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Geochemical Investigations of Calcite Fracture Fills and Mesa-Top Water Dynamics on the Pajarito Plateau, New Mexico

By

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CHAPTER 2: INTERFLOW PATHWAYS IN A SUBHUMID PONDEROSA PINE HILLSLOPE

ABSTRACT

The mechanisms controlling interflow in semiarid and subhumid environments have received relatively little attention despite that fact that interflow can be an important runoff process in these environments. A recent study in a subhumid ponderosa pine hillslope demonstrated that interflow can be volumetrically as important as overland flow. However, the factors controlling interflow in this system were still poorly understood. The objective of the current study, carried out on the same ponderosa pine hillslope at Los Alamos, New Mexico, is to better understand the interflow process in subhumid environments and to develop a conceptual model of interflow that will aid in evaluating the potential for contaminant transport in such environments. Natural chloride and stable isotope (δD and $\delta^{18}O$) tracers were used to investigate the interflow process and the chemical changes that occur as a result of interflow. Observed differences in chemistry between soil matrix water and interflow were large (for example, chloride concentrations in matrix soil water samples were over 200 mg/L, compared with only 2 mg/L in interflow samples obtained at the same time), indicative of a two-domain

flow system in which preferential flow paths generate interflow that is not in chemical or hydrological equilibrium with the soil matrix. In addition, there were large changes in the chemistry of interflow water with time; for example, chloride concentrations were 10 times greater under saturated conditions than under unsaturated conditions. Additional evidence for the movement of interflow through preferential flow paths is provided by observed rapid shifts in interflow isotopic composition (e.g., 6 ‰ $\delta^{18}O$ in 24 hours) which are not consistent with the very low bulk hydraulic conductivity of the soils. The chloride and stable isotope tracer results show that interflow travels mainly through preferential flow paths, which affects the chemistry, residence time, and distribution of water in the soils–and thereby probably affects contaminant distribution and mobility as well.

INTRODUCTION

Interflow or subsurface stormflow is the lateral movement of water through near-surface soils, regolith, and bedrock (Anderson and Burt, 1990; Satterlund and Adams, 1992). Most of our knowledge about interflow processes comes from studies conducted at humid sites. In contrast, interflow in subhumid/semiarid regions of the United States has been little studied, probably because it was not considered an

important hydrologic process: these regions have few perennial streams, receive relatively small amounts of precipitation, and have low soil-moisture content throughout much of the year. Humid systems receive much larger amounts of water and are characterized by less variable moisture conditions.

Interflow in humid areas can be generated under both unsaturated and saturated conditions (McDonnell, 1991). Under unsaturated conditions, the flow is usually thought to occur via macropores (Germann, 1990; McDonnell, 1991). The macropores either channel water from the surface, generating interflow directly, or they feed shallow, perched saturated zones overlying bedrock that is impermeable or of low permeability, indirectly producing interflow (Whipkey, 1965; McDonnell, 1991; Peters et al., 1995; Turton et al., 1995). Macropore flow under unsaturated conditions occurs when the flux of water (precipitation or snowmelt) is greater than the hydraulic conductivity of the matrix (Germann, 1990; Sklash, 1990; McDonnell, 1991). This process will be enhanced in areas that are characterized by large or intense rains or snowmelt events and/or where the soils have low matrix conductivities.

Stable isotope tracers have been used to ascertain how much of the interflow in humid environments is "old" water and how much is

"new" water. Old, or pre-event, water has been in storage and is forced out of the soil by a current storm or snowmelt event. New, or event, water comes directly from the current event. Many of the stable isotope studies have shown that most of the hydrograph rise in streams is caused by old water (e.g., Bottomley et al., 1984; Pearce et al., 1986; McDonnell, 1991; Anderson et al., 1994; DeWalle and Pionke, 1994). To explain this dominance of old water, Sklash and Farvolden (1979) proposed the groundwater ridging concept, in which water in the tension saturated zone near the stream channel is quickly converted to a phreatic state by a relatively small additional input of water, causing a fast release of old water to the channel. Although there is always a new-water component from in-channel precipitation, overland flow, or subsurface macropore flow, it is generally small. However, new water can be important in upslope areas away from stream channels. For example, based on rainfall experiments on a humid zone soil block, Turton et al. (1995) suggested that new water was an important component of interflow during infrequently occurring high intensity storms, while old water dominated during the frequent small storms that occur during a given year.

The need for an improved understanding of how rain and snowmelt move through the soil as interflow, and the importance of

hillslope hydrology tracer studies to gain that understanding, was pointed out by Sklash (1990). Such an understanding is vital because the processes that control the movement of water through soils affect the mobility of contaminants, the distribution of nutrients, and the acid-base chemistry of surface waters (Mulholland et al., 1990). As outlined above, great strides have been made in understanding the process of interflow generation in humid environments. This study attempts to develop a comparable understanding for dry subhumid-semiarid systems.

The current study uses natural chloride and stable isotope tracers ($\delta^{18}O$ and δD) which originate from precipitation, to investigate the processes that control interflow in a subhumid hillslope. Conducted at a ponderosa-pine hillslope site at Los Alamos, New Mexico (Figure 2.1), this study builds on the work of Wilcox et al. (1996). Overland (surface) flow had been considered to be the major mechanism of runoff generation in these environments, but Wilcox et al. found that interflow can be the major runoff mechanism under particular circumstances. They found that interflow was most active during spring snowmelt and, surprisingly, was moving mostly through a dense, clay Bt soil horizon having a very low saturated hydraulic conductivity (2.5×10^{-10} m/sec). Their observation of the dynamic nature of interflow led Wilcox et al. to

hypothesize that interflow was moving, at least in part, through macropores. The hydrometric measurements used in their study suggested that interflow rates were rapid enough to require such an explanation, but these measurements alone are not sufficient for a detailed model of interflow generation. Other methods are needed to determine the pathway of interflow (macropores vs. soil matrix), the rate of water movement, and the effects of interflow on soil water chemistry. In our study, therefore, a multiple-tracer approach was used to develop a model for subhumid-semiarid interflow processes. In addition, this approach enables the processes controlling interflow in subhumid-semiarid regions to be compared with those in humid regions, thereby increasing our understanding of interflow generation across a wider range of environmental conditions.

METHODS AND MATERIALS

Site Description

The 870-m^2 ponderosa-pine hillslope study area lies within the Los Alamos National Laboratory's Environmental Research Park on the Pajarito Plateau of north-central New Mexico; it is described in detail in Wilcox et al. (1995) (Figure 2.1). The site slopes gently (average 6%) and drains into a nearby canyon. The elevation is 2,315 m and the average

annual precipitation is 510 mm (Bowen, 1990), just above the subhumid-semiarid boundary of 500 mm. The depth to groundwater is about 250 m.

The stratigraphy of the site was described by Watt and McFadden (1993) and Davenport (1994), and consists of Bandelier tuff (R horizon) at the base; a dense, smectite clay Bt horizon; and, at the top, sandy-loam (A and AB) horizons (Figure 2.2). The Bt horizon shows well-developed soil structure and contains root channels, cracks, and voids between ped faces. The sandy-loam horizons, which are both loess, are hereafter collectively referred to as the A horizon. Measured saturated hydraulic conductivities are 5.7×10^{-9} to 7.5×10^{-7} m/sec for the A horizon, 2.5×10^{-10} m/sec for the B horizon, and 1.3×10^{-9} to 7.1×10^{-9} m/sec in for the weathered tuff (Stephens, 1993). The hydraulic conductivity of unweathered Bandelier tuff ranges from 2×10^{-7} to 2.35×10^{-6} m/sec (Abeele et al., 1981).

A trench, 16 m x 2 m x 1.5 m deep, was dug across the bottom of the hillslope, perpendicular to the slope of the hill. It is equipped with two collectors so that water can be collected separately from the A and B horizons (Figure 2.2). A French drain at the bottom of the trench (not shown in the figure) collects water from a portion of the tuff. The A- and B-horizon collectors drain into separate stilling wells, in which pressure

transducers electronically measure the interflow produced at least every 15 min. A meteorological station at the site continuously records temperature, humidity, wind speed and direction, and precipitation. Soil moisture is monitored on a weekly basis, by neutron thermalization, at 10 locations on the hillslope. Additional information on the hydrometric data collection system can be found in Wilcox et al. (1996).

Sampling and Analytical Methods

Collection of hydrometric data from the hillslope has been ongoing since November 1992. Stable isotope and chloride tracer sampling began in June 1993 and included samples of interflow, precipitation and the bulk soils.

Interflow and Precipitation Samples. To obtain interflow samples, small-volume (about 50-mL) PVC collectors were inserted into the pipes that feed the stilling wells; the collectors were designed to minimize evaporation, which would adversely affect the stable isotope results. Precipitation samples were collected with a large, polyethylene funnel that drained into a 1/4-in.-diameter tube which had an elbow bend. After a rain, the bend held a plug of water preventing evaporation from the main reservoir. Water drained from the tube into a 1-L polyethylene

bottle having an overflow spout which was also bent and held a plug of water once the bottle overflowed.

The samples were usually collected on a daily basis, and two duplicate samples were taken whenever possible; one was analyzed for stable isotopes, the other for chloride. Samples were stored in 10- or 20-mL glass vials with polyseal caps (vials were rinsed with a small amount of sample before being filled). Samples were then stored in a refrigerator, at 4 °C, pending analysis.

Stable isotope analyses were conducted at the New Mexico Tech and Southern Methodist University stable isotope laboratories. The hydrogen and oxygen isotopes are reported in delta (δ) notation, as per mil (‰) differences relative to the international standard SMOW (Craig, 1961):

$$\delta D \text{ or } \delta^{18}O = \left[\frac{R_{sample} - R_{SMOW}}{R_{SMOW}}\right] \times 1000, \qquad (1)$$

where R is the D/H or $^{18}O/^{16}O$ ratio. The $\delta^{18}O$ analyses were based on the CO_2 equilibration method (Socki et al., 1992). The δD analyses were based on the uranium method (Bigeleisen et al., 1952). Sample splits were analyzed at both laboratories to ensure consistency of the data. Variation in $\delta^{18}O$ of sample splits analyzed at both laboratories was less

than 0.2 ‰. All of the δD analyses were performed at the Southern Methodist laboratory, so no interlaboratory comparison of δD analyses was done. Analytical precision was better than 0.2‰ and 2‰ for the $\delta^{18}O$ and δD analyses, respectively.

Chloride concentrations were determined using a Dionex ion chromatograph. A 1.08-mM Na_2CO_3/1.02 mM $NaHCO_3$ eluant was used with a Dionex AS4A column and self-regenerating suppressor. Calibration curves were established on the basis of five standards prepared by serial dilution from an NIST Standard chloride solution. After every five analyses, one standard and deionized water (DI) blank were run, and after every ten samples, duplicate sample analyses were run, to ensure that adequate accuracy and precision were maintained. Accuracy was 10% or better based on the periodic analyses of standards, and precision was better than 2%.

Soil Cores. Soil water chloride concentrations were determined from cores taken through the entire soil profile, in July 1993, August 1994, and June 1995. For each core, soil stratigraphy was described; the core was then split into 10-cm lengths and stored in clean mason jars or zip-lock bags. Latex gloves were worn at all times to prevent contamination of the samples. At the Los Alamos Environmental Science Group Laboratory, the core samples were first air-dried for 48 hours, and then

100-g splits were mixed with 100 mL of 17-mΩ DI water. The solutions were stirred with a glass stirring rod and allowed to equilibrate for 48 hours. A control, consisting of a beaker filled with 100 mL of the DI water, was prepared for every six soil samples. After equilibration, the solutions (leachates) were decanted, centrifuged, and filtered using disposable 0.2-μm Gelman ion chromatography filters. Leachates were analyzed for chloride using the same ion chromatography procedure described earlier for the interflow and precipitation samples. Soil moisture contents were determined gravimetrically or by neutron probe. The soil bulk density values that are needed to calculate the soil water chloride concentrations from the leachate concentrations were measured previously and are given in Wilcox et al. (1996).

Mixing Models. A simple mass balance mixing model approach was used to estimate old/new water percentages and matrix/preferential flow percentages. Stable isotope data were used in the old/new water estimates, and chloride data were used in the matrix/preferential flow estimates. Both chemical concentrations and isotope delta values can be used in this approach because the equations are based on differences between values, thus absolute quantities are not required.

Old and new water percentages were calculated from isotope values using the two component-mixing equations of Pearce et al. (1986):

$$Q_o = \left[\frac{C_i \quad C_n}{C_o - C_n}\right] Q_i \qquad (2)$$

and

$$Q_n = Q_i - Q_o \qquad (3).$$

where Q is discharge, C is tracer concentration, and the subscripts i, o, and n correspond, respectively, to interflow, old water, and new water. C_o was obtained from the interflow event immediately preceding the event of interest. C_n was obtained from the precipitation sample. Note that this determination of old/new water percentages assumes that the bulk samples are reasonably representative. In other words, it has been assumed that the failure to take intrastorm variations into account in the C_n value, and assuming that the isotopic value of the preceding interflow event represents old water will not result in errors large enough to cause misleading estimates of old and new water percentages. To partially compensate for this potential source of error, estimates were made only for days when there was greater than a 0.5‰ $\delta^{18}O$ or 15‰ δD difference between the isotopic values of old and new water. Because

there is no groundwater or streamflow at the hillslope, errors cannot be attributed to those components.

To estimate the relative percentages of matrix and preferential flow, a mass balance approach similar to that of Sharma and Hughes (1985) was used. The relative proportions of preferential and matrix flow were determined as follows:

$$I = I_p + I_M \tag{4}$$

and

$$IC_I = I_P C_P + I_M C_M, \tag{5}$$

where I is the interflow rate measured at the collection point, C_I is the interflow concentration at the collection point, I_P is the preferential flow rate, C_P is the preferential flow water concentration, I_M is the matrix interflow rate, and C_M is the matrix water concentration. If I, C_I, C_P, and C_M are known, the matrix interflow rate can be determined as follows:

$$I_m = I\left(\frac{C_I \quad C_P}{C_M - C_P}\right) \tag{6}$$

This approach assumes (1) conservation of mass and solutes in the soil horizon, for example, no vertical losses to the underlying soil; and (2) no interaction between the preferential flow paths and macropores. The C_p values were estimated from typical low chloride concentrations in A- or

B-horizon interflow during unsaturated conditions. C_M was estimated on the basis of average chloride concentrations from the cores.

RESULTS

Hydrometric Results

The hydrometric results for the hillslope are discussed in detail in Wilcox et al. (1996), so only a brief overview and update are given here. Interflow from the hillslope is episodic, the largest events generally occur in the spring as a result of saturated conditions from melting snow or spring rainfall. Most of this water (83%) flows through the B horizon, the balance through the A horizon. Interflow volumes from the A and B horizons and precipitation for the period February 1993 through June 1995 are shown in Figure 2.3. Other than in the spring and summer of 1993, periods of continuous A-horizon interflow generation were generally short, compared to that of the B-horizon. In the three years of observation, there were three periods during which combined A- and B-horizon interflow rates exceeded 200 L/day, spring 1994, fall 1994, and spring 1995 (Figure 2.3). Small quantities of interflow were measured at other times, generated by individual storms or fronts. Although it makes up only a small portion of the annual water budget (≈2%), interflow can be important for shorter periods; for example, in the 1993 water year

(defined by Wilcox et al., 1996 as November to October), interflow accounted for 19% of the winter-spring water budget.

Stable Isotope Results

A $\delta^{18}O$-δD plot (Figure 2.4) comparing A- and B-horizon interflow to the Los Alamos local meteoric waterline (LMWL) of Vuataz and Goff (1986) shows that very little evaporation of the interflow waters occurs, because all of the data fall on or near the LMWL. The close agreement between interflow water and the LMWL shows, in addition, that sample integrity was preserved during the sampling process and that the analyses are of good quality.

For the A horizon, the $\delta^{18}O$ of interflow and of precipitation from June 1993 to April 1995 are shown in Figure 2.5. The variability of A-horizon interflow $\delta^{18}O$ values is much less than that of precipitation. Percentages of old and new water, calculated from equations 2 and 3, are shown in Table 2.1a for nine days of interflow. For the small volume interflow events (<0.5 L/day) that occurred during unsaturated conditions, old water dominated; new-water contributions were large (>30%) only on July 15, 1993 and February 15, 1995 when large rain or snowmelt events occurred. Unfortunately, for the very large interflow event in March 1995, which accounted for the vast majority of A-horizon

interflow generated over the sampling period, old/new water percentages could not be determined. A new water sample was not collected because the event was initiated by a thaw and not enough snowmelt was generated produce sufficient surface runoff for sampling.

For the B horizon, the $\delta^{18}O$ of interflow and of precipitation for the period June 1993-April 1995 are given in Figure 2.6. As in the A horizon, B-horizon interflow $\delta^{18}O$ values do not show as large a seasonal variation as precipitation. Some isotopically distinct precipitation events produced fast changes in the $\delta^{18}O$ values of B-horizon interflow, (e.g., the October 15, 1994 period, Figure 2.6), while other isotopically distinct events produced little change at all (e.g. events between November 12, 1994 and March 6, 1995). This difference in the effect of various storms on interflow isotopic composition is probably related to the volume of precipitation and antecedent moisture conditions. Percentages of old and new water, determined for 15 days of interflow, are shown in Table 2.1b. As in the A horizon, old water dominates in small volume events. New water contributions were large (>30%) only on days with low antecedent soil-moisture-contents and prolonged rain events or when the soil was at or near saturation. For the large March 1995 event, the lack of a new water sample prevented calculation of old/new water percentages. However, if we assume that the isotopic composition of A-

horizon interflow is approximately representative of new water than old/new percentages can be estimated (Table 2.1). The result indicates that new water was dominant in this 1400 L/day event. Although the result has a very high uncertainty, the new water dominance seems reasonable given the fact that B-horizon interflow became 0.7% lighter than the old water value, and because chloride concentrations decreased substantially (the chloride results are discussed in the next section).

Old and new water percentages for four B-horizon interflow events, calculated on the basis of δD are shown in Table 2.1C. Comparison with the $\delta^{18}O$ results for those events in Table 2.1b shows that the percentages are similar (within 20%).

Finally, a comparison of old-water percentages for eight dates on which both A- and B-horizon interflow were measured is shown in Table 2.1d. With two minor exceptions, A-horizon interflow has a higher proportion of new water than does B-horizon interflow.

Because of the assumptions discussed in the Methods section regarding the selection of parameter values for the old/new water estimates, it needs to be noted that the results in Table 2.1 have large and unquantifiable errors. For example, some estimates show greater than 100% old water, a physical impossibility. Thus, it needs to be kept in mind that while the results probably show the general trends in

old/new water percentages, there is considerable uncertainty about the values themselves.

Chloride Results

Variations in volumetric moisture content and in chloride concentrations with time are shown for the A horizon in Figure 2.7. Prior to the very wet fall 1994-spring 1995 period, chloride concentrations usually rose with increases in moisture content (e.g., Figure 2.7, July-August 1993 peak) and then declined as moisture content decreased. The maximum chloride concentration was measured in March 1995, when the A horizon became saturated. Matrix/preferential flow estimates for the A horizon are shown in Table 2.2 and suggest that preferential flow was dominant during the relatively dry 1993-1994 period, whereas during the wet period in 1995, matrix flow was dominant. One problem with this rather simple and crude method is that dilution of matrix soil waters by infiltrating water is not considered. In other words, the concentrations from the soil cores were assumed to be representative of the matrix water. It is likely that infiltrating water in the soil matrix would dilute this concentration which makes the matrix concentration input values in the mixing model unrepresentatively high. Dilution does not affect the interflow

concentration inputs because these concentrations are determined after any dilution has occurred. In order to attempt to account for the possibility of dilution of the matrix waters, the mixing model calculations were redone assuming a 50% dilution of the estimated matrix water concentrations (Table 2.2). This results in a doubling of the matrix flow percentages. The result is that during unsaturated periods prior to 1995, matrix flow still only contributes a small percentage if any of the total flow. However, when conditions were at or near saturation, matrix flow percentages increase dramatically suggesting that if significant dilution occurs, matrix flow may be dominant.

Variations in chloride concentration and volumetric moisture content for the B horizon are shown, for June 1993 through April 1995, in Figure 2.8. Between June 1993 and mid-October 1994, chloride concentrations in interflow were relatively low (below 10 mg/L), and moisture content remained below about 33%. Starting in mid-October 1994, the soils began to approach saturation; chloride concentrations increased dramatically, reaching peak levels about 6 weeks before the soils became completely saturated (March 1995). Chloride concentrations then dropped throughout the spring of 1995. Unlike their fairly steady rise, the decline in chloride concentrations was marked by sharp dips (e.g., February 12, and March 6, 1995–see arrows

in Figure 2.8). These correspond to large-volume, multiple-day interflow events during which chloride was diluted by new precipitation or snowmelt waters; they may also reflect increases in the rate of macropore flow, which would reduce the time available for diffusion of chloride from the soil matrix into the macropores. Nevertheless, in terms of total mass, more chloride was removed during the March and February events than during the rest of the June 1993-April 1995 period, because these discharges were so large.

Soil Core Results

Soil cores were taken in the summers of 1993, 1994, and 1995 from which soil water chloride concentrations were determined. Chloride concentrations increased nonlinearly with depth, and increases were especially large in the B and deeper horizons (Figure 2.9). Comparison of chloride concentrations in the cores with interflow concentrations shows that interflow was much more dilute than water in the soil matrix (Table 2.3). Additional details of the core results and determinations of vertical flux rates through the ponderosa pine soils using the soil water chloride concentrations can be found in Newman (1996).

DISCUSSION

Before a description of the conceptual model of interflow at the hillslope can be made, some discussion of how macropore flow might occur under unsaturated conditions is in order. According to McDonnell (1991), the minimal requirement for water flow via macropores is a flux density of rain that is greater than the hydraulic conductivity of the matrix. In other words, the rate of input of water exceeds that of infiltration into the matrix, causing the excess water to flow into macropores. Such a mechanism could explain the generation of both A- and B-horizon interflow under unsaturated conditions. However, in the case of the ponderosa pine hillslope, the excess water may be accumulating at the A/B interface as well as on the soil surface. The measured saturated hydraulic conductivity of the B-horizon clays is extremely low (2.5×10^{-10} m/sec) and could readily be exceeded by the flux of water through the A horizon. The excess water could then flow into B-horizon macropores and eventually be released as interflow; or, if the excess water does not encounter any macropores, it could pond on top of the B horizon, generating interflow in a portion of the A-horizon matrix.

The ponding hypothesis is supported by evidence that some matrix flow occurs in the A horizon. Matrix flow is suggested by: (1) varying

chloride concentrations shown by the multiple peaks in Figure 2.7, which are apparently caused by chloride accumulation during periods of no flow and flushing when flow begins; (2) the decrease in soil matrix water chloride concentrations over time (Table 2.3); and (3) the large matrix flow percentages calculated for certain days (Table 2.2). The infrequency and small volumes of A-horizon interflow during unsaturated periods suggest that ponding may be a small-scale or localized occurrence. Time-domain reflectometry probes were installed at the A/B horizon interface, and would have been valuable in determining whether ponding on the B horizon really occurs. However, because they never functioned properly, the ponding hypothesis remains in question.

It does not appear however, that A-horizon interflow is entirely matrix flow. Even though Table 2.2 suggests that matrix flow is important, preferential flow is evidently dominant at times. However, the sandy loam soil of the A horizon does not seem to be conducive for the formation of macropores. Field examination showed little macropore development, especially compared to the B horizon. Allison and Hughes (1983), working at a semiarid site in Australia, found that stemflow was an important preferential flow mechanism, facilitating rapid movement of water into the soil. Stemflow occurs in the area between plant stems or

roots and the surrounding soil. This same mechanism could be operating at the study site, given the considerable grass and tree cover.

In contrast to the ambiguity of how A-horizon interflow occurs, preferential flow, probably via macropores, is apparently dominant in the B horizon. This conclusion is supported by two of this study's findings. First, comparison of the $\delta^{18}O$ of precipitation with that of B-horizon interflow shows that new water can move through the system within 24 hours (see Figure 2.6, events of October 14 and November 12, 1994); such movement would not be consistent with matrix flow, given the low bulk hydraulic conductivities of the soils. Second, comparison of chloride concentrations in matrix soil water with those in interflow (Table 2.3) shows that the two are not equilibrated, which indicates that interflow water is bypassing the salt-rich matrix and moving through macropores. This is consistent with the suggestion of Thomas and Phillips (1979) and Luxmore et al. (1990) that macropores serve mainly as physical conduits and have only minor effects on aqueous chemistry. Other evidence that macropore flow can occur in the B horizon is the presence of shrinkage cracks, root channels, voids between soil peds, and mineral accumulation on the walls of some macropores.

The importance of macropore flow in the B horizon is further supported by the results in Table 2.2. The maximum matrix interflow

contribution apparently occurred on January 12, 1995, when the highest interflow chloride concentration of the study was recorded. Assuming a 50% dilution of the matrix water chloride concentrations indicates that matrix flow may have been dominant on that day. However, the rest of the results show that even with a 50% dilution factor, macropore flow is dominant even when the soils are saturated, as on March 6, 1995 (see Table 2.2).

It needs to be emphasized that interflow generation under unsaturated conditions represents only a very small fraction of the total interflow for the sampling period. Well over 90% of the total interflow generated occurred when the soils were at or near saturation. This period was also when most of the chloride was mobilized.

Conceptual Model of Interflow

The conceptual model describes interflow generation under two volumetric soil moisture regimes: (1) moisture content below approximately ≈33%; and (2) moisture content above ≈33%, and is illustrated in Figure 2.10. The ≈33% moisture content value is based on the interflow chloride data and appears to be a threshold above which major changes in interflow chemistry and old/new water percentages occur (see Figures 2.5 and 2.8, and Table 2.2). Stated another way,

moisture contents less than ≈33% represent interflow processes under unsaturated conditions, whereas moisture contents greater than ≈33% represent soil moisture conditions that are at or near saturation. Also, the 33% threshold is probably only applicable to the local area or sites with similar soils and probably cannot be universally applied to other systems.

When volumetric moisture content is below about 33%, infiltrating water is either removed by evapotranspiration or bypasses the A horizon matrix and enters the macropore system in the B horizon, where it then moves laterally. Bypassing of the A horizon matrix (which might occur by stemflow) is suggested because interflow waters are not subject to much evaporation (Figure 2.4), which indicates that they move relatively quickly through the shallow evaporative zone in the soil. Matrix bypassing is also suggested by the low chloride concentrations of B-horizon interflow (see June 1993-October 1994 period in Figure 2.8). When B-horizon interflow is generated under unsaturated conditions small amounts (<0.5 L/day) of dominantly old water interflow typically result. Under this moisture-content regime, the B horizon acts as a two-domain system: the macropore domain generates interflow, while in the matrix domain, water that manages to infiltrate moves very slowly and is subject to transpiration, which causes the chloride concentration to

increase. Any A-horizon interflow that occurs is very small in volume and may come from localized ponding on top of the B horizon. If intense or large-volume rains occur under this soil moisture regime, moderate volumes of B-horizon interflow can be generated (on the order of 30 L/day) that will be dominated by new water, as was observed in July 1993 and October 1994 (Table 2.1).

When the high moisture-content regime is in effect (and maintained longer than a few days), the apparent independence of the macropore- and matrix-flow domains disappears and the importance of new water increases as compared to typical interflow during unsaturated periods. Interflow volumes also rise dramatically and can constitute greater than 90% of the annual interflow volume (see Figure 2.3). Solute flushing and diffusion of chloride from the matrix increase, causing chloride concentrations in interflow to rise substantially (see Figures 2.7 and 2.8). It seems counter-intuitive that chloride (and probably other solute) concentrations would rise, rather than decline because of dilution, as the soils become wetter; however, in this region, the two-domain flow regime is in effect for most of the year, causing the soil matrix to act as a solute sink. Only when the soils are at or near saturation is there a continuous fluid phase that allows solutes to be transported out of the soil matrix.

Subhumid- vs. Humid-Region Interflow

When the results from this study are compared with those from humid regions (discussed in the Introduction), more similarities than differences are found. Macropore flow generation by the excess water mechanism appears to be important in both environments, effecting rapid movement of water through the soil. In addition, aspects of new/old water relationships appear to be similar. For example, Turton et al.'s (1995) finding that new water becomes important during infrequent large storms, while old water is dominant during the frequent small storms appears to hold for subhumid-semiarid environments as well. During unsaturated periods at the ponderosa pine site, large and infrequent storms produce substantial percentages of new water, while during the majority of the year, most of the interflow is old water.

However, during saturated periods, there is an apparent difference in old/water percentages at the ponderosa pine site compared to what has been observed at most humid sites. The humid sites show a dominance of old water, while the results here indicate a possible dominance or at least a much more important contribution of new water. However, this difference is probably related to the fact that there is no stream at the ponderosa pine site. Thus, the groundwater ridging mechanism suggested by Sklash and Farvolden (1979) does not apply.

Instead, it is likely that for large events, old water is readily flushed from the system allowing a substantial new water contribution to interflow.

Another difference between subhumid and humid interflow involves changes in interflow chemistry. In humid regions, evidence of solute flushing has been observed as the soils become wetter, however, changes in interflow chemistry appear to be much greater in subhumid environments. Studies in humid environments have shown changes in chloride concentrations on the order of a few mg/L (e.g., DeWalle and Pionke, 1994; Mulholland et al., 1990), whereas the flushing that occurs at or near saturation in subhumid environments produces changes on the order of 30-40 mg/L. These larger changes are caused by the tremendous salt enrichment that occurs during extended dry periods, when interflow is isolated in the macropores. With no mechanism for their removal, soluble species build up and are released in large quantities when saturated conditions are approached.

SUMMARY AND CONCLUSIONS

Stable isotope and chloride tracer results show that interflow in a subhumid ponderosa pine hillslope in New Mexico is largely controlled by preferential flow processes, which not only influences water movement, but dramatically affects soil water chemistry. Most of the

interflow is generated in the B horizon and travels mainly via macropores, which can result in extremely rapid water movement even in soils having very low bulk hydraulic conductivities. Throughout most of the year, the flow system has two domains: a macropore domain, in which water can move relatively rapidly and in which evapotranspiration has a minor effect; and a matrix domain, in which water movement is extremely slow and evapotranspiration is high, resulting in substantial water loss and increased salt concentrations. When the soils are at or near saturation (greater than ≈33% volumetric water content), a connection between the two domains is established, and concentrations of chloride and other aqueous species in interflow can rise dramatically. Under these conditions, very large volumes of interflow can be produced.

The interflow generation processes that have been identified in humid environments are also operational in this subhumid environment, but in subhumid regions the chemistry of interflow can change to a much greater extent than has been observed at humid sites. This potential for major changes in interflow chemistry may significantly influence the mobility of some contaminants.

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Table 2.1. Old and New Water Precentages for A- and B-Horizon Interflow

Table 2.1a. A Horizon $\delta^{18}O$

Date	Qs (L/day)	Cs ‰	Cn ‰	Co ‰	Qo (L/day)	% Old	% New
7/12/93	0.06	-12.4	-5.6	-12.6	0.06	98	2
7/13/93	0.02	-12.2	-9.3	-12.4	0.02	93	7
7/14/93	7.38	-12.0	-8.9	-12.2	6.94	94	6
7/15/93	0.24	-10.1	-9.9	-12.0	0.02	10	90
8/31/93	0.22	-8.9	-13.1	-8.8	0.21	97	3
7/12/94	0.02	-10.9	-2.2	-12.2	0.02	87	13
7/18/94	0.01	-9.7	-3.9	-10.4	0.01	90	10
11/13/94	0.35	-12.7	-6.4	-14.6	0.27	77	23
2/15/95	22.9	-13.6	-14.8	-12.9	14.45	63	37

Table 2.1b. B Horizon $\delta^{18}O$

Date	Qs (L/day)	Cs ‰	Cn ‰	Co ‰	Qo (L/day)	% Old	% New
7/13/93	0.02	-13.8	-9.3	-13.8	0.020	100	0
7/14/93	1	-13.5	-8.9	-13.8	1	93	7
7/15/93	0.02	-10.6	-9.9	-13.5	0.00	19	81
8/31/93	0.02	-8.9	-13.1	-8.6	0.02	94	6
7/12/94	0.05	-12.0	-2.2	-11.4	0.05	107	-7
7/18/94	0.02	-12.1	-3.9	-12.1	0.02	101	-1
8/18/94	0.01	-11.3	-3.4	-11.3	0.01	100	0
8/25/94	0.005	-10.3	-3.8	-10.3	0.005	100	0
10/15/94	61	-15.5	-18.3	-10.5	22	36	64
10/16/94	28	-12.1	-10.5	-16.2	8	28	72
10/17/94	18	-10.3	-10.6	-10.1	11	60	40
11/13/94	114	-12.3	-6.4	-13.6	93	81	19
2/15/95	26	-12.5	-14.8	-11.8	20	78	22
3/6/95[1]	1454	-12.6	-12.9	-11.9	484.6	33	67
4/12/95	0.25	-12.9	-16.7	-12.4	0.2	88	12

Table 2.1c. B Horizon δD

Date	Qs (L/day)	Cs ‰	Cn ‰	Co ‰	Qo (L/day)	% Old	% New
7/13/93	0.02	-99	-78	-97	0.02	110	-10
8/18/94	0.01	-82	-32	-80	0.01	103	-3
10/15/94	61	-112	-126	-75	17	28	72
10/16/94	28	-86	-64	-112	13	47	53
10/17/94	18	-75	-63	-86	9	51	49

Table 2.1d. Comparison of A- and B-horizon % old water (estimated from $\delta^{18}O$).

Date	A % Old	B % Old
7/13/93	93	100
7/14/93	94	93
7/15/93	10	19
8/31/93	97	94
7/12/94	87	107
7/18/94	90	101
11/13/94	77	81
2/15/95	63	78

[1]3/6/95 New water value was estimated from the A-horizon, because there was no precipitation to sample.

Table 2.2. Calculation of matrix flow contribution to interflow.

A-Horizon Interflow

				No Dilution			50% Dilution of C_m		
Date	I (L/day)	C_I (mg/L)	C_P (mg/L)	C_M (mg/L)	I_M (L/day)	% matrix flow	50%C_M (mg/L)	I_M (L/day)	% matrix flow
7/19/93	0.03	3.2	2	101	0.0004	1.2	50.5	0.0007	2
8/16/94	0.01	2.5	2	58	0.0001	0.9	29	0.0002	2
3/6/95	144	24	2	58	57	39	29	117	81
3/13/95	0.11	33	2	58	0.06	55	29	0.126	115
6/12/95	0.01	9.2	2	18	0.005	45	9	0.01	103

B-Horizon Interflow

				No Dilution			50% Dilution of C_m		
Date	I (L/day)	C_I (mg/L)	C_P (mg/L)	C_M (mg/L)	I_M (L/day)	% matrix flow	50%CM (mg/L)	I_M (L/day)	% matrix flow
7/19/93	0.06	2.3	2	206	0.0001	0.1	103	0.0002	0.3
8/16/94	0.02	2.8	2	206	0.0001	0.4	103	0.0002	0.8
1/12/95	0.01	63	2	206	0.003	30	103	0.006	60
3/6/95	1400	23	2	206	144	10	103	291	21
6/12/95	0.01	29	2	206	0.001	13	103	0.003	27

Table 2.3. Comparison of chloride concentrations in interflow and soil matrix water.

A Horizon		
Date	Interflow Cl^- (mg/L)	Matrix Cl^- (mg/L)[1]
7/19/93	3.2	101 (59-184)
8/16/94	2.5	58.4 (32-113)
6/12/95	9.2	20 (14-33)

B Horizon		
Date	Interflow Cl^- (mg/L)	Matrix Cl^- (mg/L)[2]
7/19/93	2.3	216 (55-455)
8/16/94	2.8	329 (161 593)
6/12/95	29	298 (106 477)

[1] Average and range of concentrations of the A horizon in the core.

[2] Average and range of concentrations of the B horizon in the core.

All cores were taken in the same part of the hillslope.

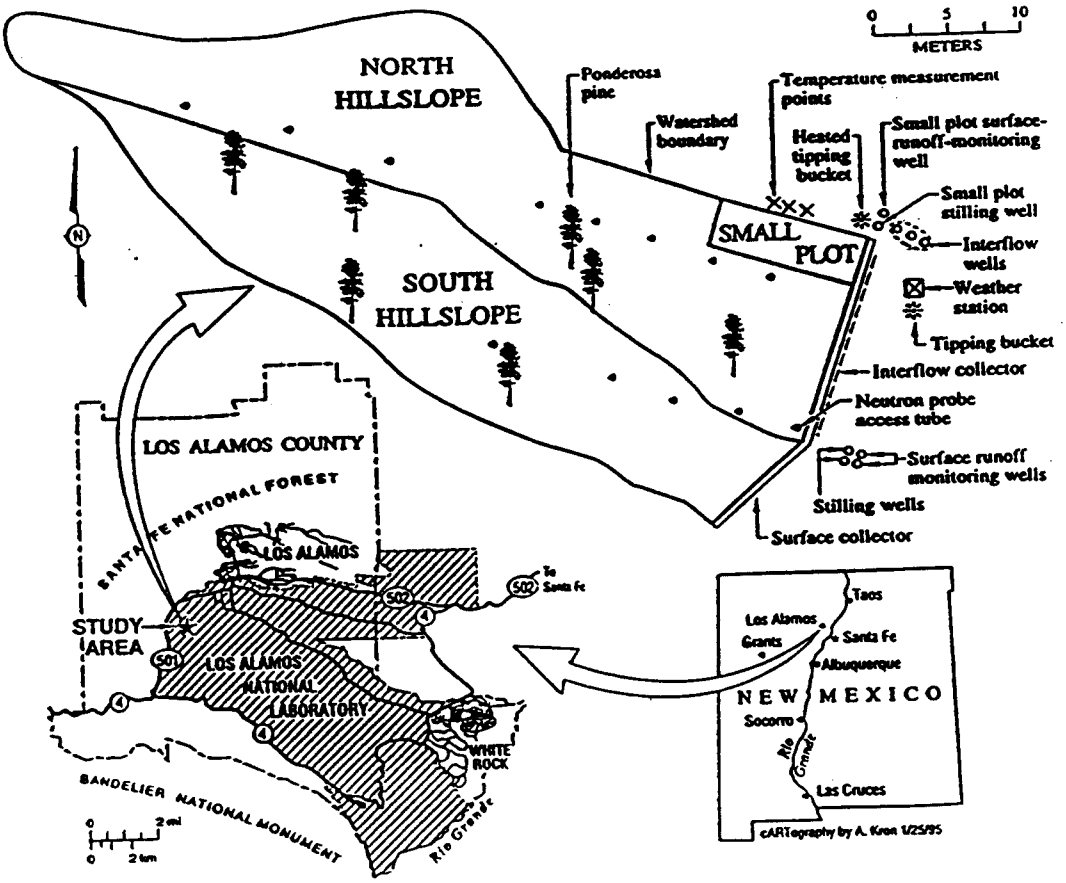

Figure 2.1. Location map and schematic of the ponderosa pine hillslope.

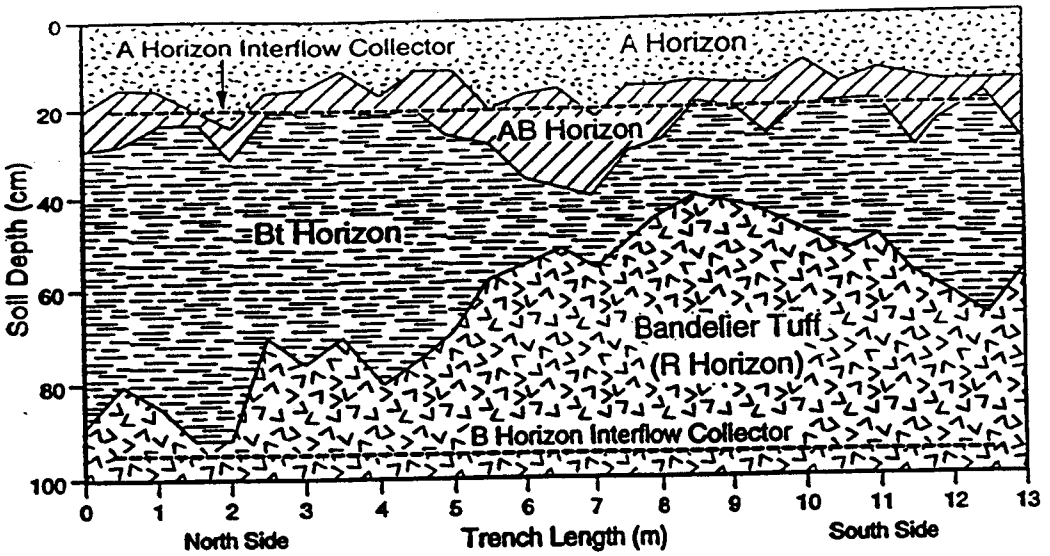

Figure 2.2. Diagram of the interflow collection trench showing soil stratigraphy and location of the interflow collectors.

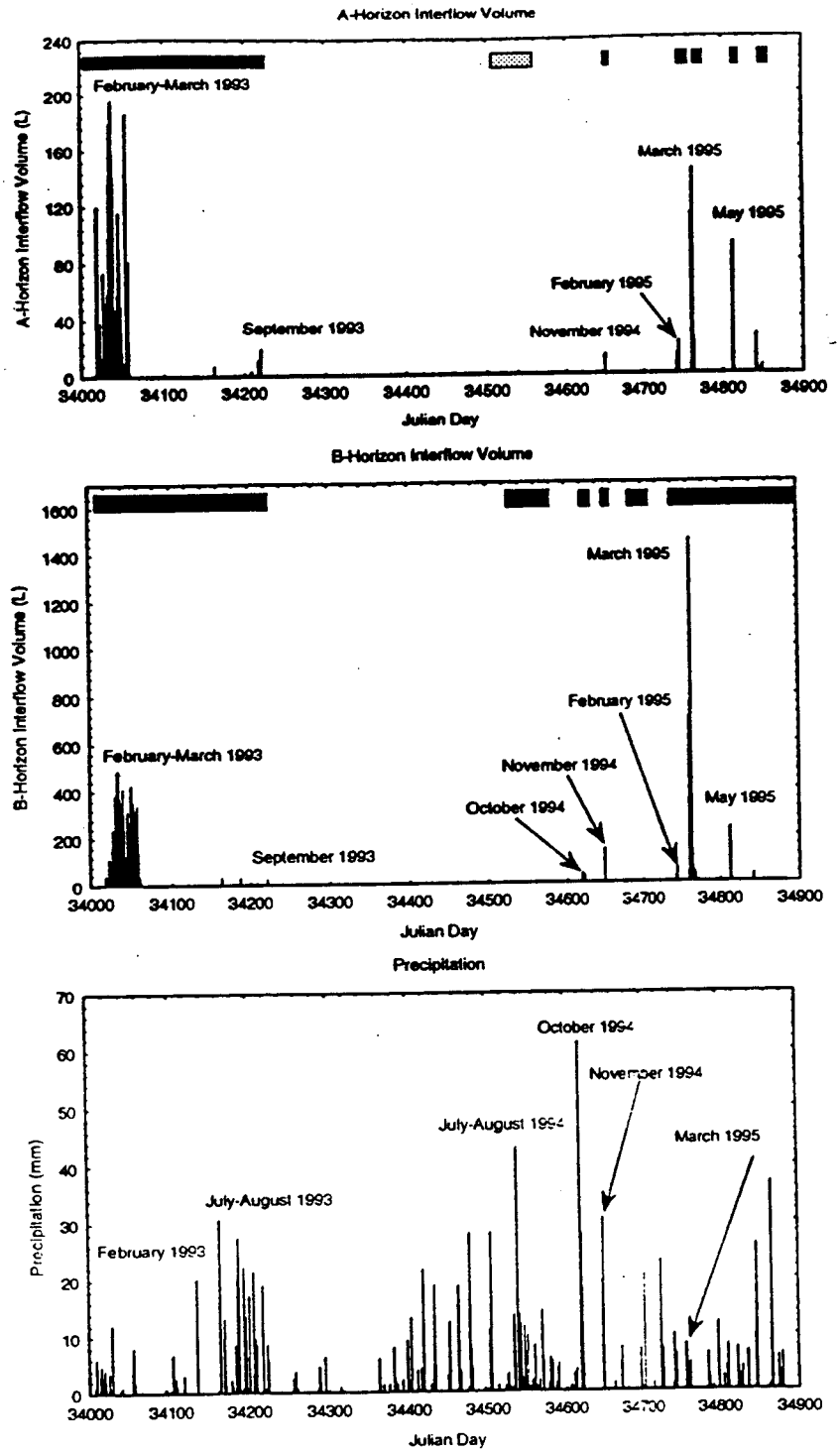

Figure 2.3. Graphs of A- and B-horizon interflow volume, and precipitation. Solid bars at the top of the interflow graphs indicate periods of continuous interflow generation. Stippled bars indicate periods of frequent, but discontinuous interflow generation.

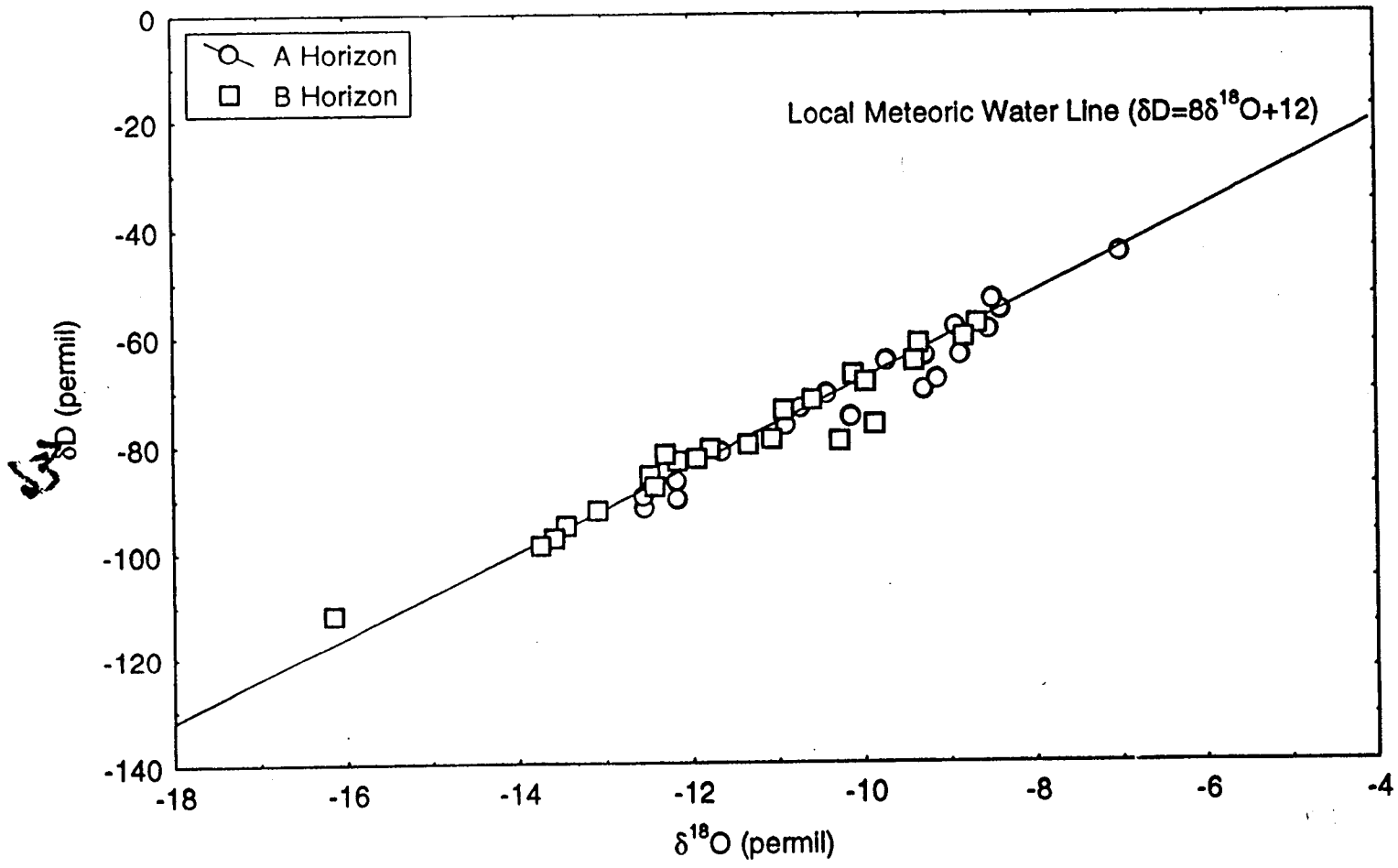

Figure 2.4. $\delta^{18}O$ versus δD for A- and B-horizon interflow plotted against the Los Alamos local meteoric water line of Vuataz and Goff (1986).

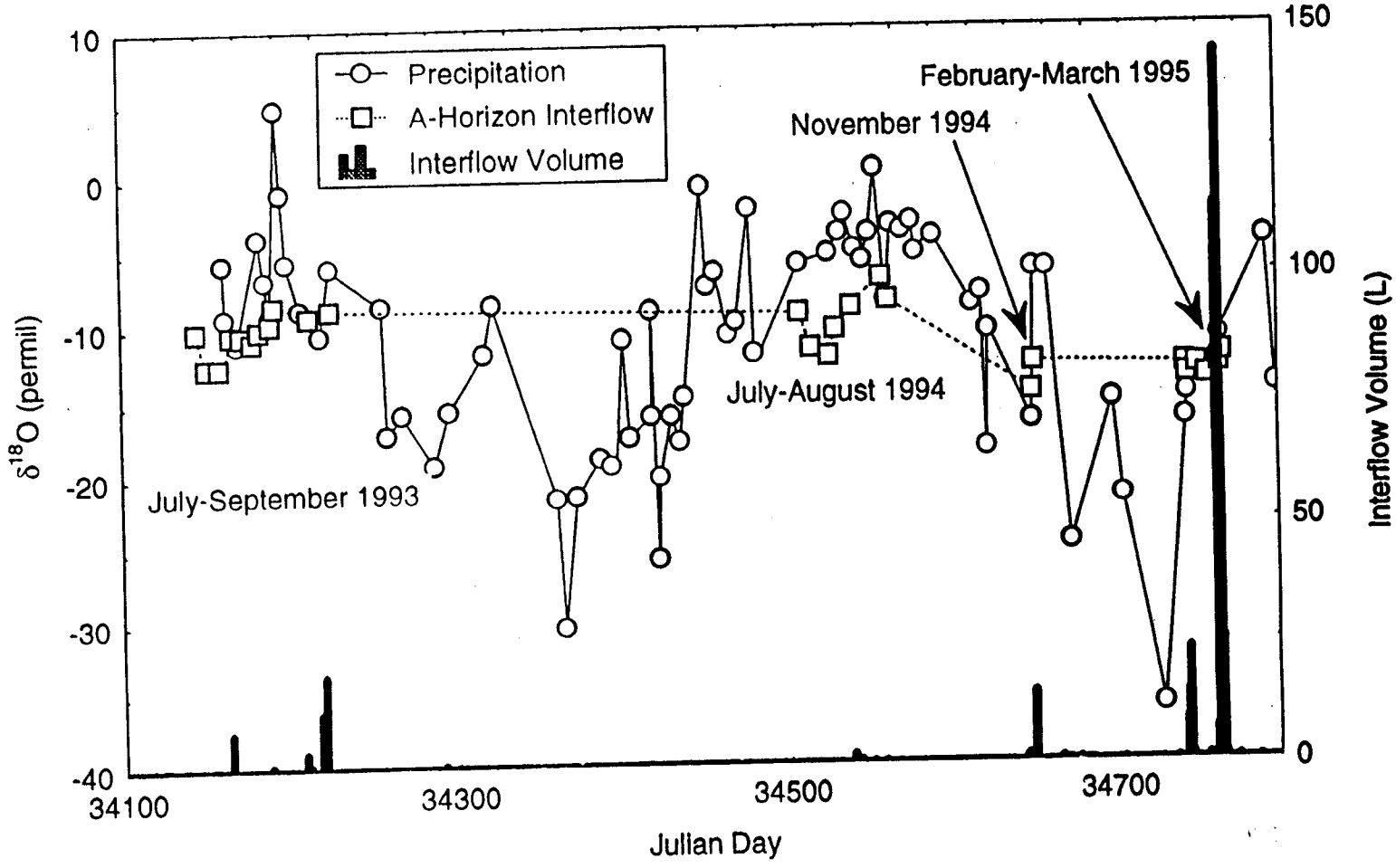

Figure 2.5. Changes in the δ^{18}O of precipitation and A-horizon interflow through time. A-horizon interflow volume is also shown.

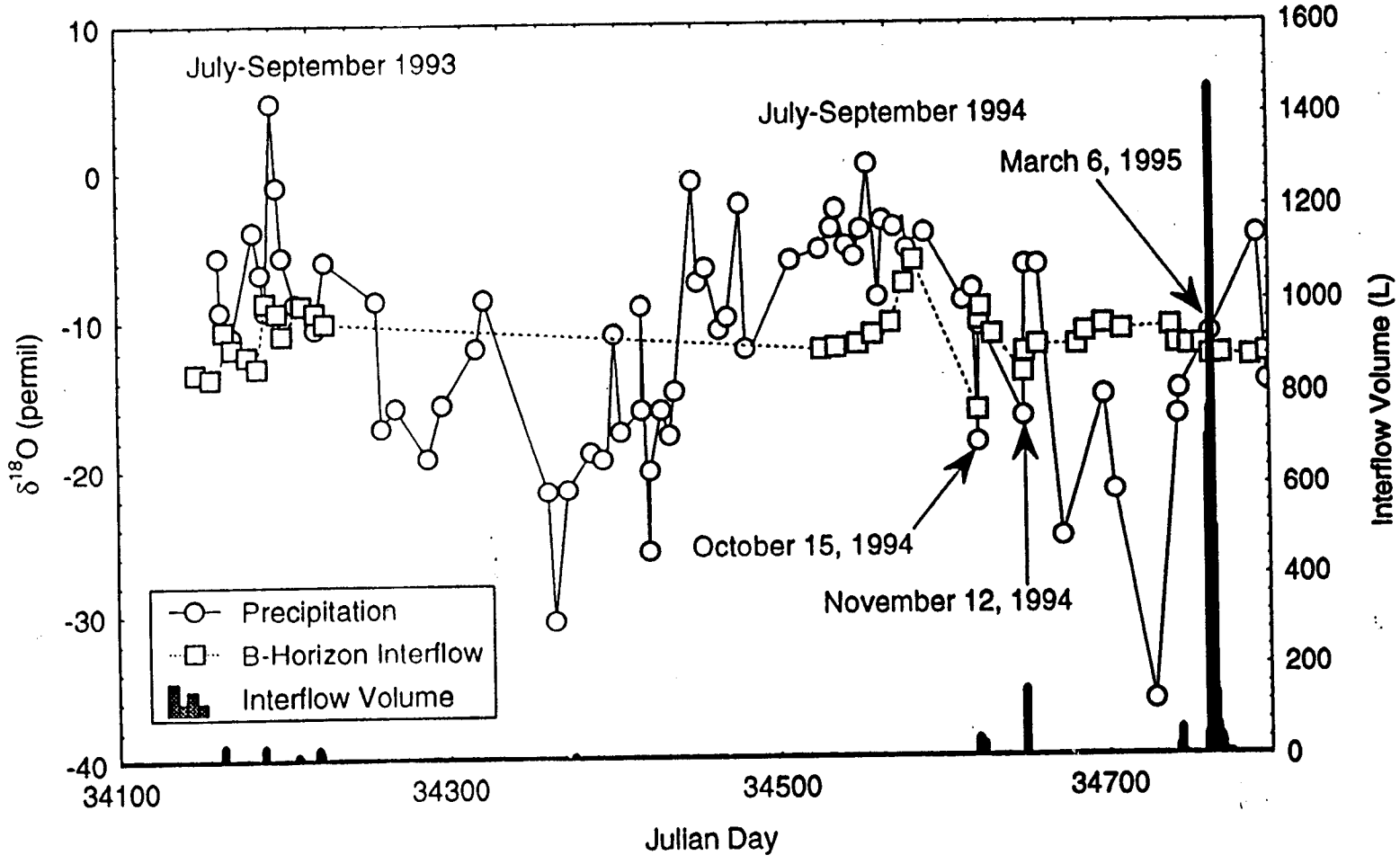

Figure 2.6. Changes in the $\delta^{18}O$ of precipitation and B-horizon interflow through time. B-horizon interflow volume is also shown.

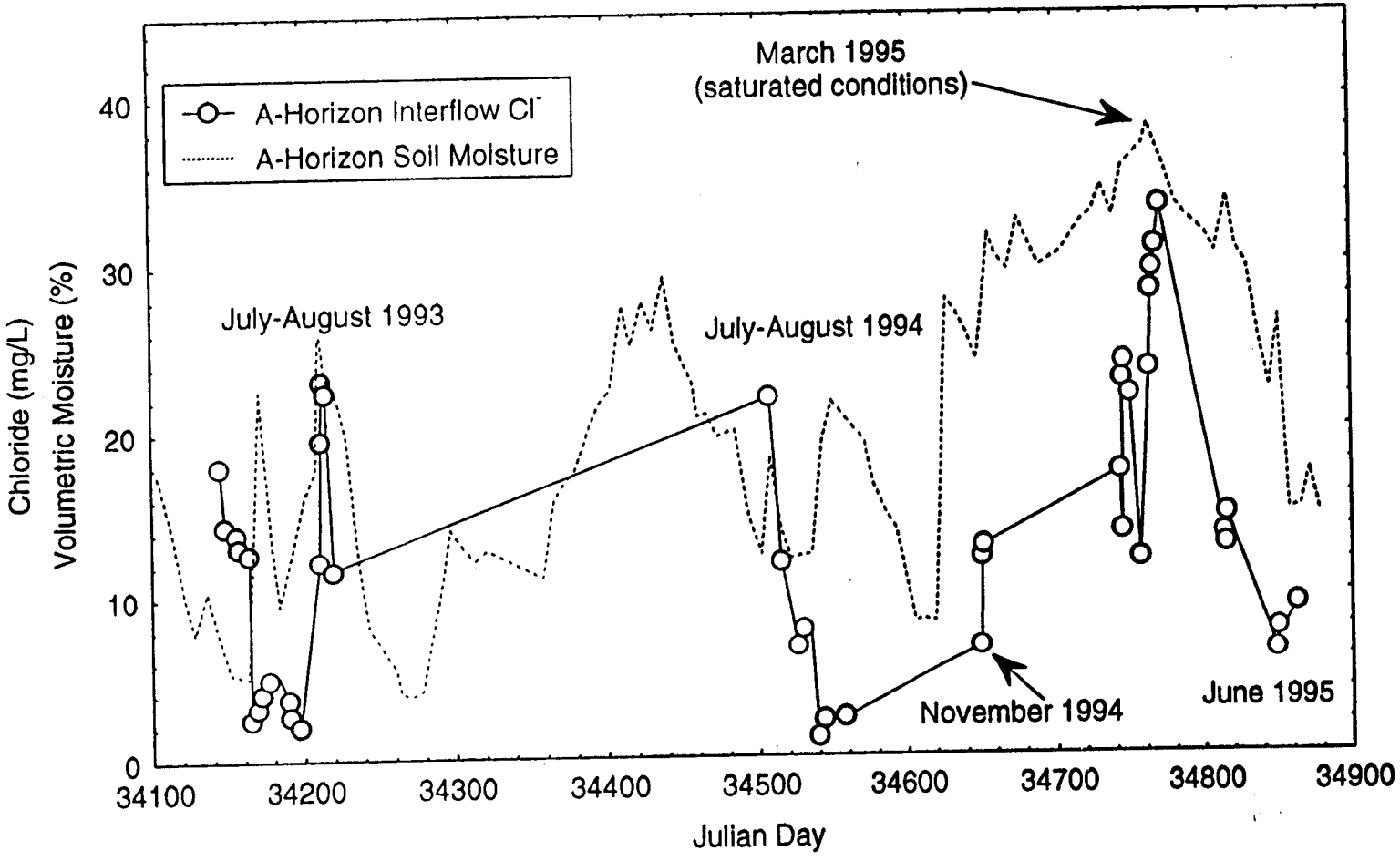

Figure 2.7. Changes in A-horizon interflow chloride concentration and volumetric moisture content through time. Volumetric moisture is the average for all neutron access tubes on the hillslope for the horizon.

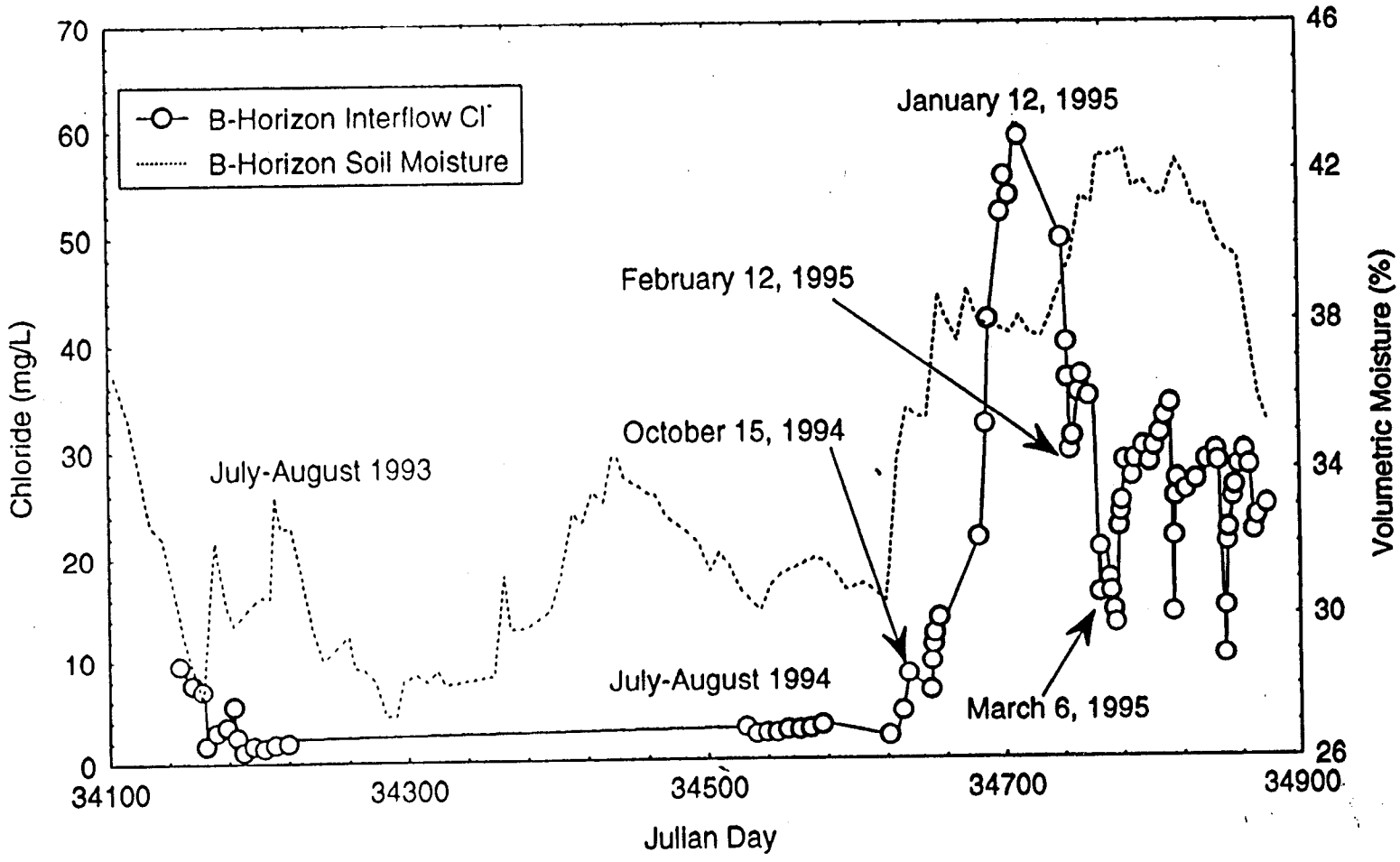

Figure 2.8. Changes in B-horizon interflow chloride concentration and volumetric moisture content through time. Volumetric moisture is the average for all neutron access tubes on the hillslope for the horizon.

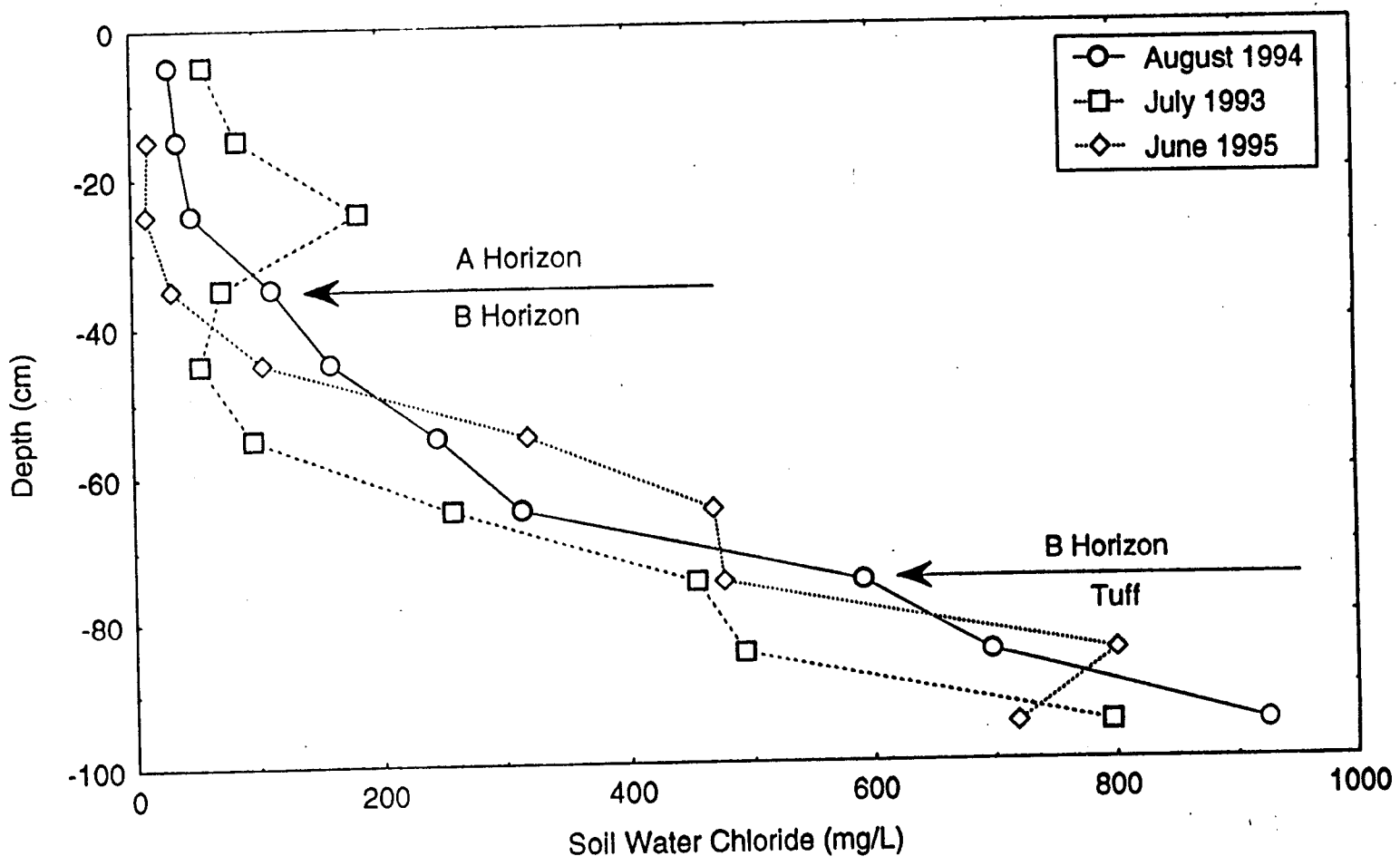

Figure 2.9. Chloride soil-water concentration profiles with depth from soil cores taken in July 1993, August 1994, and June 1995. Indicated depths of the soil horizon interfaces are approximate because of small differences in interface depths between the three cores.

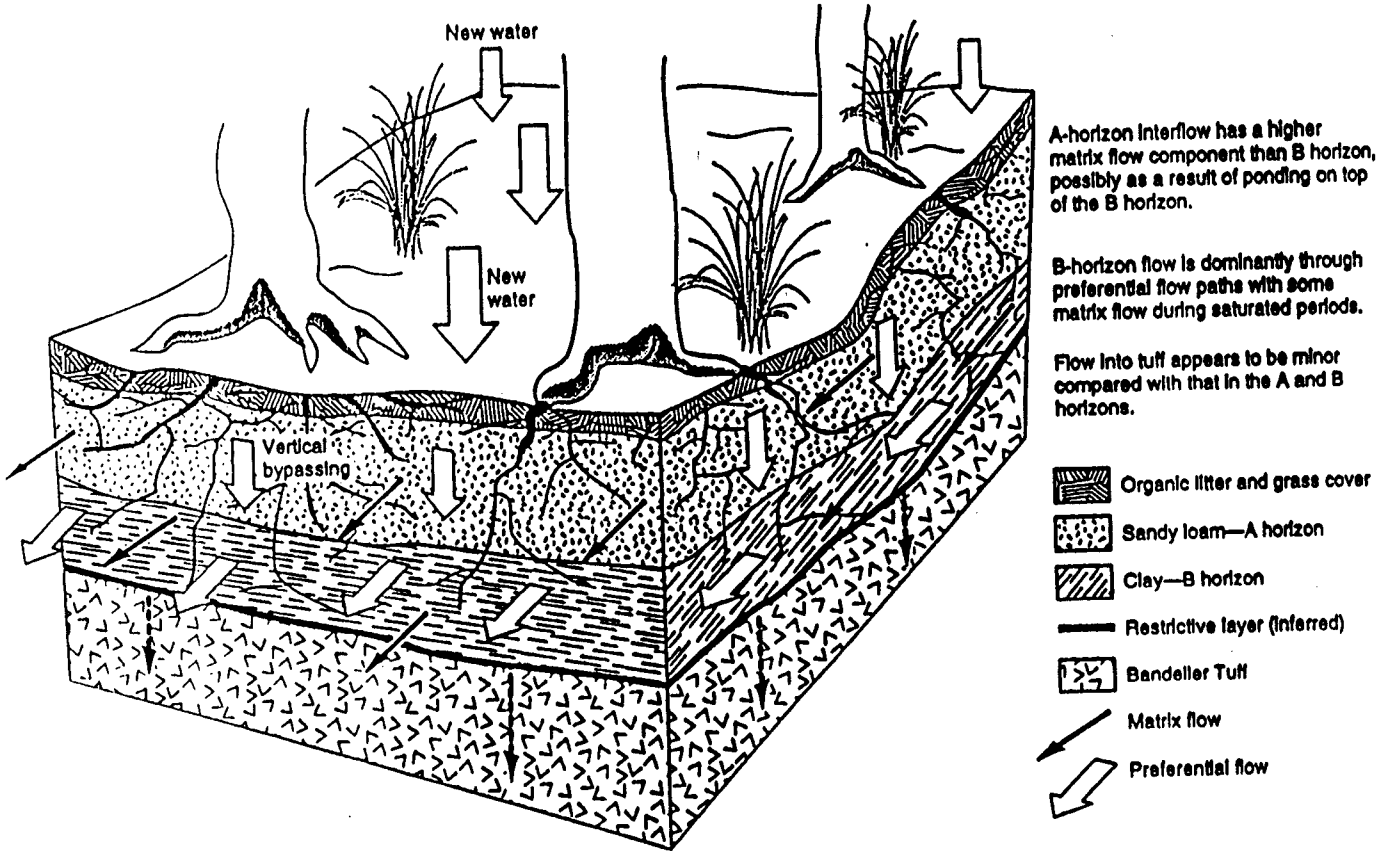

Figure 2.10. Illustration of the conceptual flow model for the hillslope.