HYDROLOGIC FLOWPATHS OF SNOWPACK RUNOFF IN A HIGH-ELEVATION CATCHMENT

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INTRODUCTION

High-elevation basins will show hydrologic and hydrochemical effects from global or regional climate change much sooner than will basins located at lower elevations. Research at several high-elevation sites in the western U.S. has shown that the hydrology and hydrochemistry of these alpine areas are sensitive indicators of climatic change [Caine, 1989a; Baron, 1990; Williams and Melack, 1991a]. The combination of small hydrologic storage in groundwater reservoirs, the predominance of intrusive igneous rocks that weather slowly, the thin acidic soils, large amount of precipitation, and low buffering ability of alpine basins result in high-elevation areas of the western U.S. responding quickly to changes in the quantity and quality of precipitation. In particular, global or regional climate change may increase both the magnitude and frequency of floods from snowpack runoff [*e.g.* Lettenmaier and Gan, 1989].

Hydrologic and hydrochemical effects of snowpack runoff are a function of the path water takes as it leaves the snowpack and moves towards surface waters. Previous work on hydrographic separation has shown that much of stream flow is composed of pre-event or baseflow, rather than snowpack runoff (e.g. Bottomley [1984]; Bottomley, [1986]; Hooper and Shoemaker [1986]). However, much of this research on hydrographic separation of stream waters has been conducted in forested sites in eastern North America, which generally have large deposits of glacial till and relatively large water storage, compared to high-elevation basins in the western United States.

Little is known on hydrologic pathways in alpine areas of western North America. In general, stream and lake waters in alpine basins in western North America have been considered to be a mixture of groundwater and relatively dilute snowpack runoff [Miller and Drever, 1977; Loranger and Brakke, 1988]. Infiltration of meltwater into soil and groundwater reservoirs of alpine basins has generally been ignored [e.g. Drever and Hurcomb, 1986].

It is essential to understand hydrological flowpaths in high-elevation basins if we are to forecast the hydrologic and hydrochemical consequences of a changing climate. Here I evaluate the hydrologic flowpaths of snowmelt runoff in the Emerald Lake watershed (ELW), a high-elevation catchment in the southern Sierra Nevada, through

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the use of introduced and natural geochemical tracers in the snowpack, snowpack meltwater, soil waters, and stream waters.

SITE DESCRIPTION

The Emerald Lake basin is a north-facing granitic cirque located on the upper Marble Fork of the Kaweah River drainage, in the southern Sierra Nevada of California, USA (36°35'49"N, 118°40'30"W). Basin area is 120 ha; elevation ranges from 2800 m at the lake outlet to 3416 m at the summit of Alta Peak. Emerald Lake is a 2.72 ha cirque lake at the bottom of the basin, fed by two main inflows and six intermittent streams, and drained by a single outflow (Figure 1). Massive rock outcrops cover 33 percent of the basin area; unconsolidated sand, gravels, and talus cover about 23 percent. The remaining area is mapped as a rock-soil complex, which is about half soil and half rock outcrop. Vegetation covers about 20 percent of the basin area, of which 3 percent has scattered trees. Snow provides about 95% of the precipitation input to the basin [Kattelmann and Elder, 1991].

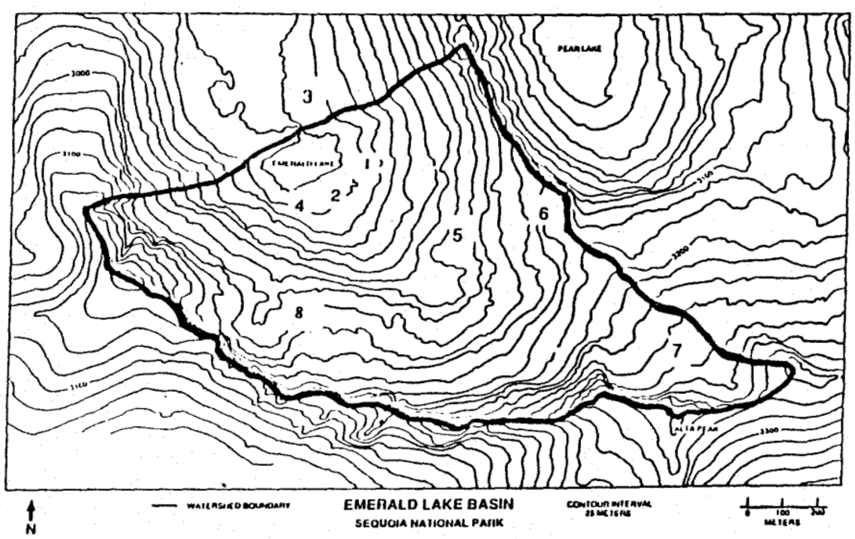

Figure 1. Topographic map of the Emerald Lake watershed and location of sites: 1 inflow 1, 2 inflow 2, 3 outflow, 4 inflow 4, 5 bench, 6 ridge, 7 cirque, 8 hole.

The free water capacity of soils in the basin is approximately 29,000±7800 m^3, defined as saturated capacity less field capacity and estimated using measured depths and physical properties of the soils. Combining the free water capacity of soils with estimates for unconsolidated materials, the total free water capacity of the basin is about 48,000±13,000 m^3 [Brown et al., 1990]. The heterogeneous distribution of soil water storage (as free water) is governed by the depth and texture of soils and unconsolidate materials, with zones of higher water capacity tending to follow perennial and ephemeral stream channels. Laboratory measurements of the saturated hydraulic conductivity ranged from 0.1 to 0.01 mm s^{-1} [Brown et al., 1990].

METHODS

Stage height in 1986 and 1987 was measured in the outflow and two main inflows with a Montedero-Whitney pressure transducer and recorded on an Omnidata data logger. Stage-discharge relationships were developed using a salt dilution technique [Kattelmann and Elder, 1991]. Sampling sites for stream water were located immediately above Emerald Lake for inflows 1, 2 and 4 and immediately below the lake for the basin outflow (Figure 1). Water samples for chemical analysis were collected weekly from the onset of snowpack runoff to September. Additionally, the following measurements were made in soils at the ridge site (elevation 3100 m) and at the bench site (elevation 2900 m) (Figure 1) at depths of about 100 mm and 300 mm: duplicate measurements of the solute composition of soil water with membrane-covered porous polyethylene plates under tension, temperature of the soil solution, and moisture content [Brown et al., 1990]. Chemical analysis consisted of measuring cation concentrations via atomic absorption and anion concentrations via ion chromatography. Reactive silicate (Si) was measured by the silic-molybdate method.

I separated the stream hydrograph of the ELW and Martinelli catchment into groundwater and surface runoff using a simple mixing model with two components:

$$Q_p = Q_t \left[\frac{C_t - C_e}{C_p - C_e} \right]$$

where Q is discharge (m^{3}day^{-1}); C is the tracer concentration; t is total stream discharge, e is the surface runoff component, and p is the groundwater component. Basic assumptions of this mixing model are: i) the chemical content of surface water (C_e) is significantly different from that of groundwater (C_p); ii) C_e is characterized by a single value or variations in the value are accounted for; iii) contributions of soil water are insignificant; iv) surface storage contributions are negligible; and (v) the mixing of water from different sources in the channel is complete [e.g. Sklash and Farvolden, 1979; Wels et al., 1990]. Groundwater was therefore the component of stream water stored in soil (vadose) and groundwater reservoirs prior to the initiation of snowpack runoff, or "pre-event" water. Surface flow was from snowpack meltwater and rain and is "event" water as defined by Hooper and Shoemaker [1986]. Therefore, this mixing model does not consider as baseflow water that infiltrated into groundwater and vadose reservoirs during the period of snowpack runoff, with subsequent discharge to surface flow.

Hydrologic residence time at the ELW during maximum discharge was determined experimentally using an applied tracer composed of $^{6}LiBr$. The tracer was introduced as a 1-L solution to a small ephemeral stream in an area of unconsolidated materials located at the source area of inflow 4, at an elevation of 2990 m (site 8, Figure 1). There is no surface discharge from this area; water is discharged through springs and seepage at an elevation of about 2900 m. Samples were collected at the gauging site in inflow 4, which is located close to the lake at an elevation of about 2815 m. The vertical relief between the application site and the collection site was about 175 m and the linear distance was approximately 350 m, with an average slope of about 30°. Water samples were collected before the tracer was introduced to determine background concentrations of Br^- and $^{6,7}Li^+$ at the application site and the collection site.

RESULTS

The mixing model indicates there was little difference in the pre-event and snowpack runoff contributions to stream flow in 1986 and 1987. The pre-event component of runoff for inflow 2 was calculated using values of C_p = 58 μmol L^{-1} and C_e = 0 μmol L^{-1}. The C_p value is representative of Si concentrations in inflow 2 during the period of low-flow [Williams et al., 1992]. At peak flow, in both 1986 and 1987, the mixing model calculated pre-event water to be about 30 percent of total flow (Figure 2). Over the entire period of snowpack runoff for inflow 2, the mixing model calculates that pre-event water contributed about 41% of total discharge in 1986 and 44% in 1987. Other inflows had similar values.

However, the assumption that a single value can represent pre-event water does not work at the ELW. I calculated the mean residence time of groundwater in the ELW for 1986 by dividing the volume of groundwater discharge (Q_{GW}) by the size of the groundwater reservoir (V_{GW}). Kattelmann [1989] has calculated V_{GW} for the ELW as $120{,}000 \pm 60{,}000\,m^3$. For the time period of April 10 to August 30, 1986, the mixing model calculated that Q_{GW} supplied about $1.124 \times 10^6\,m^3$ of the $2.284 \times 10^6\,m^3$ of the water discharged in the outflow. Using the minimum ($60{,}000\,m^3$) and maximum ($180{,}000\,m^3$) values for V_{GW}, the groundwater reservoir was flushed from six to nineteen times during this period. The mean hydrologic residence time for this period of 143 days was in the range of seven to twenty-three days. Clearly, this flushing rate is incompatible with the assumption of the mixing model that measured Si in surface waters during snowpack runoff represents pre-event waters.

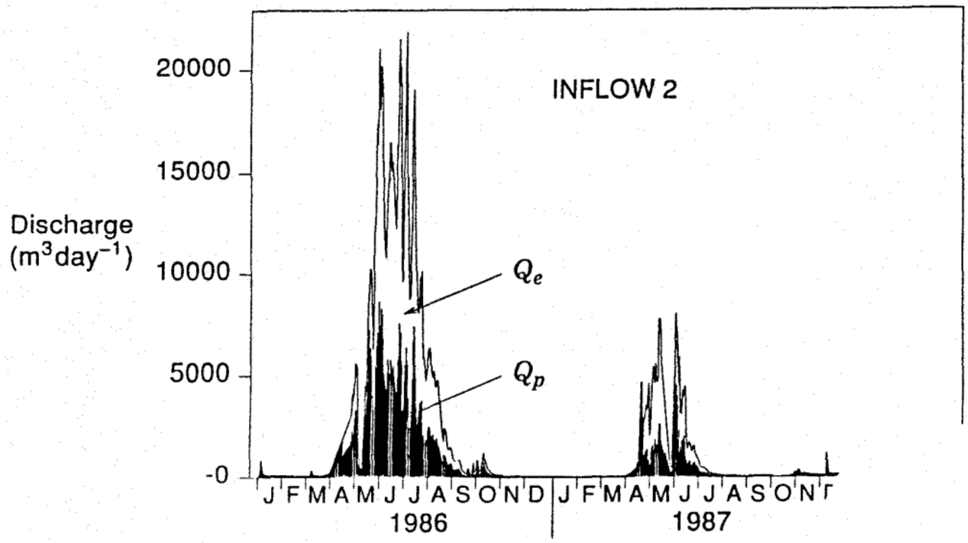

Figure 2. Hydrograph separation of inflow 2 into contributions from pre-event water (Q_p) and snowpack runoff, or event water (Q_e).

Hydrologic Residence Time. The tracer experiment with LiBr shows that the residence time of groundwater at maximum discharge in 1987 was about 12 hours at this site. The first detection of the Br^- tracer was observed 9 hours after injection and the last measurable amount was observed 10 hours later (Figure 3). The peak occurred 12 hours after injection and about 1 hour after discharge measured in inflow 4 reached its annual maximum. The well-defined temporal response curve for Br^- suggests that Br^- was conserved. In contrast, the results for the $^6Li^+/^7Li^+$ ratio were not well-defined (Figure 3). The ratio of $^6Li^+/^7Li^+$ increased to a maximum which was temporally coincident with that of Br^-. However the ratio of $^6Li^+/^7Li^+$ did not return to within one standard deviation of background levels at the end of the experiment. The temporal response of ^6Li suggests that this ion was not conserved.

The hydrologic residence time of water in soils and talus was estimated at hours to days during snowpack runoff, based on two additional methods. First, hydrologic residence time was estimated as the amount of free water in the basin divided by the amount of discharge from the basin. The calculated free water capacity of soil and talus at the ELW was estimated at approximately 48,000±13,000 m^3. Peak daily

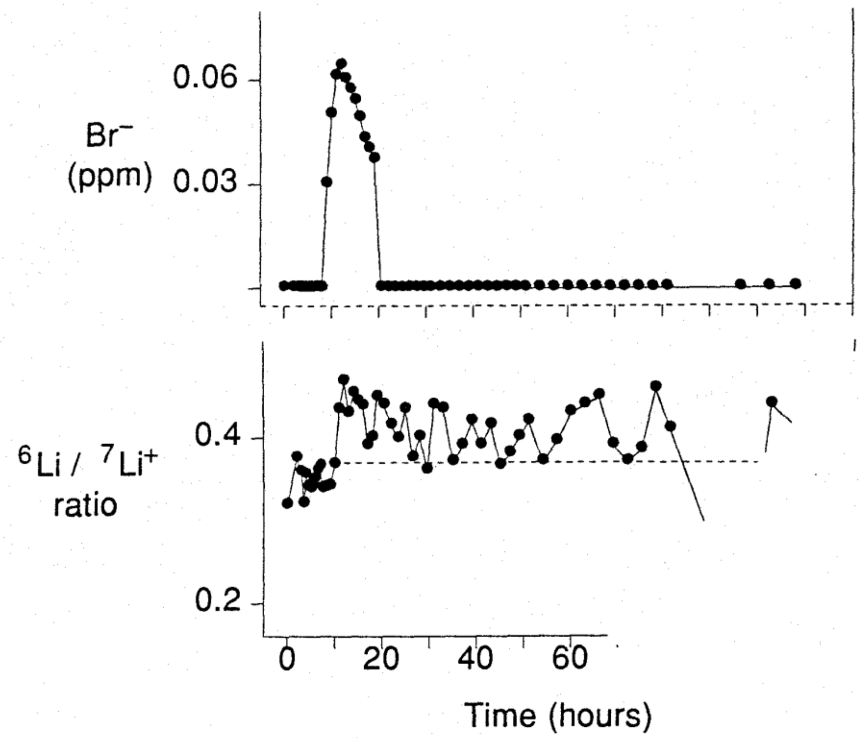

Figure 3. A time series of Br^- concentrations and the ratio of $^6Li^+/^7Li^+$ in inflow 4, after a salt of ^6LiBr was applied to an ephemeral stream during the period of snowpack runoff in the recharge area of inflow 4. The dashed line is the average background ion intensity ($^6Li^+$) or ratio ($^6Li^+/^7Li^+$), plus one standard deviation.

discharge from the basin was 36,000 m^3 in May of 1986 and in May of 1987 was about 17,000 m^3. Therefore, water storage in soils and talus could turn over almost daily in 1986 and every other day in 1987, assuming all snowpack runoff flowed through soils and talus. The second method of determining hydrologic residence time was to calculate the saturated hydraulic conductivity (K_w) of each soil type in the ELW, then multiply K_w by soil depth. During spring runoff soils are saturated and water movement through the soil may be described by K_w. Given average soil depths of 0.5 to 1 m in the basin and saturated conditions, water would pass through the soil profile in a matter of minutes to hours, and then be available for surface runoff.

Snowpack runoff-soil interactions. Some amount of snowpack runoff appeared to infiltrate soils and unconsolidated materials. I evaluated the importance of interactions between snowpack runoff and soils by analyzing the molar ratio of Na:Ca in the soil solution. Selectivity coefficients of soils in the ELW result in soil exchangers preferentially retaining Ca^{2+} over Na^{+}. Consequently, Na^{+} is leached from soils more readily than Ca^{2+}. This leads to the prediction that if cation exchange was responsible for the neutralization of the H^{+} in wet deposition, the molar ratio of Na:Ca in soil solution should increase during snowpack runoff, and then decrease towards the end of snowpack runoff.

The Na:Ca molar ratio increased by 230%, at depths of both 100 mm and 300 mm, at the bench site during snowpack runoff (Figure 4). The Na:Ca ratio at the bench site then decreased back to pre-snowmelt levels towards the end of snowpack runoff. If cation exchange in soils had an effect on the solute composition of surface waters, the molar ratio of Na:Ca in stream waters should have increased during the period of snowpack runoff, provided that: (1) exchange of the H^{+} in snowpack runoff for base cations on soil exchange sites was important at the scale of the basin, and (2) discharge from soil reservoirs was an important component of stream flow. The molar ratio of Na:Ca in all three streams increased by a factor of 230 percent during the period of snowpack runoff (Figure 4). The large increase in the Na:Ca ratio of stream waters during the period of snowpack runoff was comparable to that in soil reservoirs, and indicates that cation exchange was important at the scale of the basin. The ratio of Na:Ca then decreased back to pre-snowmelt levels, as did the Na:Ca ratio in the soil solution. During the autumn months the Na:Ca ratio in soils and stream became progressively out-of-phase. The Na:Ca ratio at this time reached an annual maximum in the soil solution and an annual minimum in stream waters (Figure 4). Discharge from soil reservoirs does not appear to be an important source of stream flow during the autumn and winter months.

DISCUSSION

Pre-event water at the ELW contributes only a small fraction of stream flow during the period of snowpack runoff. The use of a single C_p for hydrograph separation of stream flow during runoff generally underestimates the contribution of groundwater to stream flow in basins with small groundwater reservoirs [Wels et al., 1990]. My results suggest that much of snowpack runoff in the ELW infiltrates soils and unconsolidated materials, undergoes reactions with soil water and soil exchangers, and then is discharged to stream flow.

I re-evaluated the contribution of subsurface water to stream flow by allowing C_p to vary over time, the "piecewise linear" method of hydrographic flow separation developed by Hooper and Shoemaker [1986]. We made the assumption that the Si content of the soil solution and groundwater were the same [e.g. Wels et al., 1990].

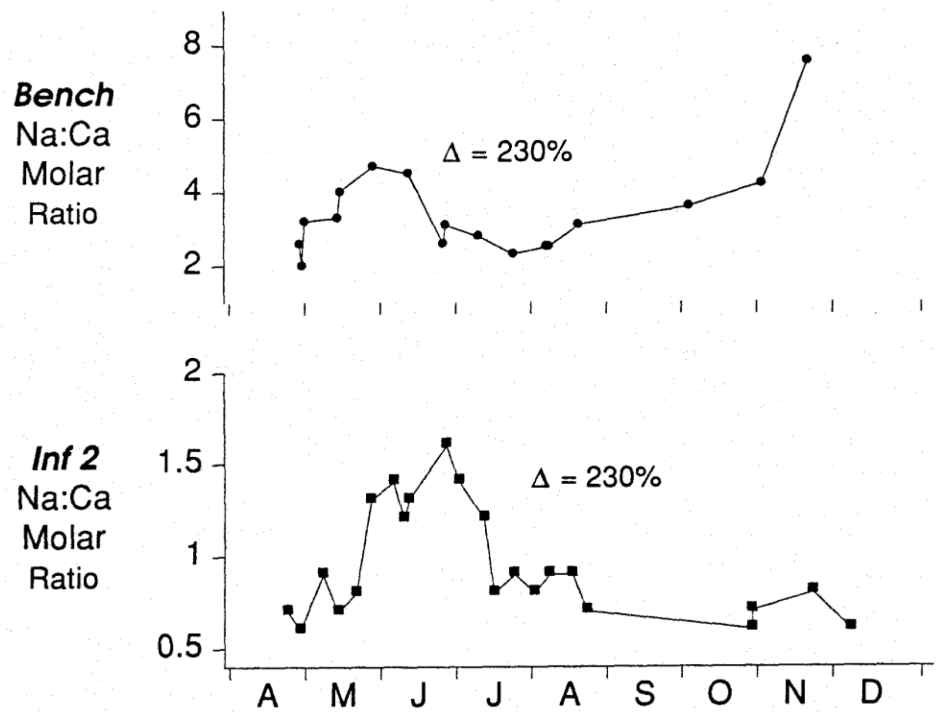

Figure 4. Time series from 1987 of the molar ratios of Na:Ca from soil lysimeters at the bench site and for inflows 1 and 2. Both soil lysimeters and stream waters showed an increase of 230% in the molar ratio of Na:Ca during the period of snowpack runoff.

The concentrations of Si used for C_p were calculated by averaging all Si measurements in the soil solution for a particular day; this value of C_p was used in the mixing model until the next date of Si measurements in the soil solution. The time-frame represented by Si measurements of the soil water component of the hydrograph was therefore shortened from seasonal to daily.

The hydrograph for snowmelt runoff in 1987 shows two melt episodes, one starting in mid-April and the second starting in early June (Figure 5). The new hydrograph shows that subsurface discharge in 1987 accounted for about 62 percent of total flow in inflow 2; surface runoff accounted for about 38 percent of flow. Subsurface contributions to stream flow were near 100 percent at the initiation of each melt episode. The contributions of subsurface water to stream waters then decreased and the relative contributions of surface runoff to stream flow increased. Much of the subsurface contribution to stream flow was new water that had infiltrated into subsurface reservoirs. Caine [1989b] reports similar results for an alpine basin in Colorado; up to 50 percent of snowpack runoff was routed through surficial deposits in the Martinelli catchment of the Green Lakes Valley.

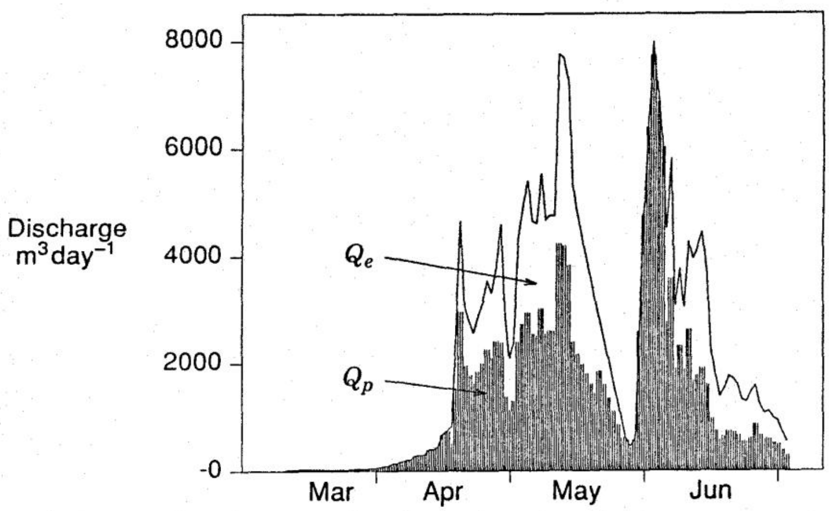

Figure 5. Hydrograph separation of inflow 2 into contributions from soil discharge (Q_p) and surface runoff (Q_e), where the value of C_e varies on a daily basis.

These results suggest that the first fraction of snowpack meltwater during a melt episode infiltrates soils and talus. Soils and talus then become saturated and discharge water to streams as saturated overland flow. The volume of snowpack runoff then exceeds the infiltration capacity of soils and an increasingly larger percentage of snowpack runoff flows towards streams as Hortonian overland flow. The congruency between changes in the Na:Ca molar ratio of soils and stream water is consistent with return flow from soils supplying much of stream discharge. Furthermore, the non-conservative behavior of Li suggests that cation exchange reactions modified the tracer introduced to the recharge area of inflow 4. The location of zones of higher water capacity along stream channels also suggests that much of snowpack runoff finds its way into soil and talus before becoming stream flow. Processes with rapid kinetics that occur within subsurface reservoirs may therefore be the primary controls on the composition of surface waters in alpine basins during the period of snowpack runoff.

REFERENCES

Baron, J. (1990), *Biogeochemistry of a subalpine ecosystem: Loch Vale Watershed,* Ecological Studies 90, Springer-Verlag, New York, NY.

Bottomley, D. J., D. Craig, and L. M. Johnston (1984/1985), Neutralization of acid runoff by groundwater discharge to streams in Canadian Precambrian Shield watersheds, *Journal of Hydrology.* 75:1-26.

Bottomley, D. J., D. Craig, and L. M. Johnston (1986), Oxygen-18 studies of snowmelt runoff in a small Precambrian Shield watershed: implications for streamwater acidification in acid-sensitive terrain, *Journal of Hydrology.* 88:213-234.

Brown, A. D., L. J. Lund, and M. A. Lueking (1990), Integrated Soil Processes Studies at Emerald Lake Watershed, Final Report, Contract A5-204-32, California Air Resources Board, Sacramento, CA.

Caine, N. (1989b), Hydrograph separation in a small alpine basin based on inorganic solute concentrations, *Journal of Hydrology.* 112:89-101.

Caine, N. (1989a), Diurnal variations in the quality of water draining from an alpine snowpack, *Catena.* 16:153-162.

Drever, J. I. and D. R. Hurcomb (March 1986), Neutralization of atmospheric acidity by chemical weathering in an alpine basin in the North Cascade Mountains, *Geology.* 14:221-224.

Hooper, R. P. and C. A. Shoemaker (1986), A comparison of chemical and isotopic hydrograph separation, *Water Resources Research.* 22:1444-1454.

Kattelmann, R. (1989), Groundwater contributions in an alpine basin in the Sierra Nevada, *Proceedings of the Headwaters Hydrology Symposium*:361-369, American Water Resources Association, Bethesda, MD.

Kattelmann, R. and K. Elder (1991), Hydrologic characteristics and water balance of an alpine basine in the Sierra Nevada, *Water Resources Research.* 27:1553-1562.

Lettenmaier, D. P. and T. H. Gan (1989), An exploratory analysis of the hydrologic effects of global warming on the Sacramento-San Joaquin River Basin, California, *Water Resources Research.*

Loranger, T. J. and D. F. Brakke (1988), The extent of snowpack influence on water chemistry in a north Cascades Lake, *Water Resources Research.* 24:723-726.

Miller, W. R. and J. I. Drever (1977), Chemical weathering and related controls on surface water chemistry in the Absaroka Mountains, Wyoming, *Geochimica et Cosmochimica Acta.* 41:1693-1702.

Sklash, M. G. and R. N. Farvolden (1979), The role of groundwater in storm runoff, *Journal of Hydrology.* 43:45-65.

Wels, C., R. J. Cornett, and B. D. LaZerte (1990), Groundwater and wetland contributions to stream acidification: an isotopic analysis, *Water Resources Research.* 26:2993-3003.

Williams, M. W., A. Brown, and J. M. Melack (1992 (in press)), Geochemical controls on the composition of surface waters in an alpine basin, Sierra Nevada, *Limnology and Oceanography.*

Williams, M. W. and J. M. Melack (1991), Solute chemistry of snowmelt and runoff in an alpine basin, Sierra Nevada, *Water Resources Research.* 27:1575-1588.