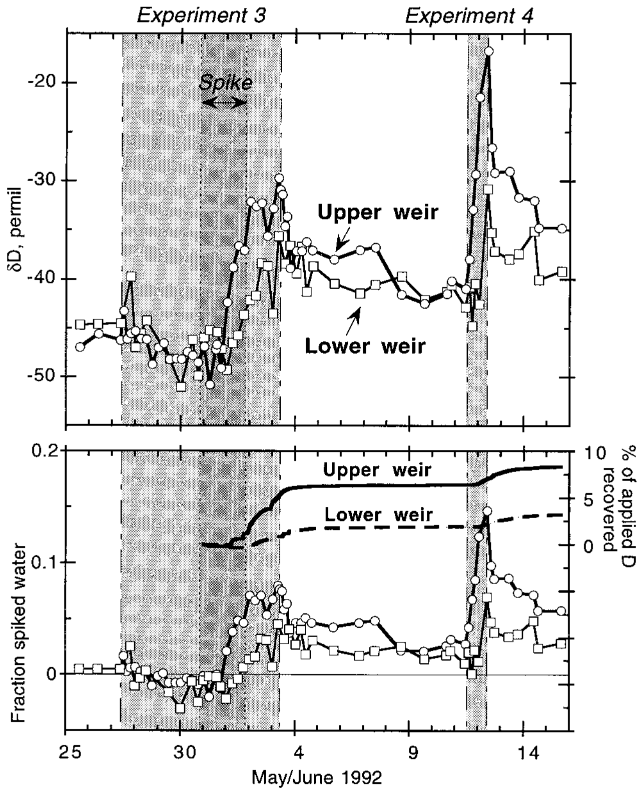

Figure 11. The δD of the runoff at the upper and lower weirs during sprinkling experiments 3 and 4. The bottom panel shows the results of a two component mixing model analysis, using the average runoff composition before the spiking began as old water, and the volume-weighted average rainfall composition during spiking (+155‰) as new water. Also shown is the integrated flux of the spike for each weir, expressed as a fraction of the total deuterium applied. Light shading indicates time of sprinkling; dark shading shows the interval in which rain was spiked with deuterium.

not exceed the background fluctuations in δD for another 18 hours. The δD at the lower weir climbed until the sprinklers were shut off, fell slightly during the falling limb, and then reached a plateau of about −40‰. Sprinkler experiment 4 pushed the lower weir runoff to higher δD values, but with a 13- to 21-hour lag relative to the appearance of higher δD water at the upper weir.

There are two routes for the spiked water to reach the lower weir. It either travels down through the thin soil outside the upper weir walls or passes through bedrock under the upper weir, emerging between the two weirs. Although the sprinkler heads were all inside the catchment boundaries (Figure 1), some water was thrown outside the margins. Thus we cannot rule out spiked sprinkler water traveling down through the soil as a source for labeled water to the lower weir. The labeled water appeared at the lower weir over the same time interval during which δD values started to climb at the base of the colluvium just upslope of the upper weir, as represented by lysimeter 2-1C (Figure 12b). It is possible therefore that the rise in δD at the lower weir is from water that traveled through the bedrock, infiltrating into the rock as far upslope as lysimeter 2-1C. If this is the case, then bedrock flow paths must be highly variable in space and bedrock flow velocities very high, as one might expect in a fractured medium.

By the end of both experiments, after a total of 210 mm of sprinkling and rainfall, 11.4% of the applied deuterium had left the catchment, slightly more than two thirds of the total passing through the upper weir (Figure 11). Because such a small amount of the tracer had traveled through the catchment at the end of monitoring, it is not possible to characterize well the average transit time for water in the catchment [e.g., *Rodhe et al.*, 1996]. A linear extrapolation suggests that it would take about 2 m of precipitation, an amount equal to the mean annual rainfall at CB1, to displace all of the applied deuterium through the catchment. Although this estimate is crude and likely to underestimate the actual transit times of water through the catchment, as the vadose zone flow is thicker on the upper slopes, it is in line with the transit time distributions observed in larger catchments by *Rodhe et al*. [1996] and *Nyström* [1985]. This suggests that the average residence time for water in this steep, 860 m^2 headwater catchment is on the order of 6 months.

5. Discussion

Two important flow paths through the CB1 catchment have been identified: vertical percolation in the vadose zone and rapid saturated flow. These paths meet in a small area of partially saturated colluvium near the channel head, merging to form the discharge from the catchment. In effect, this area near the channel head is the variable source area [*Hewlett and Hibbert*, 1967] of the CB1 catchment. This differs from the usual conception of a variable source area in that it is defined by a subsurface-saturated area rather than surface saturation, but it appears to be comparable in its importance in controlling the sources of runoff. Our tracer data show that this region is where relatively short residence time percolating water in the vadose zone mixes with high-velocity but long residence time water emerging from the bedrock. The dynamics of this zone have important control on runoff chemistry, and the proportions of old and new water in runoff.

In the natural winter storm of February 1992 it was impossible to distinguish between release of stored old water and preferential flow of new water as sources of the peak storm runoff, because the water in the colluvium and the rainfall were both lower in δD than the base flow (Figure 5). This ambiguity was removed, however, during the precipitation events after our application of deuterium-spiked water to the catchment. In sprinkling experiment 4 and the brief storm that followed it, much of the water in the vadose zone had δD values ≫0, but the sprinkler and rain water had low (approximately −47‰) δD values. In this situation the shift in the runoff δD to more positive values can be attributed unambiguously to release of spiked vadose zone water by the rain (cf. Figures 11 and 9). The water that was mobile in the colluvium during this time was old, spiked water (Figure 9h), arguing against preferential flow of new water contributing to runoff during sprinkling experiment 4 and the storm after it. The absence of preferential flow in other forest soils [e.g., *Rawlins et al*., 1997] has been attributed to low rainfall intensities typical of these environments, and this may explain the CB1 observations as well. Thus, under typical conditions, flow through the vadose zone is best characterized as plug flow, in which old water is pushed out at the base of the soil column by incoming new water.

The bromide injections into saturated rock and colluvium in a midslope location showed that there is a high velocity saturated flow path through the bedrock between partially satu-

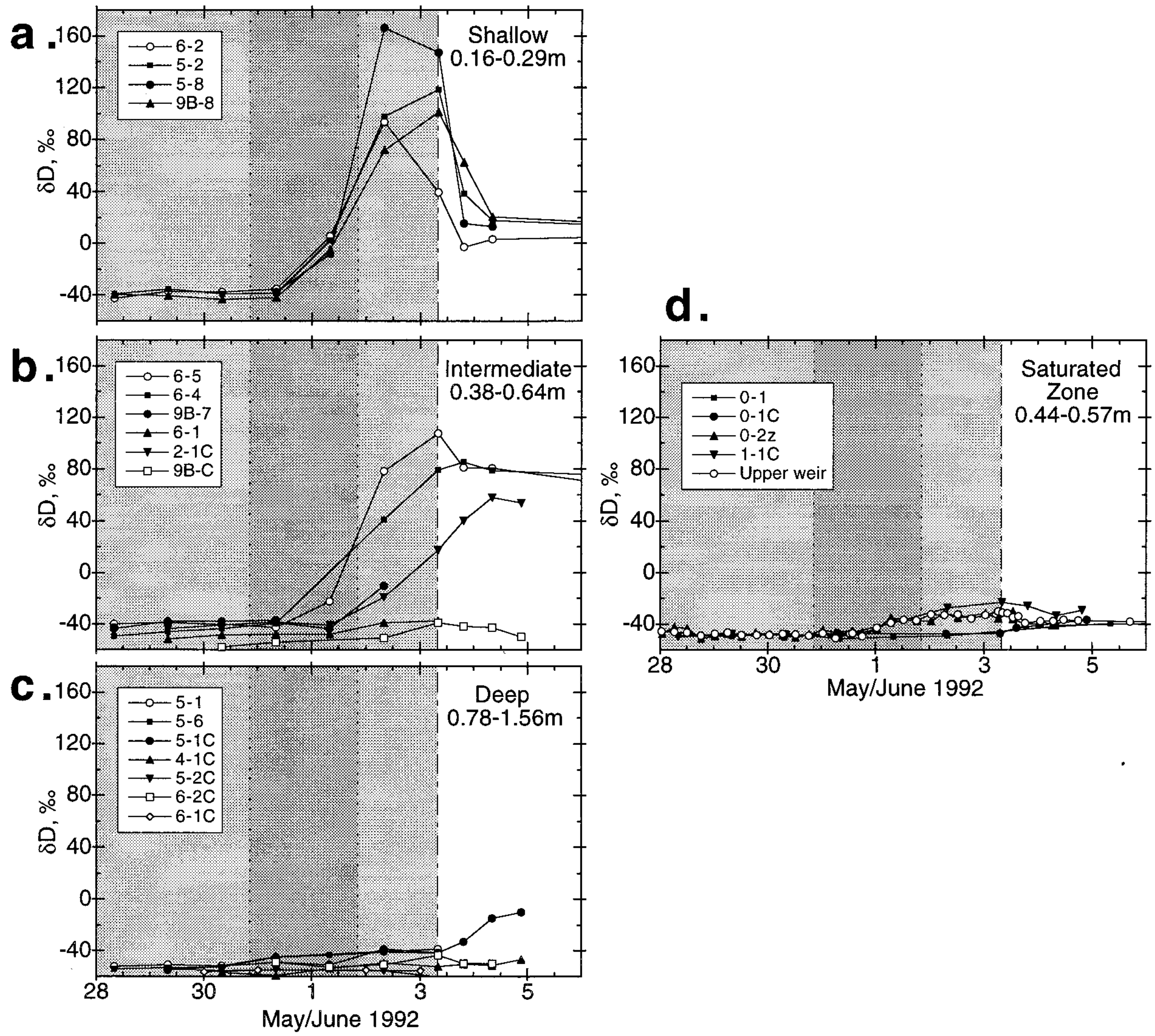

Figure 12. Comparisons of temporal δD records of the vadose zone in (a) shallow, (b) intermediate, and (c) deep lysimeters with (d) the δD of runoff at the upper weir and in saturated colluvium near the upper weir. All the saturated colluvium lysimeters are in the same depth interval as the intermediate depth vadose zone lysimeters. Lysimeters are ordered by increasing depth in colluvium in the legends. Note that scales are the same. Shading as in Figure 11.

rated patches in the colluvium. We now return to the bromide data to illustrate how water fluxes from bedrock varied with changing discharge. Bromide concentration increased in the runoff through the falling limb of experiment 3 and then showed sharp declines as discharge increased in experiment 4 (Figure 13). This pattern demonstrates that the flow path bringing water out of the bedrock, into the partially saturated colluvium near the weir, and then into runoff, contributes a greater proportion of the total discharge as the discharge declines. The flux from this pathway varies directly with the discharge (as indicated by the falling and rising bromide flux during the falling and rising limbs of experiments 3 and 4), but the relative contribution of this flow path to runoff varies inversely with discharge. Taken together, the bromide concentrations (Figure 13) and the deuterium data (Figure 11) show that increasing discharge yields an increasing contribution from the vadose zone near the weir and a consequent decline in the relative contribution from the saturated flow path emerging from the bedrock. Falling discharge reverses the pattern. At very low discharges, such as those that occurred between experiments 3 and 4, all saturation within the colluvium disappears [*Montgomery et al.*, 1997]; nonetheless, the deeper flow path predominates over vadose zone percolation.

The runoff is extremely sensitive to changes in these flow path contributions, as shown by the synchronicity and opposing directions of the concentration trends of both the bromide and deuterium tracers. The deeper bedrock flow merges with percolating water from the vadose zone in the partially saturated zone near the upper weir. Therefore the areal size of this partially saturated area and the thickness of the vadose zone prescribe the relative contributions of deep flow and near-channel vadose zone flow to runoff. A pair of recording piezometers located in this area, just 0.5 m upslope from the entrance to the upper weir and with their inlets at the colluvium/bedrock interface, provide the clearest picture of the health of the saturated zone. In the 40 hours after the sprin-

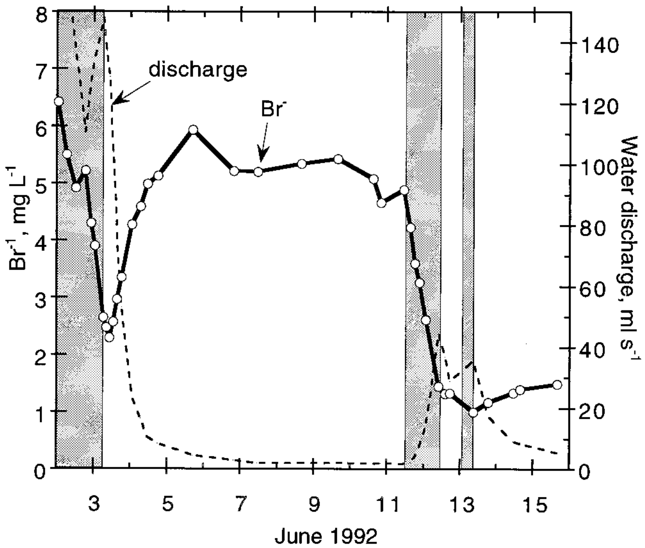

Figure 13. Bromide concentration, shown with thick line and data points, and discharge at the upper weir, shown with dotted line, from the falling limb of experiment 3 through experiment 4. Shading shows times of sprinkling or natural rainfall. Concentration of bromide rose with falling discharge at the end of experiment 3 and fell with increasing discharge during experiment 4. The same behavior, although at lower concentrations, was observed at the lower weir.

klers were shut off in experiment 3, both piezometers fell from their steady state pressure heads of ~0.2 m to 0 (dry) or 0.05 m; this is the precise time interval over which the bromide concentration rose and the δD fell in the runoff to the plateau values maintained during the low-flow period between sprinkling experiments. The saturated zone in the colluvium during long intervals between storms is so small and thin that it is difficult to measure with our instrumentation; however, the high bromide and low δD values between sprinkling experiments 3 and 4 show that exfiltration from bedrock dominates over percolation from the colluvium under these conditions.

Our view of the flow paths during the steady flow conditions achieved during sprinkling experiment 3 is shown in Figure 14. Water percolates through the unsaturated zone in the colluvium at a rate controlled by rainfall intensity. It then enters and flows through bedrock, guided across the interface and through the rock by exfoliation fractures. Where the colluvium or fractures are saturated, the average linear velocity increases by about 3 orders of magnitude. Thus, although there are rapid flow paths in the catchment, these contribute old water to runoff, because the water must first travel a slow, Darcian path through unsaturated colluvium. These experiments have demonstrated that during steady flow conditions, the velocity of flow across the bedrock/colluvium boundary in the saturated zone can be comparable to that of flow within the saturated colluvium. In order to maintain these high velocities under steady flow, the saturated zone must be limited in extent or of low porosity (in the bedrock). Thinning of the layers of fractured weathered rock downslope brings deep flow paths toward the surface near the channel head. There, old water mixes with new water; that is, water that has percolated down through the colluvium to intersect the saturated zone. Because this zone of mixing is small relative to the entire catchment area, runoff is predominantly old water at high flow. Flow velocities are low through the vadose zone, such that new water takes more than a day to appear under typical rainfall intensities, and average transit times are on the order of 6 months. The growth and decay of the saturated flow path dictates the mix of old and new water to runoff in nonsteady conditions. Flow paths through bedrock will predominate at all discharges; a contribution from a colluvium-only flow path is possible only during conditions wet enough to produce partial saturation of the colluvium near the channel head.

These flow paths can explain much of the concentration-discharge behavior at CB1 [*Anderson et al.*, 1997]. The magnitude of solute concentration changes are much smaller than discharge changes because old water dominates runoff at all flows. Variations in solute concentrations with discharge can be attributed to the changing transit times and flow paths under different hydrologic conditions. At high flow, a small but important portion of the discharge derives from water that has traveled only through the colluvium and, because of short flow path lengths, has relatively short transit times. The remainder of the flow travels through bedrock in addition to colluvium and therefore may be exposed to fresher, more reactive minerals. The average transit time of water in the saturated bedrock flow path is likely to decrease with increasing discharge, driven by greater head gradients. The old water collected passively in our lysimeters during the storm following experiment 4, shows that rainfall can mobilize stored water in soil; this may be a way to release solutes from water in small pores. The shortest transit time for water through the catchment, set by the depth to the water table near the channel head, is sufficiently long for exchange reactions to reach equilibrium; unreacted rainwater does not therefore play a role in the solute dynamics of the runoff.

A long-standing difficulty in hillslope hydrology is deciphering a mechanism or mechanisms that can produce sufficient volumes of subsurface flow to yield storm discharge [*Bonell*, 1993]. Here we see two distinct mechanisms: (1) rapid flow through shallow fractured bedrock and (2) formation of local subsurface saturated areas in the colluvium due to emergent bedrock flow, which divert vertically penetrating vadose zone water to runoff (as observed near the upper weir). The latter process is particularly important in controlling storm variations in the chemistry of runoff and the apportioning of old and new water. The volumes at peak flow seem consistent with these source generation mechanisms. The tracer data presented here unequivocally call for a driving mechanism that generates runoff by displacement of prestorm water. Data reported by R. Torres et al. (submitted manuscript) suggest that head gradients are rapidly transmitted through the vadose zone. Our tracer data show that the distance travelled by new water during a storm may be only a short distance through the colluvium mantle; we therefore infer that it is the head gradients associated with moisture increase that cause deeper water displacement from the vadose zone and expansion of the saturated zone. Our tracer data argue that once water is released from the vadose zone into saturated bedrock or colluvium, it quickly leaves the site. This quick delivery explains why spiked water reached the lower weir and why storm variations in water chemistry [*Anderson et al.*, 1997] occur there.

6. Conclusions

These tracer studies have demonstrated that in the CB1 catchment water moves through the vadose zone slowly at a

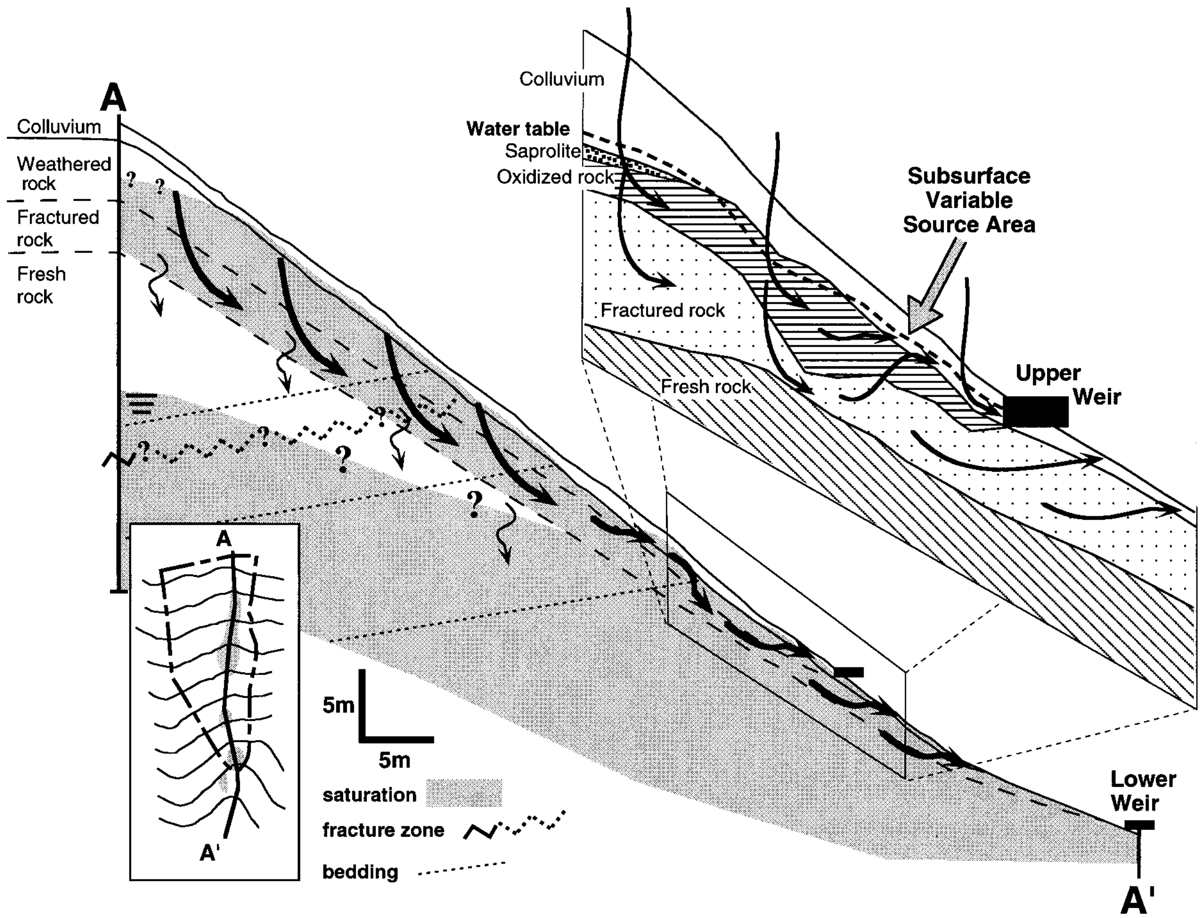

Figure 14. Longitudinal profile of the CB1 catchment, without vertical exaggeration, showing flow paths and geology, and the subsurface variable source area near the upper weir. In the exploded view at top, thickness of subsurface layers has been exaggerated two-fold. The subsurface saturated variable source area near the weir, shown here at its maximum extent, expands and contracts in response to rainfall, and controls the contribution of relatively short travelled, young water from the vadose zone to runoff. In the main figure, shading depicts areas believed to be saturated during the steady conditions of the sprinkling experiments. Saturated zone thickness in the colluvium has been exaggerated to make it visible. Extent of saturation in colluvium is constrained by over 200 piezometers; saturation in fresh rock is constrained by the 35 m well at the ridge crest. Solid arrows show inferred flow paths, predominantly infiltrating into the weathered rock layers high on the slope, but with considerable rapid flow back and forth between colluvium and weathered rock in the lower portion of the slope. The average transit time for water through this catchment is believed to be of order 6 months.

rate predicted for steady state conditions by the rainfall rate and the volumetric water content of the soil. Under nonsteady wetting conditions, water is mobilized in the soil, but this water is dominantly old water. We did not see evidence for preferential flow of new water under either the steady low-intensity conditions of our sprinkling experiments or in two natural storms. Flow in the saturated zone at the base of the colluvium is rapid and, taking advantage of exfoliation fractures, crosses back and forth between rock and colluvium in the zone of convergent topography (of both the surface and the bedrock/colluvium interface) down the axis of the catchment. The area of subsurface saturation in the colluvium near the channel head expands and contracts during storm events and, in so doing, exerts important control on the flow paths contributing runoff. Water percolating through the vadose zone in the colluvium mixes in this saturated area with water emerging from the bedrock. The bedrock flow path predominates under all flow conditions, but when the area of saturation near the weir grows both in areal extent and thickness during a storm, an increasing proportion of runoff is derived from vertical percolation through colluvium above this area. As *Hewlett and Hibbert* [1967] predicted in their conceptual model, we see a greater contribution to storm runoff from the colluvium near the channel, while other parts of the slope contribute a higher proportion of base flow runoff. Moreover, the timing of generation of runoff does not appear to be tied to the rate at which water moves through the catchment. We believe that this type of subsurface variable source area is likely to be important in runoff generation and stream chemistry of other steep lands with shallow soils overlying fractured bedrock.

Acknowledgments. This work was supported by the Weyerhaeuser Company, NSF (EAR 8417467) and USGS Water Resources (14080001G2111) grants to W.E.D., and a NASA Graduate Student Fellowship in Global Change Research (NGT 30083) to S.P.A. M.E.C. was supported by the Director, Office of Energy Research, Office of

Basic Sciences, Engineering and Geosciences Division of the U.S. Department of Energy under contract DE-AC03-76F00098. We thank T. Giambelluca of the University of Hawai'i for his meteorological instrumentation of the site. We appreciated thoughtful reviews by H. Elsenbeer, J. M. Buttle, and an anonymous reviewer. In addition, we thank F. Booker, H. Zender, C. Phillips, M. DeNiro, A. Brown, J. Repka, and R. S. Anderson for assistance in various aspects of this work.

References

Allan, C. J., and N. T. Roulet, Runoff generation in zero-order Precambrian Shield catchments: The stormflow response of a heterogeneous landscape, *Hydrol. Process.*, *8*, 369–388, 1994.

Anderson, S. P., Flow paths, solute sources, weathering, and denudation rates: The chemical geomorphology of a small catchment, Ph.D. thesis, 380 pp., Univ. of Calif., Berkeley, 1995.

Anderson, S. P., W. E. Dietrich, R. Torres, D. R. Montgomery, and K. Loague, Concentration-discharge relationships in a steep, unchanneled catchment, *Water Resour. Res.*, *33*, 211–225, 1997.

Baldwin, E. M., Eocene stratigraphy of southwestern Oregon, *Oreg. Dep. Geol. Miner. Ind. Bull.*, *83*, 1–40, 1974.

Bonell, M., Progress in the understanding of runoff generation dynamics in forests, *J. Hydrol.*, *150*, 217–275, 1993.

Brusseau, M. L., and P. S. C. Rao, Modeling solute transport in structured soils: A review, *Geoderma*, *46*, 169–192, 1990.

Buttle, J. M., Isotope hydrograph separations and rapid delivery of pre-event water from drainage basins, *Prog. Phys. Geogr.*, *18*, 16–41, 1994.

Buttle, J. M., and D. L. Peters, Inferring hydrological processes in a temperate basin using isotopic and geochemical hydrograph separation: A re-evaluation, *Hydrol. Process.*, *11*, 557–573, 1997.

Cheng, J. D., Subsurface stormflows in the highly permeable forested watersheds of southwestern British Columbia, *J. Contam. Hydrol.*, *3*, 171–191, 1988.

Dietrich, W. E., S. L. Reneau, and C. J. Wilson, Hollows, colluvium, and landslides in soil-mantled landscapes, in *Hillslope Processes*, edited by A. D. Abrahams, pp. 361–388, Allen and Unwin, Boston, Mass., 1986.

Driscoll, C. T., N. van Breemen, and J. Mulder, Aluminum chemistry in a forested Spodosol, *Soil Sci. Soc. Am. J.*, *49*, 437–444, 1985.

Dunne, T., Field studies of hillslope flow processes, in *Hillslope Hydrology*, edited by M. J. Kirkby, pp. 227–294, John Wiley, New York, 1978.

Dunne, T., T. R. Moore, and C. H. Taylor, Recognition and prediction of runoff-producing zones in humid regions, *Hydrol. Sci. Bull.*, *20*, 305–327, 1975.

Eshleman, K. N., J. S. Pollard, and A. K. O'Brien, Interactions between groundwater and surface water in a Virginia coastal plain watershed, 1, Hydrological flowpaths, *Hydrol. Process.*, *8*, 389–410, 1994.

Freeze, R. A., Streamflow generation, *Rev. Geophys.*, *12*, 627–647, 1974.

Freeze, R. A., and J. A. Cherry, *Groundwater*, 604 pp., Prentice-Hall, Englewood Cliffs, N. J., 1979.

Haagen, J. T., *Soil Survey of Coos County, Oregon*, 269 pp., Soil Conserv. Serv., U.S. Dep. of Agric., 1989.

Hammermeister, D. P., G. F. Kling, and J. A. Vomocil, Perched water tables on hillsides in western Oregon, I, Some factors affecting their development and longevity, *Soil Sci. Soc. Am. J.*, *46*, 811–818, 1982.

Harr, R. D., Water flux in soil and subsoil on a steep forested slope, *J. Hydrol.*, *33*, 37–58, 1977.

Herwitz, S. R., Evaluation of the assumption of watertight bedrock underlying an experimental watershed, *Phys. Geogr.*, *14*, 465–477, 1993.

Hewlett, J. D., and A. R. Hibbert, Factors affecting the response of small watersheds to precipitation in humid areas, in *International Symposium on Forest Hydrology*, edited by W. E. Sopper and H. W. Lull, pp. 275–290, Pergamon, Tarrytown, N. Y., 1967.

Hornberger, G. M., and E. W. Boyer, Recent advances in watershed modelling, *U.S. Natl. Rep. Int. Union Geol. Geophys. 1991–1994*, *Rev. Geophys.*, *33*, 949–957, 1995.

Humphrey, N. F., Pore pressures in debris failure initiation, M.S. thesis, 169 pp., Univ. of Wash., Seattle, 1982.

Hursh, C. R., Storm-water and adsorption, *Eos Trans. AGU*, *17*, 301–302, 1936.

Iida, T., A hydrological method of estimation of the topographic effect on the saturated throughflow, *Trans. Jpn. Geomorphol. Union*, *5*(1), 1–12, 1984.

Kennedy, V. C., C. Kendall, G. W. Zellweger, T. A. Wyerman, and R. J. Avanzino, Determination of the components of stormflow using water chemistry and environmental isotopes, Mattole River basin, California, *J. Hydrol.*, *84*, 107–140, 1986.

Litaor, M. I., Aluminum mobility along a geochemical catena in an alpine watershed, Front Range, Colorado, *Catena*, *19*, 1–16, 1992.

Luxmoore, R. J., Micro-, meso-, and macroporosity of soil, *Soil Sci. Soc. Am. J.*, *45*, 671–672, 1981.

Luxmoore, R. J., and L. A. Ferrand, Towards pore-scale analysis of preferential flow and chemical transport, in *Water Flow and Solute Transport in Soils*, edited by D. Russo and G. Dagan, pp. 45–60, Springer-Verlag, New York, 1993.

McDonnell, J. J., A rationale for old water discharge through macropores in a steep, humid catchment, *Water Resour. Res.*, *26*, 2821–2832, 1990.

McDonnell, J. J., M. Bonell, M. K. Stewart, and A. J. Pearce, Deuterium variations in storm rainfall: Implications for stream hydrograph separation, *Water Resour. Res.*, *26*, 455–458, 1990.

Montgomery, D. R., W. E. Dietrich, R. Torres, S. P. Anderson, J. T. Heffner, and K. Loague, Hydrologic response of a steep, unchanneled valley to natural and applied rainfall, *Water Resour. Res.*, *33*, 91–109, 1997.

Mulholland, P. J., Hydrometric and stream chemistry evidence of three storm flowpaths in Walker Branch Watershed, *J. Hydrol.*, *151*, 291–316, 1993.

Nielsen, D. R., M. T. van Genuchten, and J. W. Biggar, Water flow and solute transport processes in the unsaturated zone, *Water Resour. Res.*, *22*, 89S–108S, 1986.

Nyström, U., Transit time distributions of water in two small forested catchments, in *Lake Gårdsjön—An Acid Forest Lake and Its Catchment*, edited by F. Andersson and B. Olsson, pp. 97–100, Ecol. Bull., Stockholm, 1985.

Ogunkoya, O. O., and A. Jenkins, Analysis of runoff pathways and flow contributions using deuterium and stream chemistry, *Hydrol. Process.*, *5*, 271–282, 1991.

Pearce, A. J., M. K. Stewart, and M. G. Sklash, Storm runoff generation in humid headwater catchments, 1, Where does the water come from?, *Water Resour. Res.*, *22*, 1263–1271, 1986.

Peters, D. L., J. M. Buttle, C. H. Taylor, and B. D. LaZerte, Runoff production in a forested, shallow soil, Canadian Shield basin, *Water Resour. Res.*, *31*, 1291–1304, 1995.

Rawlins, B. G., A. J. Baird, S. T. Trudgill, and M. Hornung, Absence of preferential flow in the percolating waters of a coniferous forest soil, *Hydrol. Process.*, *11*, 575–585, 1997.

Robinson, J. S., and M. Sivapalan, Instantaneous response functions of overland flow and subsurface stormflow for catchment models, *Hydrol. Process.*, *10*, 845–862, 1996.

Rodhe, A., L. Nyberg, and K. Bishop, Transit times for water in a small till catchment from a step shift in the oxygen 18 content of the water input, *Water Resour. Res.*, *32*, 3497–3511, 1996.

Sidle, R. C., Y. Tsuboyama, S. Noguchi, I. Hosoda, M. Fujieda, and T. Shimizu, Seasonal hydrologic response at various spatial scales in a small forested catchment, Hitachi Ohta, Japan, *J. Hydrol.*, *168*, 227–250, 1995.

Sklash, M. G., Environmental isotope studies of storm and snowmelt runoff generation, in *Process Studies in Hillslope Hydrology*, edited by M. G. Anderson and T. P. Burt, pp. 401–436, John Wiley, New York, 1990.

Sklash, M. G., M. K. Stewart, and A. J. Pearce, Storm runoff generation in humid headwater catchments, 2, A case study of hillslope and low-order stream response, *Water Resour. Res.*, *22*, 1273–1282, 1986.

Sposito, G., and W. A. Jury, The lifetime probability density function for solute movement in the subsurface zone, *J. Hydrol.*, *102*, 503–518, 1988.

Trudgill, S., Hillslope solute modelling, in *Modelling Geomorphological Systems*, edited by M. G. Anderson, pp. 309–339, John Wiley, New York, 1988.

Turton, D. J., C. T. Haan, and E. L. Miller, Subsurface flow responses of a small forested catchment in the Ouachita Mountains, *Hydrol. Process.*, *6*, 111–125, 1992.

Vennemann, T. W., and J. R. O'Neil, A simple and inexpensive method of hydrogen isotope and water analyses of minerals and rocks based on zinc reagent, *Chem. Geol.*, *103*, 227–234, 1993.

Williams, M., R. Kattelmann, and J. Melack, Groundwater contributions to the hydrochemistry of an alpine basin, in *Hydrology in Mountainous Regions*, vol. 1, *Hydrological Measurements; the Water Cycle*, edited by H. Lang and A. Musy, pp. 741–748. Int. Assoc. of Hydrol. Sci., Gentbrugge, Belgium, 1990.

Wilson, C. J., and W. E. Dietrich, The contribution of bedrock groundwater flow to storm runoff and high pore pressure development in hollows, in *Erosion and Sedimentation in the Pacific Rim*, edited by R. L. Beschta et al., pp. 49–59. Int. Assoc. of Hydrol. Sci., Gentbrugge, Belgium, 1987.

Wilson, G. V., P. M. Jardine, R. J. Luxmoore, L. W. Zelazny, D. A. Lietzke, and D. E. Todd, Hydrogeochemical processes controlling subsurface transport from an upper subcatchment of Walker Branch watershed during storm events, 1, Hydrologic transport processes, *J. Hydrol.*, *123*, 297–316, 1991.

Yee, C. S., and R. D. Harr, Influence of soil aggregation on slope stability in the Oregon Coast Ranges, *Environ. Geol.*, *1*, 367–377, 1977.

S. P. Anderson, Department of Earth Sciences and Institute of Tectonics, University of California, Santa Cruz, CA 95064-1077. (e-mail: spa@bagnold.ucsc.edu)

M. E. Conrad, Berkeley Center for Isotope Geochemistry, Lawrence Berkeley Laboratory, Berkeley, CA 04720.

W. E. Dietrich, Department of Geology and Geophysics, University of California, Berkeley, CA 94720-4767.

K. Loague, Department of Geological and Environmental Sciences, Stanford University, Stanford, CA 94305-2115.

D. R. Montgomery, Department of Geological Sciences, University of Washington, Seattle, WA 98195.

R. Torres, Department of Geological Sciences, University of South Carolina, Columbia, SC 29208.

(Received October 15, 1996; revised September 8, 1997;
received September 11, 1997.)