Subsurface Flow of a Forested Riparian Area in the Oregon Coast Range

by

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AN ABSTRACT OF THE THESIS OF

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This study was undertaken to gain further understanding of the subsurface hydrology for a stream-adjacent riparian area in Western Oregon's Coast Range. Spatial and temporal fluctuations of the free water surface of a toeslope, adjacent riparian area, and stream channel in a forested terrace reach were monitored over a period of one year. A total of 27 piezometers were installed (nine in each of three transects established perpendicular to the stream) in a stream-adjacent terrace and hillslope. These piezometers were monitored from September 26, 1989 to July 25, 1990.

Results indicate that the direction of flow within a forested terrace can vary throughout the year. During the drier months of September and October, flow direction was generally towards the stream. However, by November the direction of flow for those areas closest to the stream had begun to change. In some cases the direction of flow changed up to 180° from the September direction.

The location of influent/effluent zones along the stream (i.e., zones along the stream where the terrace has a lower or higher hydraulic head than the stream, respectively) also varied throughout the year. During October, 1989 the furthest upvalley stream-adjacent piezometer had hydraulic heads greater than the stream for the entire month, while two other stream adjacent piezometers did not. During normal precipitation in January, 1990 both furthest upvalley and the middle stream-adjacent piezometers had hydraulic heads greater than the stream for the entire month. By July, 1990 only the middle stream-adjacent piezometer had a hydraulic head greater than the stream for the entire month, while the furthest upvalley piezometer had a higher head for 33% of the month and the hydraulic head of furthest downvalley piezometer never exceeded that of the stream.

For the Deer Creek study site, subsurface velocities were estimated to be in the range of 10^{-8} to 10^{-11} m/s resulting in a 3.8 year minimum travel time for storm water to reach the stream. Thus, soil matrix velocities using the Darcy equation were not sufficient to generate stormflow. However, no overland flow was observed during storm events, indicating that alternative subsurface flow pathways, such as macrochannels, are being utilized.

Previous research has suggested that the release of water from terrace storage in small headwater streams is not sufficient to maintain baseflow. The results of the

Deer Creek study support this conclusion. Furthermore, the results of this study indicate that forested riparian areas are hydrologically complex with respect to both space and time and that oversimplification of these systems may lead to misinterpretation of other processes associated with subsurface water dynamics.

APPROVED:

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Subsurface Flow of a Forested Riparian Area in the Oregon Coast Range

INTRODUCTION

Riparian areas are receiving increased attention from land managers, researchers and the general public. A large part of this attention stems from the recognition that riparian areas have the potential to influence large woody debris recruitment, stream temperature, streambank stability and nutrient cycling.

A component of riparian areas that is critical in influencing many riparian physical characteristics and functions is the occurrence and dynamics of subsurface water. Water table dynamics of stream-adjacent areas have the potential of influencing soil genesis, nutrient cycling, plant communities, stormflow and baseflow, yet little is known about the hydrologic character of these areas.

The focus of previous research on forested watershed stormflow and baseflow generation generally assumes a uniform hydrologic link between the hillslope and stream. However, where terraces, floodplains, or alluvial fans occur this assumption may not necessarily be valid. These areas may be a source of complex hydrological interactions between several sources of incoming and outgoing water including precipitation, streamflow, shallow subsurface flow, and deep groundwater flow. In addition, many watershed stormflow models assume a uniform mechanism of stormflow over the entire watershed, whether it be the partial area, variable source area, subsurface flow, or the groundwater ridge theory. This variety of findings does not indicate that these mechanisms are exclusive or opposing but rather suggests that under varying spatial and temporal circumstances several mechanisms may be occurring simultaneously.

Management activities such as road building, selective harvesting of riparian areas, buffer strip design, and upslope harvesting have the potential of impacting the hydrology of riparian areas. Some federal agencies specify a uniform width depending on various parameters, including stream width and order. The hydrologic implications of alternative types of riparian protection, however, have not been documented.

This study will help improve our ability to delineate riparian areas based on an understanding of hydrologic functioning of these areas. A knowledge of subsurface hydrology could aid in designing buffer strips of varying widths instead of simply assigning widths. Varying width buffer strips would allow for a wider buffer in areas where the water table is close to the surface and tree roots are more likely to be shallow, thereby creating a more wind-firm outer edge for the buffer strip.

The emerging field of riparian silviculture is beginning to address the problem of regeneration within riparian areas so as to insure a future source of large wood to streams. Knowledge of riparian subsurface hydrology will provide riparian silviculturalists with a hydrologic basis for predicting the regeneration and growth potential of various riparian tree and shrub species. Thus, an understanding of the temporal and spatial characteristics of riparian subsurface hydrology will have important research and management implications.

OBJECTIVES

The intent of this study is to provide information to help understand the spatial and temporal hydrologic functioning of the "hillslope-riparian area-stream" continuum within a watershed. This study monitored spatial and temporal fluctuations of the free water surface of a toeslope, adjacent riparian area, and stream channel in a forested terrace reach for a period of one water year.

To allow specific analytical procedures to be used and to focus the general objective of this study, five questions were identified:

(1) What is the apparent direction of flow of subsurface water within the hillslope-riparian area-stream continuum throughout the year and during storm events?

(2) Are subsurface water velocities through a saturated soil matrix sufficient to generate stormflow in the Deer Creek study site?

(3) Does vertical infiltration from precipitation account for the increased hydraulic head of terrace piezometers during storm events?

(4) Throughout the year does the stream recharge the terrace or does the terrace recharge the stream?

(5) Do stream adjacent terraces in small forested watersheds store sufficient water for sustaining baseflow during summer months?

LITERATURE REVIEW

Physical Characteristic of Riparian Areas

A simple definition of a riparian area is the zone along rivers, streams, lakes and around springs and bogs, wet meadows and ponds (USDI-Bureau of Land Management, 1987). Mitsch and Gosselink (1986) offer three characteristics of riparian areas which separates them from upslope areas:

(1) Riparian areas have a linear form as a consequence of their proximity to streams and rivers.

(2) Energy and material from the surrounding landscape converges and passes through riparian areas.

(3) Riparian areas are connected to upstream and downstream ecosystems.

Hence, the physical and functional characteristics of riparian areas make them unique and dynamic areas on the landscape.

Riparian areas are considered to be areas of plant richness. According to Gregory et al. (in press) riparian areas have twice the number of plant species than upslope areas. In the past, research on riparian vegetation has been minimal since riparian areas are smaller in areal extent and sometimes have tree species with lower economic value than the upslope (Swanson et al., 1982).

Riparian vegetation can be both generalists, that is, those species which can occupy both upslope and riparian areas, and specialists, which are those plants limited to moist streamside areas. Geomorphic characteristics, such as gradient, topography, and soils, in conjunction with surface/subsurface water determine the complexity and richness of riparian plant patterns. Other factors such as aspect and climate also play an important role in determining riparian plant patterns.

Riparian soils vary depending on parent material but are generally characterized as being alluvially and/or colluvially derived. In headwater forested riparian areas there is often periodic disturbance such as flooding and debris torrents which limits soil pedogenesis. This situation results in many riparian soils being classified as entisols (lack of diagnostic horizons indicative of advanced stages of soil genesis). In addition, the presence of a high water table can create an anaerobic environment that may restrict chemical and biological weathering processes in the soil which can also inhibit soil genesis (Platts et al. 1987).

In moist riparian areas, the soils may be classified as "hydric". A hydric soil is one that in an undrained condition is saturated, flooded or ponded long enough during the growing season to develop anaerobic conditions favoring the growth and regeneration of hydrophytic plants

(USDA-Soil Conservation Service, 1985). In these moist areas the presence or absence of gaseous oxygen can effect the rate of oxidation-reduction reactions (Platts et al. 1987) which in turn affects the soil color. The presence of mottles and/or gleying are indicative of the redox conditions in the soil. Mottles that are bright red are indicative of alternating redox processes while gleyed (grey, blue or greenish-blue) soils indicate a prolonged reducing environment where the ferric iron (brown) has been reduced to ferrous iron (grey).

Riparian Subsurface Hydrology

The hydrology of riparian areas has not been widely researched. Most studies of subsurface hydrology have concentrated on hillslope/stream hydrologic responses in relation to stormflow. The hydrologic/hydraulic characteristics of stream-adjacent areas may well be critical for not only stormflow response but also in determining bank storage and stability and riverine ecosystem characteristics such as vegetation patterns, nutrient cycling, and invertebrate and vertebrate community composition.

Working in Brazil, Nortcliff and Thornes (1984) observed:

> "Hillslope and channel coupling for both water and sediment usually takes place across floodplains, and to this extent the control of particulate matter in the channels, if not soil erosion on the hillsides, is largely operational through the floodplain...contrary to widely held beliefs among conservationists and

> forest managers, the runoff dynamics may also very largely be related to the activity of the floodplain rather than the hillslope."

Nortcliff and Thornes further suggest that floodplains (riparian areas in general) on small watersheds have complex interactions between several sources of incoming and outgoing water including subsurface flow, from the hillslope, groundwater, streamflow, and precipitation. Furthermore, they found that those areas along the stream and the toeslope were the most hydrologically dynamic while the piezometers in the center of the floodplain lagged behind the hydrograph.

Groenvold and Grienpentrog (1985) observed that groundwater levels in an alluvial floodplain on the Carmel River in California were critical in determining vegetation patterns. Heavy groundwater pumping in the area not only caused a decline in riparian vegetation but was also linked to severe streambank erosion following the mortality of streambank vegetation.

Stanford and Ward (1988) found that the floodplains in Montana's Flathead area were hydraulically connected to the channel. It had been assumed that the hyporheic zone, i.e., the interstitial habitat penetrated by riverine animals, was only a few centimeters to a few meters wide. However, they found riverine invertebrates up to 2 km from the channel in unscreened shallow wells (10 m) indicating a much broader hydrological/ecological connection between the floodplain and the stream.

Kondolf et al. (1987) found that streambank storage is important for baseflow for alluvial rivers such as the Carmel River in California. However, they suggest there are three conditions necessary for significant bank storage: (1) The stream reach must be subject to stage increase from the passage of high flows, (2) the bank material must have a high hydraulic conductivity (sands and gravels), and (3) there must be sufficient volume of permeable bank material to provide significant storage relative to streamflow. The authors suggest that condition (1) implies that downstream reaches are more favorable for bank storage than headwater streams because their greater drainage area can produce larger flood peaks and thus cause a greater change in storage.

In the Carmel River study area the water held in storage usually drained soon after the passage of the flood peak. However, as long as there was a hydraulic gradient between the bank and the stream, drainage continued to occur up to 6 weeks after the passage of a flood peak. This streambank drainage usually occurred during May and June, which are critical months for the downstream migration of steelhead trout smolts.

It is difficult to draw a strict line between the physical characteristics and functional dynamics of riparian areas since both interact to create and maintain the unique properties associated with these areas. In the preceding paragraphs general descriptions of riparian areas were identified. In the following sections, the functional aspects of riparian areas will be discussed.

Riparian Area Functions

Due to their proximity to the stream, riparian areas act as sources of large woody debris input to the stream. Large woody debris was once considered a blockage to fish and river transport and as a damaging agent to bridges and culverts. As a result, woody debris was systematically removed and salvaged from rivers and streams. However, in the last 10-15 years the importance of woody debris for fish habitat, sediment storage, and bank stability has been increasingly recognized. This recognition has led to leaving large wood in the stream and also increasing the amount of wood in streams through fish habitat structures.

The quality and quantity of woody debris recruitment from a riparian area depends on the tree species and age classes, soil stability, valley form, climate, lateral channel mobility and stream management history (Bisson et al., 1987). The quality and quantity of wood reaching the stream will influence its morphological and biological functioning within the stream. Beschta (1989) cites several morphological and biological influences of large wood in stream: (1) creating and maintaining pools, (2) causing local reductions in stream velocities that serve as foraging sites for fish feeding on drifting food items, (3) forming eddies where food organisms are concentrated, (4) supplying protection against predators, (5) providing shelter during high flows and (6) trapping and storing organic inputs from the streamside forests enabling them to be biologically processed.

Large woody debris, along with other external physical factors such as hillslope erosion and bedrock control, serves to shape channel morphology in mountain streams by acting as a roughness element. Generally, large wood serves to store sediment, create pools and backwater habitat, and also establish stepped gradients which dissipate stream energy resulting in less erosion (Beschta, 1989).

Riparian vegetation plays an important role in protecting stream banks from erosion. Meehan (1977) found that riparian herbaceous communities are effective in reducing the transport of sediment to streams from nearby anthropogenic and natural sources.

During overbank flooding, the above-ground portion of the vegetation acts as roughness elements which can exert considerable influence on the resistance to flow (Hickin, 1984). In small headwater streams, where the stream and the banks are highly interactive, varying vegetation patterns can produce an order of magnitude variation in Manning's n, which in turn influences velocity distribution over a wide range of flows.

The roots of riparian vegetation serve to bind stream adjacent sediments, which in turn influences the lateral stability of the channel. Hickin (1984) found that if discharge, water-surface slope, bend curvature, size of bank materials and bank height are all held constant, a river migrating through unvegetated land will erode twice as quickly as a river going through a forested floodplain.

Riparian areas also serve as a source of temperature amelioration for the stream. Riparian vegetation species, age, crown characteristics, distance from the stream and canopy cover influence the amount and quality of light reaching the stream. The amount of light reaching the stream, along with other factors, affects the level of photosynthesis, hence primary production in the stream.

Riparian areas are beginning to be researched for their role in nutrient cycling. Green and Kauffman (1989) found that riparian areas represent unique sites for chemical transformation due to the presence of an anaerobic soils throughout the year. The saturated anaerobic zone, which is common in low-lying riparian areas, acts as a site of denitrification due to the presence of denitrifying bacteria. These bacteria are a significant means of nitrate removal by reducing nitrate to nitrous gas (N_2O) and nitrogen gas (N_2).

Preventing abnormally high levels of nitrate in streams is important since high levels could lead to a reduction in water quality due to algal blooms, turbidity, and ultimately oxygen depletion (Green and Kauffman, 1989). Rhodes et al. (1985) found that forested riparian areas removed 99% of precipitation nitrogen inputs. However, they point out that riparian areas can also act as rapid conduits of nutrients during rainfall and snowmelt events. Similar results were also encountered in agricultural lands by Lowrance (1983). He found that the riparian area acted as a buffer or sink for the high nutrient load from the fertilized upslope lands that could have potentially reached the stream. Several studies have indicated that riparian areas "clean up" nutrient containing waters with a high degree of efficiency before they enter the stream.

As riparian areas are further understood in terms of their dynamic functioning on the landscape, it becomes important to understand the physical, chemical, and biological bases which determine the direction and magnitude of the riparian functions. A better understanding of riparian hydrological characteristics will hopefully provide an important link in connecting physical processes to functional relationships.

Water Movement Through Soil

For this study only saturated flow will be considered. This limitation is due largely to the difficultly in characterizing and quantifying unsaturated flow within both the vadose and tension-saturated zone.

In a complicated porous media such as soil, the flow pattern can only be described at a macroscopic level since at a microscopic level the velocity varies greatly from point to point. Thus, flow through soil can be treated in terms of a macroscopic flow velocity vector whereby the microscopic flow velocity vectors are averaged over the total volume of soil. The volume of soil is then treated as a uniform media, with flow spread out over the entire cross-section, solid and pore space alike (Hillel, 1982).

The flow through a porous media such as soil is given by Darcy's Law, named after Henri Darcy, a French hydraulic engineer for the city of Dijon, France who developed it in the mid-1800's:

$$v = - K \frac{\Delta H}{\Delta L} \qquad (1)$$

where v is the specific discharge or Darcy velocity, ΔH is the change in head from h_2 to h_1, ΔL is the distance from h_2 to h_1, and k is the proportionality constant generally designated as the saturated hydraulic conductivity.
Another form of Darcy's law is given as:

$$Q = - K \frac{dh}{dl} A \qquad (2)$$

where Q is discharge, and A is cross-sectional area. These equations are critical in groundwater movement analysis and apply to groundwater flow in any direction in space.

The application of Darcy's law to a soil matrix assumes the flow to be laminar. In true soil matrix flow, this assumption is met due to the narrowness of the pores. However, within macropores turbulent flow may occur. For example, Kirkby (1988) indicated:

> "Movement of soil water can, in many cases, be described in Darcian flow, with only modest dispersion of a sharp front between wet and dry soil between identifiable slugs of water. Darcy's law breaks down where there are continuous, connected large voids which allow significant bypassing of the main flow which may be turbulent."

Sloan and Moore (1984) also indicated that direct application of Darcy's law to stormflow in steeply sloping forested environments may not be realistic. Megahan and Clayton (1983) also questioned whether Darcy's law can be applied to steeply sloping forested environments. However, as a result of their studies in steep forested areas in Idaho, they concluded Darcy's law is applicable to those areas.

When flow is unsteady or a soil is non-uniform, hydraulic head may not decrease linearly along the direction of flow. Where the hydraulic gradient or hydraulic conductivity is unsteady or varying, localized gradient, flux, and K values must be used rather than values for the whole system. For these situations Slichter (1899), as indicated by Hillel (1982), generalized Darcy's Law for saturated porous media into a three-dimensional, differential form:

$$q = - K \nabla H \tag{3}$$

This form of Darcy's Law indicates that the flow of a liquid (q) through a porous medium is in the direction of, and at a rate proportional to, the hydraulic gradient (which is the driving force) and also to the hydraulic conductivity which is ability of a medium to transmit liquid (Hillel, 1982).

Hydraulic conductivity (K) is an important component of Darcy's Law and is generally described as the ratio of flux to hydraulic gradient, or the slope of the flux versus gradient curve. K has the same units as flux which is L/T and varies with soil chemical, biological and physical properties. Hydraulic conductivity is a function of not only of the porous medium properties but also the properties of the fluid. Thus, any factor that affects the pore geometry of the soil and/or the fluid properties of the liquid moving through the soil, affects hydraulic conductivity.

In general, the flow of water through soil or a geologic formation occurs in response to a potential gradient (from higher to lower potential) that is established in the subsurface environment (Freeze and Cherry, 1979). This potential is described by the Bernoulli equation:

$$H = \frac{v^2}{2g} + \frac{P}{\gamma} + z \qquad (4)$$

where z is the height above datum, g is the acceleration of gravity, P is pressure, and γ is the specific weight of the fluid. In groundwater systems, velocity is so slow that the velocity term is dropped and gage pressures are used, leaving the hydraulic head equal to the height of the water (P/γ) plus the height above a datum.

Models of Stormflow Generation

In most watershed models, stormflow is generated by one of the following categories: (1) partial area-overland flow, (2) variable source area-overland flow, (3) variable source area-subsurface flow, (4) subsurface flow-direct hillslope-channel coupling, and (5) the groundwater ridge-capillary fringe theory.

The partial area-overland flow concept suggests that runoff is produced mainly from specific areas within the watershed. The partial area concept was developed after estimates of runoff, calculated from rainfall minus evaporation and infiltration, produced linear errors.

These linear errors were explained by assuming that only rainfall on a small, fixed portion of the watershed contributed to runoff during a hydrograph peak (Betson, 1964). The location of these small, fixed areas is determined mainly by soil properties. When these areas become saturated from infiltrating precipitation and surface storage requirements are met, the excess water runs off rapidly as Hortonian overland flow to the stream (Freeze, 1974).

The variable source area-overland flow concept suggests that runoff is generated from watershed areas generally near the stream that become saturated from below by a rising water table. The location of these source areas depends mainly on topography, geology and soil type and their relative sizes may expand and contract in response to climatic factors. Similar to the partial area concept, the runoff from the stream-adjacent variable source areas enters the stream rapidly as overland flow (Sklash and Farvolden, 1979).

A study by Dunne and Black (1970) of a small watershed in New England, with a dissected terrace and a narrow marshy valley floor, suggests that overland flow from saturated areas close to the stream is the dominant runoff producing mechanism. Though they refer to their theory as in terms of the partial area concept, they also suggest the size of these saturated areas varies seasonally. Since they refer to the saturated areas as spatially and temporally dynamic, I chose to include them in the variable source area concept rather than the fixed, partial area concept category.

Dunne and Black (1970) found that subsurface storm flow contributed only minor amounts to the storm runoff hydrograph. The reason for this, they suggest, is the dampening effect of the storage and transmission of water within the soil. The role of the hillslope in generating storm runoff depends on its ability to generate overland flow in small portions of the watershed. Thus, the major portion of storm runoff is produced from saturated areas near the stream. The rest of the watershed acts as a reservoir during storms; between storms it provides water for baseflow and maintains wet areas that produce stormflow. These wet areas are produced by both water escaping from the ground surface to reach the channel as overland flow and by direct precipitation on saturated areas which act essentially as expanded stream channels.

Freeze (1972), in concurrence with Dunne and Black, found that:

> "On convex slopes with lower permeabilities and on all concave slopes direct runoff through very short overland flowpaths from precipitation on transient near-channel wetlands dominates the hydrograph. In these expanding wetlands, surface saturation occurs from below because of vertical infiltration towards a very shallow water table rather than by downslope subsurface feeding."

Freeze further suggests that subsurface flow cannot deliver sufficient water to have a significant contribution to stormflow. He states, "Theoretical simulations of runoff generation in upstream source areas show that there are stringent limitations on the occurrence of subsurface stormflow as a quantitatively significant runoff generating mechanism." These stringent limitations include a threshold hydraulic conductivity of 0.002 m/s (below which subsurface stormflow is not feasible mechanism) and then only where convex hillslopes are directly connected to steeply incised channels.

The variable source area-subsurface flow concept, like the variable source area-overland flow concept, suggests that source areas of storm runoff contribution expand and contract in response to climatic factors. However, the variable source area-subsurface flow concept suggests that water is transferred from the hillslope to the stream through subsurface routes, rather than through overland flow pathways adjacent to the stream. The combination of an expanding channel network and translatory flow (rapid displacement of stored water by new rain) causes the storm runoff to reach the stream quickly (Sklash and Farvolden, 1979).

Work by Hewlett and Hibbert (1967) relates the quick rise of streamflow to variable source areas and subsurface translatory flow. They suggest that translatory flow occurs mainly on the lower portions of the watershed where it is important in direct flow. A paper by Hewlett and Nutter (1970) states:

> "An expanding channel network...[wherein] the channel reaches out to tap the subsurface stormflow systems which, for whatever reason, have overridden their capacity to transmit water beneath the surface ...The rapidly expanding channel allows subsurface flow, even at a velocity of a few feet per day, to reach the channel in time to contribute to and sustain the upland storm hydrograph...[The] expansion is aided by rain falling directly on the wetted areas."

Harr (1977) in the Western Cascade mountain range of Oregon found that 97% of the stormflow was of subsurface origin with the remaining 3% from channel interception. Harr found that subsurface stormflow is a major contributor to storm runoff in his study watershed and was in agreement with the variable source area concept of runoff production. He found that there were saturated zones at the toe of the study slope that were visible and appeared to expand upslope and laterally with continuing rainfall, yet never produced overland flow.

The subsurface stormflow concept, unlike the variable source area-subsurface flow concept, does not include an expanding and contracting source area, but instead proposes that water moves laterally from the hillslope to the stream mainly through a series (not all necessarily connected) of large soil pores, old root channels, animal burrows, cracks, etc., termed macropores or macrochannels.

To make the distinction between subsurface flow and groundwater flow, subsurface flow will be defined as "that part of the precipitation which infiltrates the surface soil and moves laterally through the upper soil horizons towards the stream as ephemeral, shallow, perched groundwater above the main groundwater level." (Chow, 1964).

Whipkey (1965) gives four criteria for the occurrence of subsurface stormflow: (1) the soil surface is permeable, (2) the land is sloping, (3) there is a water-impeding layer near the surface, and (4) the soil is saturated. He further suggests that in undisturbed forest watersheds all the criteria for subsurface stormflow are met and that the subsurface flow is the dominant mechanism of storm runoff.

A study by Beasley (1976) suggests that subsurface flow from the upper portions of the watershed is a significant proportion of the total flow from the watershed. This is counter to both the partial and variable source area concepts which place the greatest emphasis on the role of small, near channel areas for the production of flow. Beasley suggests that macrochannels are important in conveying stormflow and that in forested environments the condition for macrochannel flow is met (i.e. the pores are open to the atmosphere and have positive heads at the openings). He also reasons that the flow must be through macrochannels since soil matrix flow (even with high saturated hydraulic conductivity values) is not sufficiently fast enough to account for the timing (sometimes less than 1 hour from the onset of precipitation) of the peaks.

Whipkey and Kirkby (1978) also found that macrochannel flow was a viable mechanism for subsurface stormflow movement. They reasoned that macropores are enlarged and maintained by hydraulic erosion which may increase the size of the pores and cracks into significant pipes able to transport large amounts of water.

Mosley (1979), working in New Zealand, found that the velocity of the flow through macropores within the soil were up to 3 times greater than that of the soil matrix. He concluded that subsurface flow through the macropores, and not the soil matrix, is the dominant mechanism for both stormflow and baseflow in the study area.

Sloan and Moore (1984) agree that macropore flow is the dominant mechanism for stormflow generation, however, they suggest that soil matrix flow is the dominant mechanism for baseflow generation. They state:

> "Based on preliminary field observations and a preliminary analysis of the field data by the authors...it is postulated that there are two subsurface flow components contributing to the hydrologic response of steeply sloping forested watersheds: (1) macropore flow-which is responsible for stormflow response, and (2) soil matrix flow-which is responsible for baseflow or delayed response."

The final watershed response concept to be discussed is the groundwater ridge-capillary fringe theory which is used in support of the variable source area concept (both surface and subsurface). Sklash and Farvolden (1979) found that subsurface flow was a major contributor in runoff producing events for relatively flat watersheds in Ontario, Canada. The problem they had with asserting that subsurface flow is an important component of stormflow was to determine how the slow moving subsurface water could appear so quickly in the stream during a storm event. Their theory for explaining the significant role of subsurface water in generating streamflow is as follows:

> "Along the perimeter of transient and perennial discharge areas, the water table and its associated capillary fringe lie very close to the ground surface. Soon after rain or snow-melt begins, infiltrating water readily converts the near-surface tension-saturated capillary fringe into a pressure-saturated zone or groundwater ridge."

This groundwater (subsurface water) ridge both provides an increased impetus for the flow of groundwater to the stream from the discharge zone, but also expands the size of the subsurface water contribution area (discharge zone). The response of the hillslope subsurface flow may become important at later times in the storm event, but has little influence in the early part of the storm event.

Gillham (1984) supports the groundwater ridge-capillary fringe theory with his work on the capillary fringe and its effect on water table dynamics. Like Sklash and Farvolden (1979), he found that subsurface flow constitutes a larger percentage of the storm runoff than would be expected using graphical methods of hydrograph separation. He also observes that it is not clear how the slow moving subsurface water could respond so rapidly to precipitation events.

The presence of a capillary fringe zone (the zone above the water table that remains saturated under negative pressure) could explain the rapid subsurface response during precipitation events (Abdul and Gillham, 1984). In these zones where the capillary fringe can extend to or near to the ground surface (especially in fine-textured soils), there is little or no available storage capacity. A small amount of precipitation in these areas can result in a rapid rise in the water table thus increasing the hydraulic gradient towards the stream. This rapidly rising water table creates a groundwater ridge which then discharges into the stream from stream-adjacent overland flow or directly to the stream channel through the stream bed.

This variety of stormflow generating concepts does not necessarily mean that the concepts are contradictory, or that it is necessary to search for one absolute mechanism, but rather suggests that over varying spatial and temporal conditions different mechanisms may predominate.

DEER CREEK WATERSHED

This study site is located in the Deer Creek Watershed which is approximately 16 km south of Toledo, Oregon, in Lincoln County (Figure 1). Deer Creek is a tributary of Horse Creek, which in turn flows into Drift Creek, ending ultimately in the Alsea Bay at Waldport, Oregon. The Deer Creek Watershed has an area of 3.03 square kilometers and is administered primarily by the U.S. Forest Service, with a small portion in private ownership. This study site was selected due to the minimum management disturbances, ease of access, and was part of the Alsea Watershed longterm study which provided a source of historical data for the watershed.

The Deer Creek Watershed is in the Oregon Coast Range physiographic region which extends from the middle fork of the Coquile River in the south to the Columbia River in the north, a distance of 350 km. The Coast Range physiographic region varies from 50 to 100 km in width. Average summit elevations range from 500 m to a maximum of 1250 m at the top of Mary's Peak near Corvallis, Oregon.

The geology of the Deer Creek Watershed is typical of the central Coast Range, with Tyee Sandstone representing the principal formation. The Tyee formation is bluish-grey to grey, rhythmically bedded, micaceous and arkosic

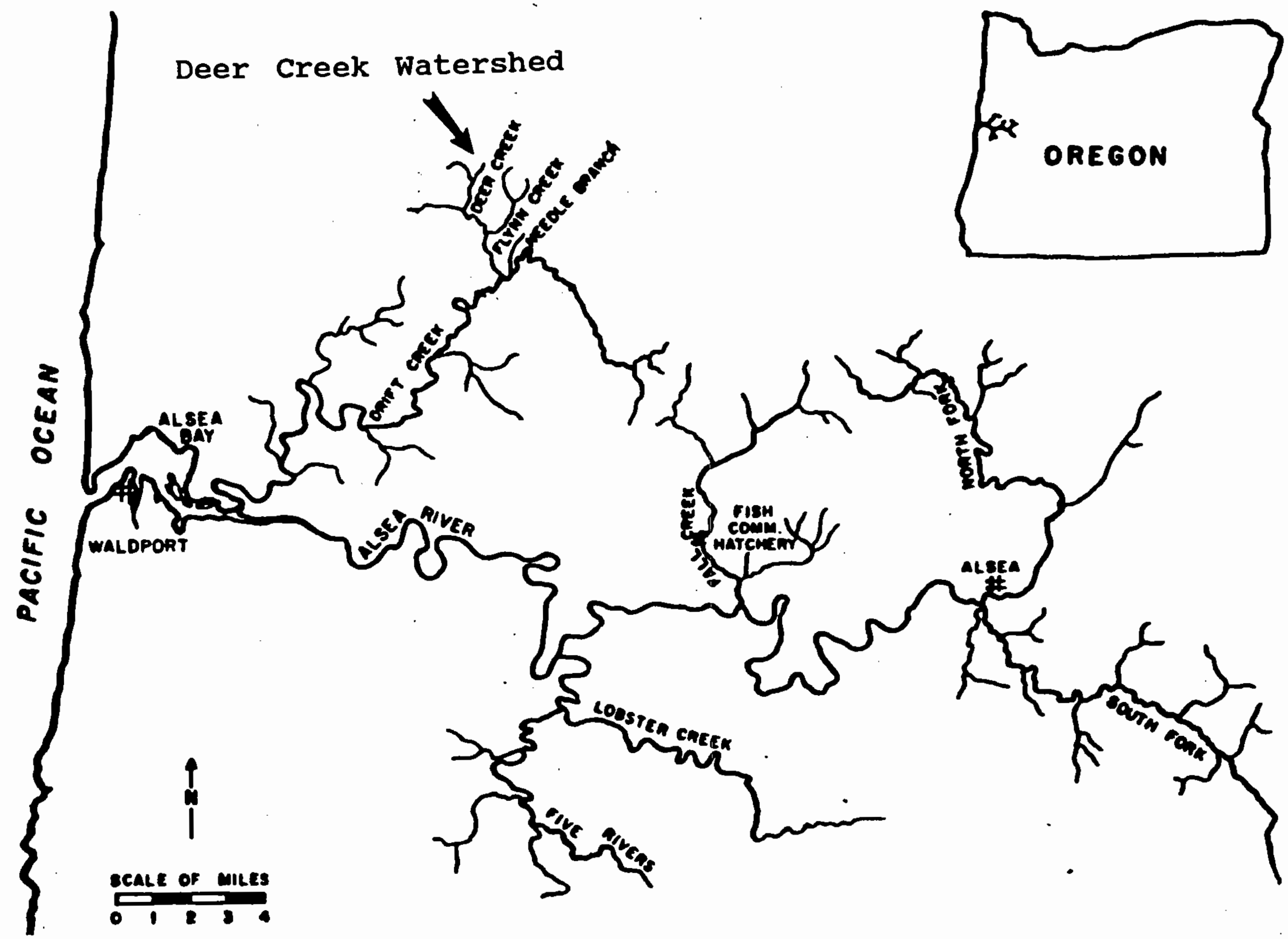

Figure 1. Alsea River System and location of Deer Creek Watershed. Adapted from Moring and Lantz (1975).

(granular sedimentary rock composed of clear quartz and feldspar or mica) sandstone, approximately 3 m thick, which grades upward into the overlying fine-grained carbonaceous siltstone (Clay Mineral Society, 1976). The sandstone is lithified and characterized by an abundance of mica flakes. Plant fragments have been mostly found in the siltstone layers. The lack of marine invertebrates suggests that the formation was deposited rapidly (Baldwin, 1964).

Soils of the Deer Creek Watershed are mainly in the Slickrock and Bohannon series, with areas in the Bohannon-Slickrock association. These soils are mainly gravelly loam soils 0.5 to 1 m deep over arkosic sandstone and gravelly clay loam soils more the 1.2 m deep to tuffaceous sandstone. The general characteristics of this series include well-drained, somewhat gravelly, moderately fine-textured, very acid soils formed in colluvium on 3-50% mountain sideslopes and footslopes. Depth to underlying sandstone varies from 1 m to several meters. These soils commonly have 6-8% organic matter in the surface horizon with more than 1% to at depths of 0.5 meters or greater.

The climate of the Deer Creek area is characterized by wet winters and fairly dry and mild summers. The area receives 2000-3000 mm of precipitation annually. During the winter months there is considerable cloud cover and frequent rains as moist air moving in from the Pacific Ocean rises and cools. Ninety percent of the annual precipitation from October to May falls as rain, with snow making up only a small part of the precipitation. Precipitation intensities tend to be low (USDA-Soil Conservation Service, 1973).

The vegetation of the area is mainly dense stands of conifers and hardwoods with a shrub, grass, and herb understory. The dominant conifer species of the area is Douglas-fir (*Psuedotsuga menziesii*) with considerable amounts of western hemlock (*Tsuga heterophylla*), and western redcedar (*Thuja plicata* Donn). The dominant hardwood species are red alder (*Alnus rubra* Bong.), and bigleaf maple (*Acer macrophyllum*). The dominant understory vegetation consists of swordfern (*Polystichum munitum*), oxalis (*Oxalis oregona*), vine maple (*Acer circinatum*), and salmonberry (*Rubes spectabilis*).

Study Site

The Deer Creek study site is located along an unconstrained reach of Deer Creek (Figure 2). The site was selected since it appeared to be a homogeneous surface. The geomorphic surface (Figure 3) at the site consists primarily of a terrace formed from old alluvial deposits. In contrast to true floodplains, which are active surfaces that can be reworked every 1-2 years,

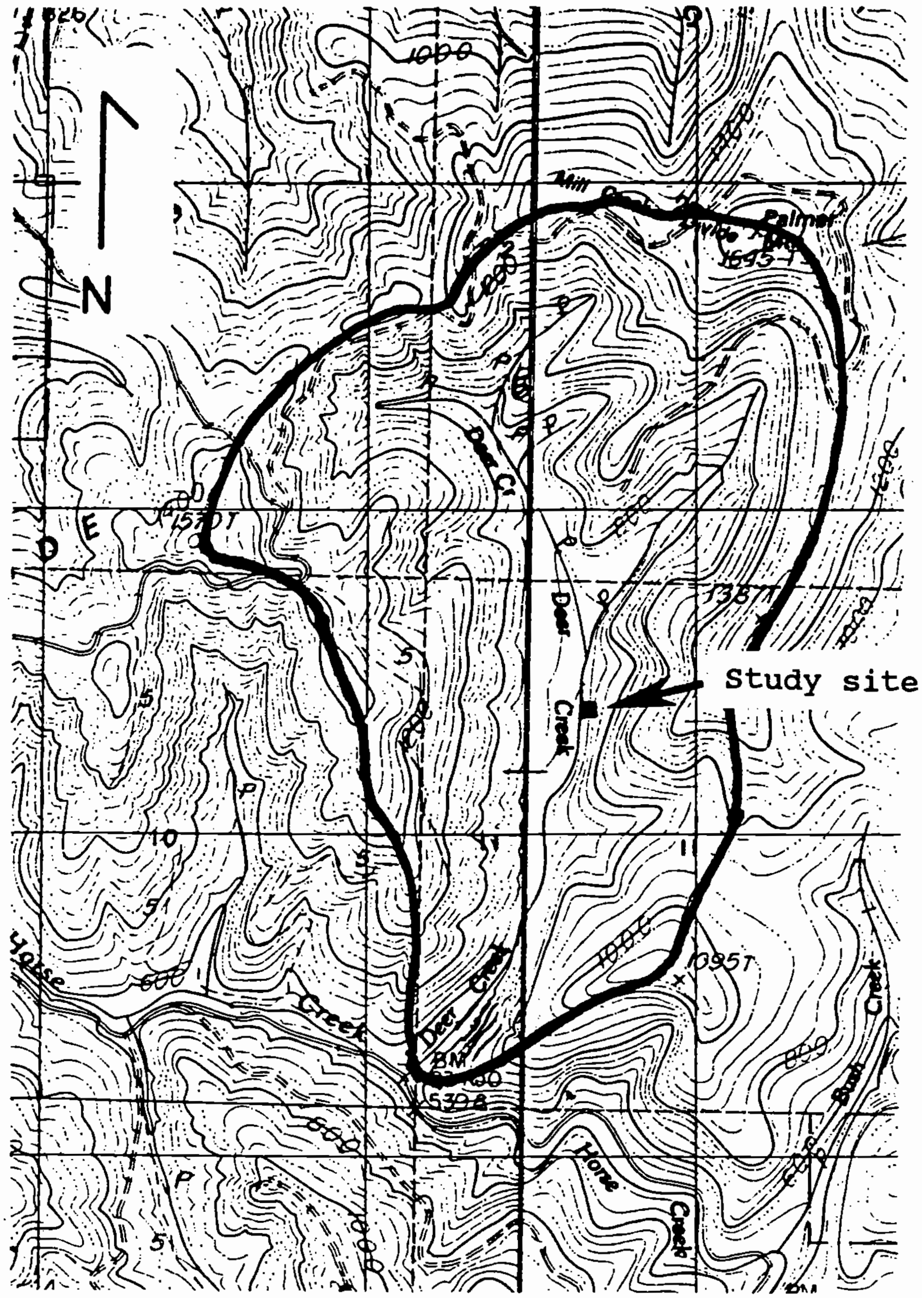

Figure 2. Topographic map of the Deer Creek Watershed showing location of study site. U.S.G.S. Elk City and Toledo South quandrangles (approximate scale of 1:24,000). T.12 S. R.10 W. Section 11.

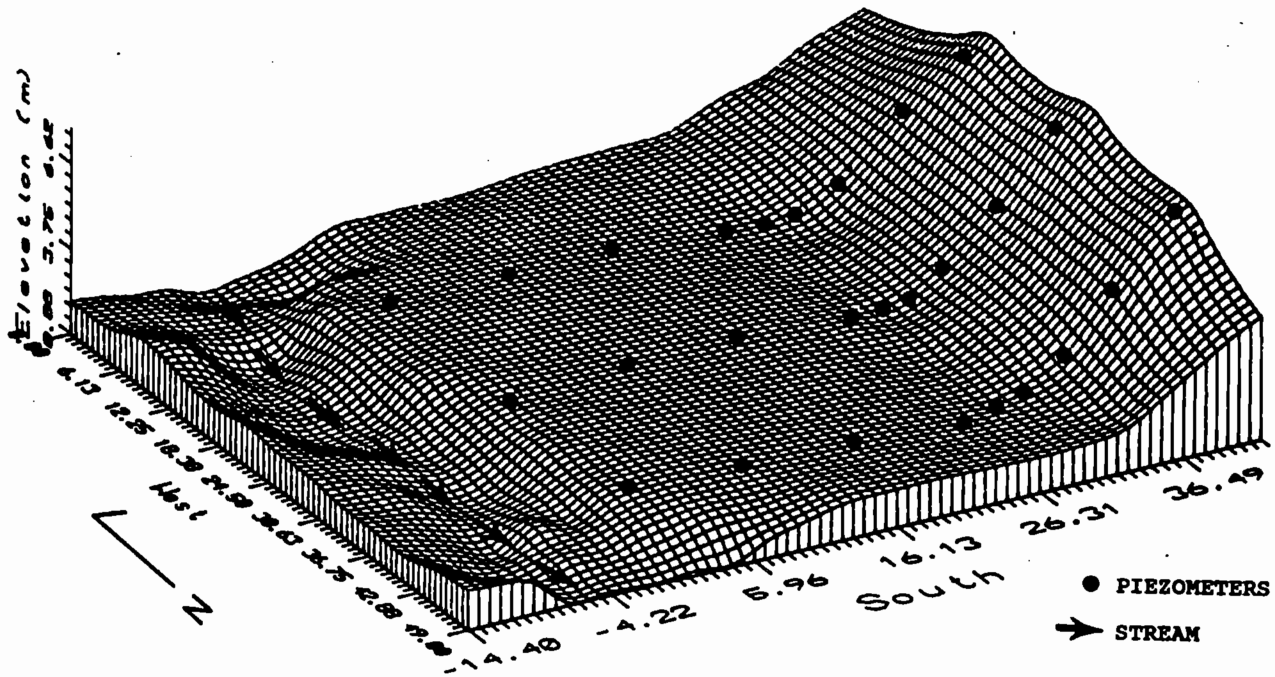

Figure 3. Deer Creek study site surface and piezometer locations.

terraces seldom experience flooding. The 25-50 year old red alder stand on the terrace indicates it has been at least that long since the last flood event. The soils of the terrace tend to be finer textured loams to silty clay loams. Detailed descriptions of soil cores from the study site are found in Appendix A.

The overstory vegetation of the terrace is dominated by red alder (Figure 4); Douglas-fir is common on the adjacent hillslopes. The understory vegetation is mainly swordfern, salmonberry, oxalis, grasses, and some elderberry. Appendix B contains detailed vegetation information for the site.

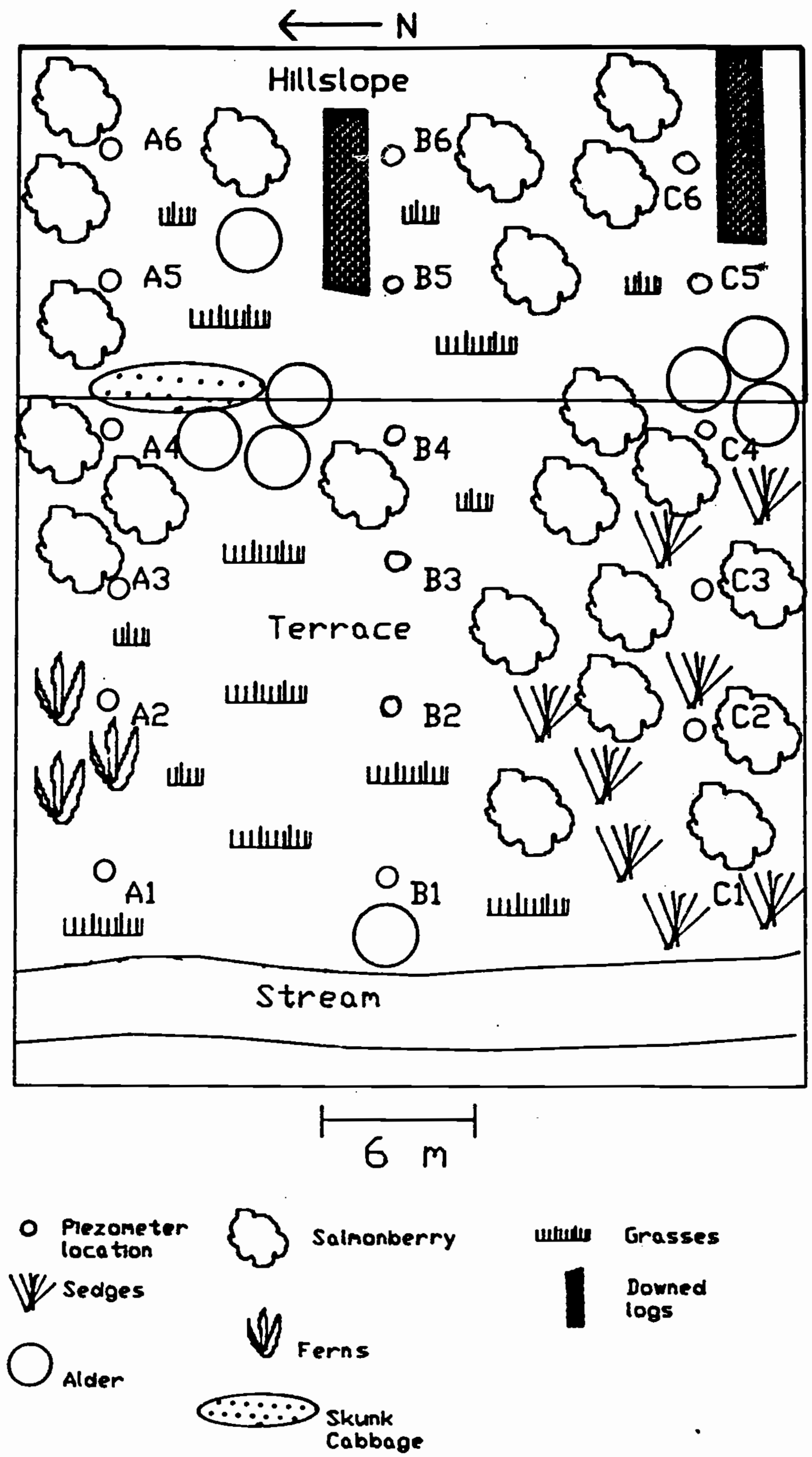

Figure 4. Generalized vegetation map of the Deer Creek study site.

METHODS

In late September of 1989, 27 piezometers were installed in a grid pattern extending from the stream, across the terrace, and onto the adjacent hillslope in the Deer Creek Watershed. Nine piezometers were laid out in each of 3 transects oriented perpendicular to the stream (Figure 5); transects were placed 12.2 meters apart. Within the transects, the terrace piezometers (numbers 1-4) were installed approximately 8.5 meters apart, the two shallow toeslope depressional area piezometers (4a and 4b) were 2.1 m apart, and the hillslope piezometers (numbers 5-7) were approximately 4.6 meters apart (Figure 5). Piezometers 1-3 were 1.8 m in depth, 4 and 5 were 1.5 m, 4a and 4 b were 0.61 m, and 6 and 7 were 2.7 m.

The type of piezometer chosen for the site was an open standpipe type. A drawback to this type of piezometer is that the measured head may be higher or lower than the free water table and, in moderately impervious soils there may be a large time lag. Thus, with open standpipes in partially saturated soils there is the problem of evaluation of measured head (Transportation Research Board, 1978). However, due to their simplicity, cost, durability, and ease of installation, standpipe piezometers were selected.

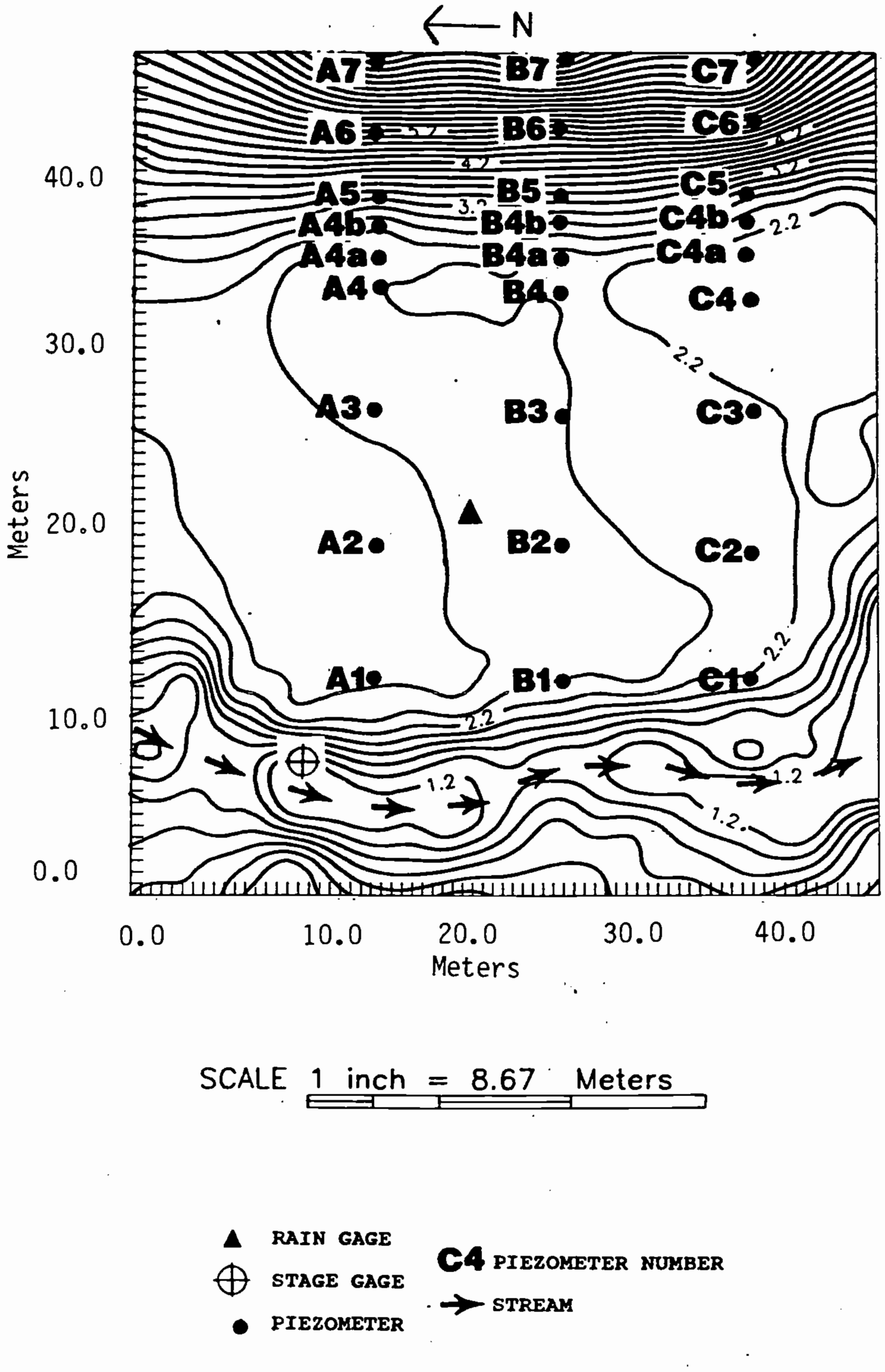

Figure 5. Plan view of the Deer Creek study site.

Single point piezometers were used instead of piezometer nests since the field technique for determining soil texture cannot distinguish orders of magnitude difference in hydraulic conductivity values. In addition, most of the soil horizons were too narrow for piezometer placement.

The piezometers were constructed from 2.54 cm diameter polyvinyl chloride (PVC) pipe with 48 3-mm holes drilled in the lower 21 cm of the pipe. The inside of the piezometer was lined with TYPAR®, a geotextile fabric, to prevent fine sediment from entering the pipe.

A soil probe with a slightly smaller diameter than the piezometer was used to dig the piezometer holes. The soil probe allowed for intact soil cores to be taken from the holes and also caused minimum disturbance to the sidewalls of the holes. The soil core was laid out in a 1/2 section PVC trough and soil properties such as USDA soil textural classes and colors were determined. Soil cores were stored in clear plastic tubes for later reference.

The depth of the piezometer holes was determined by the depth to saturation. Since installation occurred in late summer, it was reasoned that the water table would be at its lowest. After the holes were augered, the piezometers were driven directly into the holes. The soil texture of the terrace is quite fine and so leakage down the sides of the pipes was thought to be minimal. The standpipes were covered with ventilated caps and then painted black to reduce visibility.

Pressure transducers (hydrostatic water depth probes) were placed in the middle transect of the piezometers (i.e., B transect). Six of the nine piezometers were equipped with transducers (Appendix C, Figure C-1). The two shallow piezometers (B4a and B4b) and the uppermost hillslope piezometer (B7) were not equipped with transducers due to the limited number of channels available on the data logger. The transducers were manufactured by UNIDATA® Corporation, Lake Oswego, Oregon. Model 6508c was chosen which measures water height (i.e. hydrostatic pressure) from 0-5 m above the transducer. The transducer uses hydrostatic pressure exerted on a Motorola piezometric electric pressure sensor as the primary data source.

Transducers are calibrated with clean water (specific gravity of 1.0) at room temperature. Suspended sediment and/or varying temperatures may cause an error in the readings. The specifications for accuracy of the unit indicates that there is a 1% of the full scale error over the range of 0-50° celsius. The resolution is 0.4 % of full scale operating at -1 to 40° C. For model 6508c this leads to a maximum error of +/- 5 cm in accuracy and within 2 cm resolution.

Hydrostatic pressure is dependent on several factors including the density of the liquid (D), the acceleration of gravity (g), the depth of the liquid (d) and the atmospheric pressure (atm) at the top of the liquid. Thus, the absolute pressure at a particular depth is equal to (g x D x d) + atm. The vent tube in the transducer causes atmospheric pressure to cancel leaving gage pressure equal to (g x D x d).

The transducer transect was set to record water depth and accumulated precipitation every 15 minutes. In addition, depth to the free water surface in all piezometers was recorded on a weekly basis or biweekly basis. To measure depth to water in the standpipes, a multimeter was attached to two metrically incremented wires that terminated with a weight; the ends of the wires were exposed. Using resistivity, the multimeter needle would jump when the exposed wires contacted the water surface. Depth to water was then determined from the incremented wire. Weekly precipitation totals were also obtained using a standard number ten can. Stage was read from an attached meter stick on the stage gage. Time of measurements were recorded so as to be able to calibrate transducer readings.

A water level recorder was installed just upstream of the furthest upvalley transect (Figure 5). The water level recorder in Deer Creek was constructed using a 2.1 meter section of 15 cm diameter, thick-walled PVC with 1.9 cm holes drilled in the lower portion of the pipe. A 2.5 cm diameter PVC pipe had holes drilled at the lower end and then was fastened inside the larger PVC pipe. A pressure transducer identical to the ones used in the piezometers was placed inside the smaller PVC pipe. The entire gage was fastened to a stream-adjacent alder with large hose clamps.

An electronic tipping bucket precipitation gage (Appendix C, Figure C-2) was placed on the site in a clearing between the first (A) and second (B) transects (Figure 5). The tipping-bucket rain gage was mounted on a platform which was bolted onto a steel pipe that had been driven into the ground. The rain gage collector funnel filters the precipitation before it passes to the tipping bucket. The bucket tips with each 0.2 mm of precipitation and has a range of 0-150 mm per hour.

To prevent damage from animals and the elements, wires leading to the data logger from the piezometers, stage gage and rain gage were installed in 1.3 cm diameter PVC conduit. The conduit was then buried in shallow trenches.

The data logger consisted of a microprocessor that stores recorded data on a CMOS RAM 32K memory (Appendix C figure C-3). Three wires (ground, power, and data) connected to a field termination strip allowed for several incoming channels. A Toshiba® 1000 laptop computer was utilized to retrieve the data from the logger. The system included a menu driven software system that allowed for ease in data retrieval.

Hydraulic Conductivity Estimation

To estimate *in situ* values of saturated hydraulic conductivity (K) of the piezometers a procedure known as a "slug test" was performed. This test was undertaken by Ed Salminen, a graduate student in Forest Engineering, as a class project for the groundwater hydraulics class in Civil Engineering.

A slug test is initiated by first causing an instantaneous change in the water level in the standpipe by the introduction of a known volume of water or material. The method chosen for interpreting the results of the slug test and obtaining K values was the Hvorslev method (1951). The Hvorslev method was determined to be the most appropriate due to its simplicity in interpreting piezometer recovery and that it is applicable for partially penetrating wells (Freeze and Cherry, 1979).

The Hvorslev method assumes homogenous, isotropic, infinite medium in which both soil and water are incompressible. Other assumptions of the Hvorslev method include:

1. Darcy's law is valid.
2. The aquifer is confined.
3. There is no friction in the piezometer.
4. The influence of stress adjustments, air in the formation or piezometer, clogging, etc. are negligible.

The result is:

$$K = Area \;/\; F\,T = \pi\, r_c^2 \;/\; F\,T \qquad (5)$$

Where:

r_c = effective radius of the casing over the depth of change in water in the casing.

T (time lag) = time when $(H\text{-}h/H\text{-}H_0)=0.37$ on a log $(H\text{-}h/H\text{-}H_0)$ versus time graph. H is the recovered head, h is unrecovered head, and H_0 is the initial drop in water level in the well.

F = shape factor for various geometries based on flow net-type solutions and a line source or sink.

For the case in Deer creek the following shape factor was used:

$$F = \frac{2\pi\,(l-d)}{\ln[\,2m\,(l-d)\,/\,r_w]} \qquad (6)$$

Where: (l-d) = length over which water enters or leaves the piezometer = 0.21 m.

m = saturated thickness of the aquifer = 7.3 m (maximum depth of auger).

r_w = well-bore radius (as related to flow entering from or flowing to the aquifer). In this situation since the well bore radius is essentially equal to the piezometer radius, $r_w = r_c = 0.0165$ m.

In situ measurements of hydraulic conductivities in the Deer Creek study site transects included all piezometers that had a measurable water table at the time of the slug test. Piezometers A4a, A4b, A7, B4a, B4b, B6, B7, C4a, C4b, and C7 were excluded because water levels were below the piezometer on the date of measurement, February 22, 1990.

In piezometer A1, a simulated 2.4 m "slug" made from 1.27 cm diameter PVC was introduced into the piezometer and allowed to equilibrate before beginning the test. This trial test revealed that the equilibration period was extremely long and that more than one day would be necessary to complete the slug tests on the other piezometers. Thus, the remaining wells were evaluated by adding a known volume of water to the standpipe. Both methods should yield the same results. Recovery in the standpipes was monitored throughout the day and again one week later. Appendix D lists the time lags, F values, and calculated K values. The calculated values are within the ranges listed in Freeze and Cherry (1979) for clays, sands and silts. To determine the saturated thickness (m), a borehole adjacent to the site was augered by hand until what appeared to be regolith was encountered. K values for varying saturated thicknesses were also calculated in case the 7.3 m depth was not the absolute aquifer thickness. However, there was little difference in the calculated K values using a range of m=7.3-10.0 m and thus a depth of 7.3 m was used.

Equipment Malfunctions

During the study there were three periods of equipment malfunction where the data was lost, either completely or partially.

From the time of installation in early October, 1989 until January 1990, the uppermost transducer, B6, was not functioning properly. Several attempts were made to correct the problem, however, none were successful and so a new transducer was installed.

Data was lost from December 15, 1989 to January 11, 1990 due to a low battery in the data logger. The company was concerned that the reason for a 12-month battery to only last 3 months was a short in the wiring, and so the data logger was sent back to Unidata Corporation for testing. The data logger appeared to be working properly, and so was returned with a new battery.

On April 24, 1990 the rain gage ceased functioning. After rewiring the system the gage was still inoperative and was brought in from the field for testing. Since the gage appeared to be working properly when tested it was reinstalled. However, when it was reconnected, it still would not record precipitation. The rain gage resumed working properly after being connected to a new data logger sent by the company. The gage resumed functioning on May 21, 1990.

Another gap in the data occurred from February 22, 1990 to March 28, 1990. This gap was a result of the "slug test" performed on February 22, 1990. To perform the slug test, water was added to each of the piezometers and then the recovery recorded over the day. At the end of the day there were some piezometers that were not fully recovered. It was assumed that the piezometers would recover by the following week, however, this was not the case. The following week the water level in some of the piezometers was higher than the pre-slug test levels, though no precipitation fell during the intervening week. On March 28, 1990 it appeared the piezometers were all recovered and data collection was resumed.

RESULTS AND DISCUSSION

1990 Water Year

Four storm events were utilized for detailed analysis; these were chosen from the top four high-stage events. Table 1 indicates the dates, maximum stage, and previous 24-hour precipitation for each of these events. The 1990 water year appeared to be a fairly typical precipitation year for the Deer Creek Watershed (Figure 6), although November and March precipitation was below normal and June precipitation was above normal.

Table 1. Ranking of four largest storm events for the 1990 water year, Deer Creek Watershed.

Ranking	Storm period	Maximum stage (m)	24-hour ppt (cm)
1	Dec 4, 1989	1.99	12.67
2	Apr 27, 1990	1.89	*
3	Feb 8, 1990	1.71	7.11
4	Jan 28, 1989	1.64	4.36

* Recording precipitation gage inoperative.

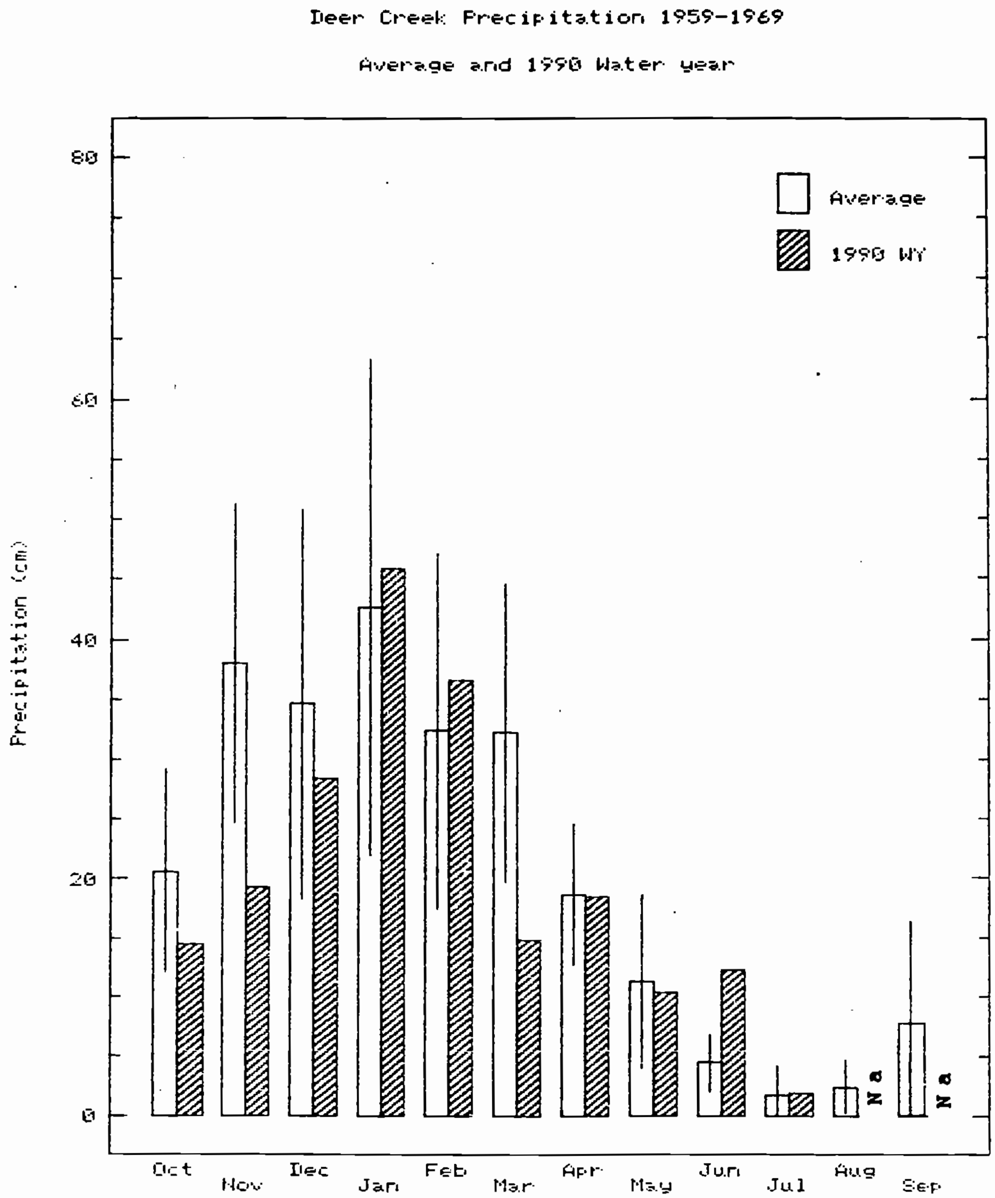

Figure 6. 1990 water year precipitation for the Deer Creek Watershed compared to average for water years 1959-1968. Vertical lines indicate standard error.

Question #1

What is the apparent direction of flow of subsurface water within the hillslope-riparian area-stream continuum throughout the year and during storm events?

Originally, this question was formulated to address the magnitude and direction of subsurface flow, but only direction was included in the final analysis. To include flow magnitude a Darcy velocity, based on saturated K values and hydraulic gradients, would be needed:

$$V_x = - K \frac{\delta h}{\delta x} \tag{7}$$

and

$$V_y = - K \frac{\delta h}{\delta y} \tag{8}$$

Darcy velocities for the Deer Creek study site were calculated to be between 10^{-8} to 10^{-11} m/s. Even if these values were converted to average linear velocities by dividing v by n (porosity, which is always less than 1.0), the soil matrix velocities are still insufficient to generate stormflow (see Question #2). Soil matrix velocities, based on *in situ* estimations of saturated K values and hydraulic gradients for the Deer Creek study area, may greatly underestimate the magnitude of subsurface flow velocity and so was not used. Thus, for this question only apparent flow direction was considered.

The component of flow in the x direction was calculated by using the partial derivative $\delta h/\delta x$, which was then approximated by:

$$x = \frac{((h_2 + h_4)/2) - ((h_1 + h_3)/2)}{\Delta x} \qquad (9)$$

Where h_1, h_2, h_3, and h_4 are the hydraulic heads at each piezometer in a four piezometer grid (Figure 7).

The y component of flow was calculated in a similar manner using an approximation for the partial derivative $\delta h/\delta y$:

$$y = \frac{((h_3 + h_4)/2) - ((h_1 + h_2)/2)}{\Delta y} \qquad (10)$$

The vectors were then plotted using the midpoint of the four piezometers as the origin.

Figure 8 illustrates the direction of water movement from September 26, 1989 to July 25, 1990. Only vectors that differ significantly from the rest are labeled with dates. The hillslope flow vectors show a consistent pattern in a downhill and downvalley direction throughout the year. A similar pattern occurs for those terrace piezometers that are near the hillslope.

The mid-terrace vectors do not show a clear and consistent pattern. For example, vectors between

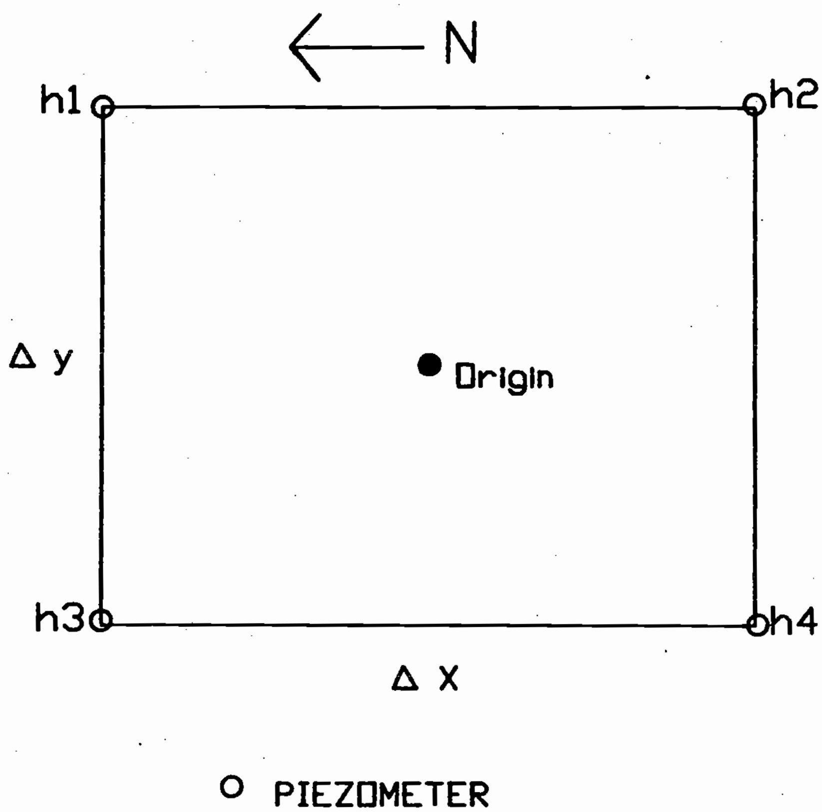

Figure 7. Piezometer grid for flow direction calculation.

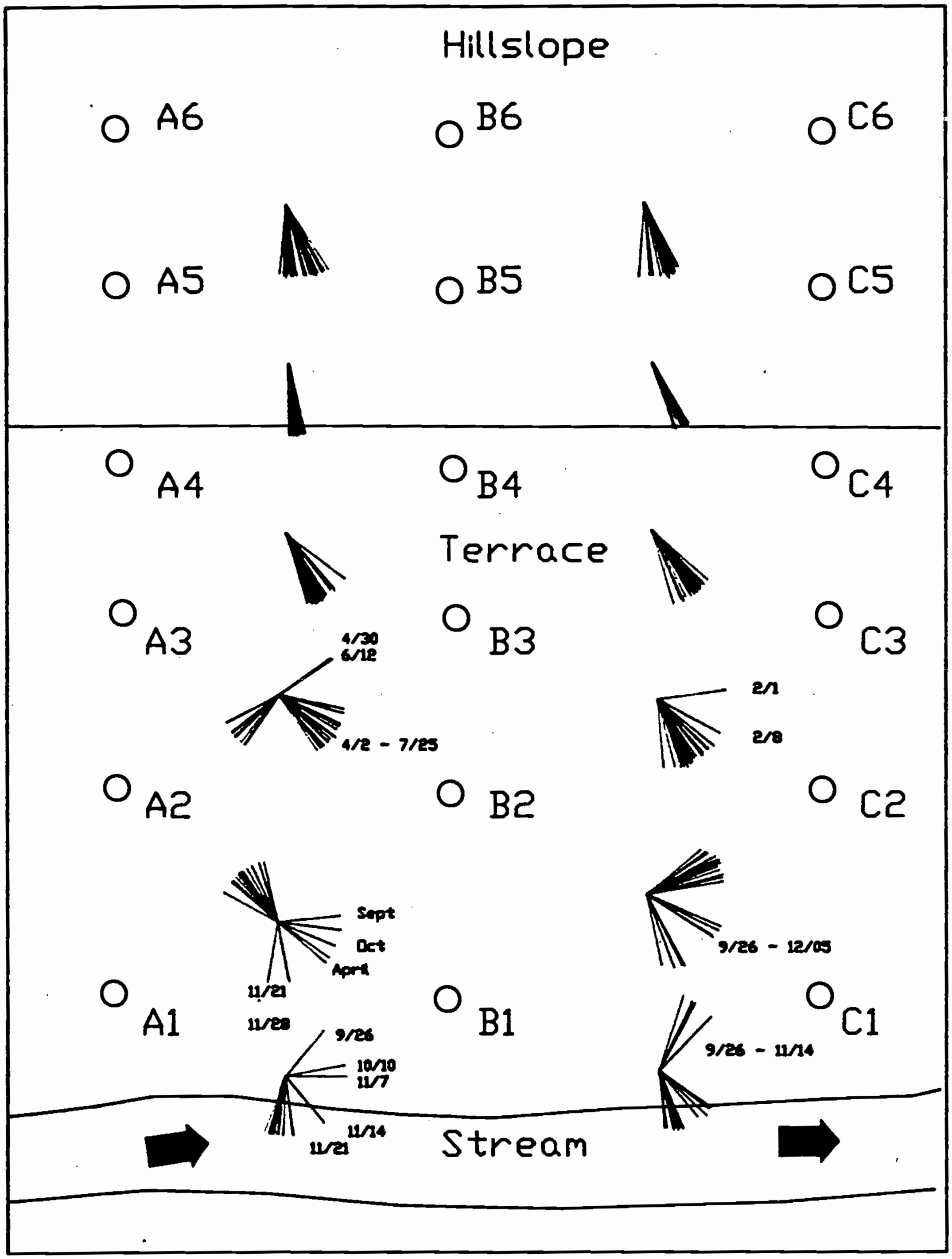

Figure 8. Seasonal direction of flow within the Deer Creek study site, September 26, 1989 to July 25, 1990.

piezometers A1-B1 to A3-B3 show a seasonal reversal of direction indicating water is flowing in an upvalley direction. A possible explanation for this situation is that approximately 10 m upvalley from the A transect is a topographic depression that runs parallel to the A transect. This topographic depression appears to remain saturated at or near the ground surface throughout most of the year. However, if the hydraulic head in A becomes sufficiently high, it may flow towards the depressional area rather than flowing downvalley. The direction of flow along the stream is mainly from the terrace to the stream (Figure 8) except from September to November where flow occurs from the stream to the terrace.

To further examine the direction of flow within the hillslope-riparian area-stream continuum on a seasonal basis, direction of flow within the B transect (i.e., the transect with recording piezometers) was analyzed for fall (October) and summer (July) periods, which were considered relatively dry periods. Direction of flow during the intervening winter storm periods will be discussed in later paragraphs.

Transducer data was analyzed for the periods October 24-28, 1989, and July 3-25, 1990 to determine the direction of flow within the transect. Hydraulic head of the stream (i.e. stage) was subtracted from the stream adjacent piezometer (B1) and then the hydraulic head in B1 was subtracted from the next piezometer in the B transect towards the hillslope (B2). This process was repeated for all B transect piezometers. A positive value indicated flow was towards the stream while a negative value indicated flow was towards the hillslope (i.e. away from the stream).

Figure 9 illustrates the direction of flow for October 24-28, 1989. The direction of flow changes with increasing precipitation (less than 5.5 cm.) from 23:45 on October 25 to 06:45 on October 26, and changes again 45 minutes after precipitation stops. On October 28, 1989 the flow again changes direction. In each situation when flow changed direction, the pattern of flow was not the same as recorded previously.

Figure 10 illustrates the direction of flow for a three week period from July 3-25, 1990. During that period only 0.5 cm of precipitation fell and the direction of flow remained unchanged. Thus, it appears that flow direction changes are dependent on precipitation. In addition, directional changes can occur relatively quickly.

To further evaluate subsurface flow patterns, the direction of water movement during three storm periods in December, 1989, February and April, 1990 were analyzed. The method used for analysis is identical to that described previously for the B transect transducer data.

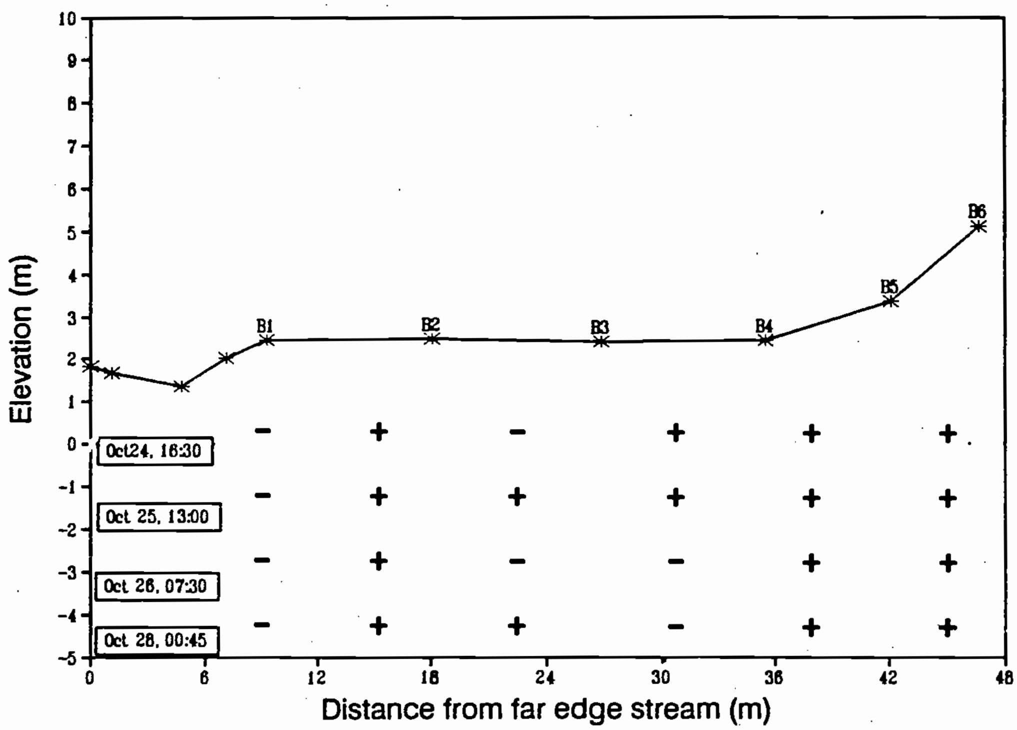

Figure 9. Direction of flow within the B transect from October 24-28, 1989. '+' indicates flow is towards the stream while '-' indicates flow is away from the stream.

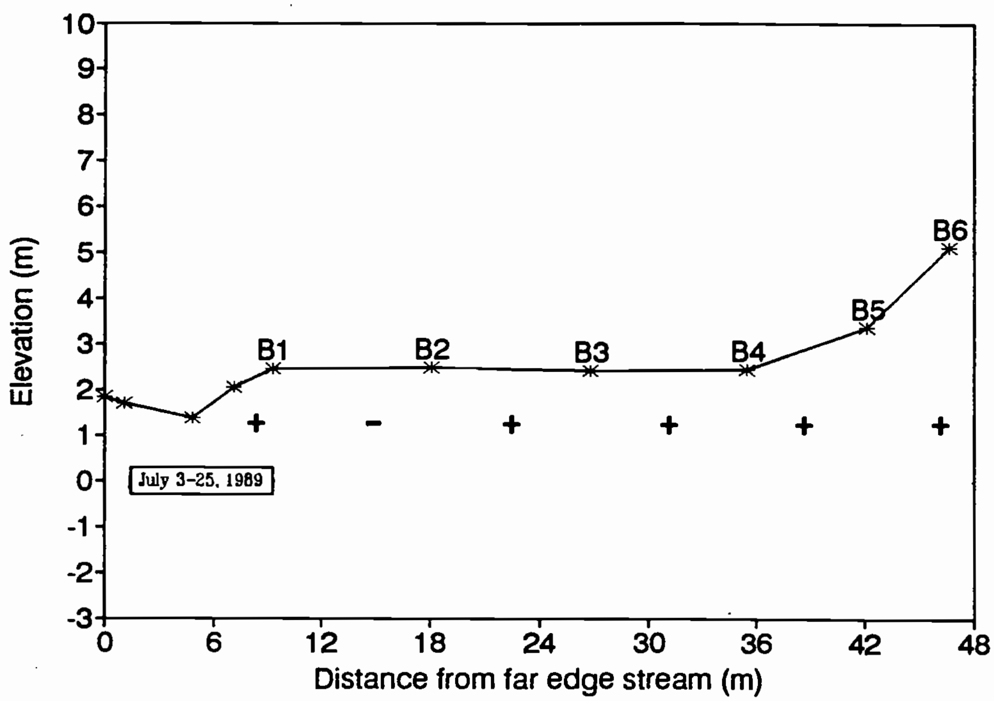

Figure 10. Direction of flow within the B transect from July 3-25, 1990. '+' indicates flow is towards the stream while '-' indicates flow is away from the stream.

The direction of flow patterns for December and February storms (Figure 11) are identical before, during, and after the peak of the hydrograph. It appears that B1 represents a ridge of relatively high hydraulic head causing flow to move towards both the stream and B2 before, during and after the peak discharge. Before the storms, the hydraulic head in B2 is lower than that of either B1 or B3. This may be a result of more rapid drainage because of the greater macroporosity associated with buried stream gravels in the B2 vicinity. However, at peak discharge, B2 develops a higher hydraulic head. This reversal of head may be due to the relatively large amounts of water moving through the buried stream gravels. Thus, during the storm, there is a ridge of relatively high hydraulic head that develops around B2 causing subsurface water to flow towards B3 and also downvalley within the buried channel. After the peak discharge occurs in the main channel the buried channel begins to drain quickly, causing the head to be lower than either B1 and B3 resulting in flow towards B2. Before, during, and after peak discharge, the direction of flow from B4 to B3, B5 to B4, and B6 to B5 shows a consistent pattern towards the stream.

The April storm provides a different pattern of flow directions within the B transect than were encountered for the December or February storms. Before the storm, the

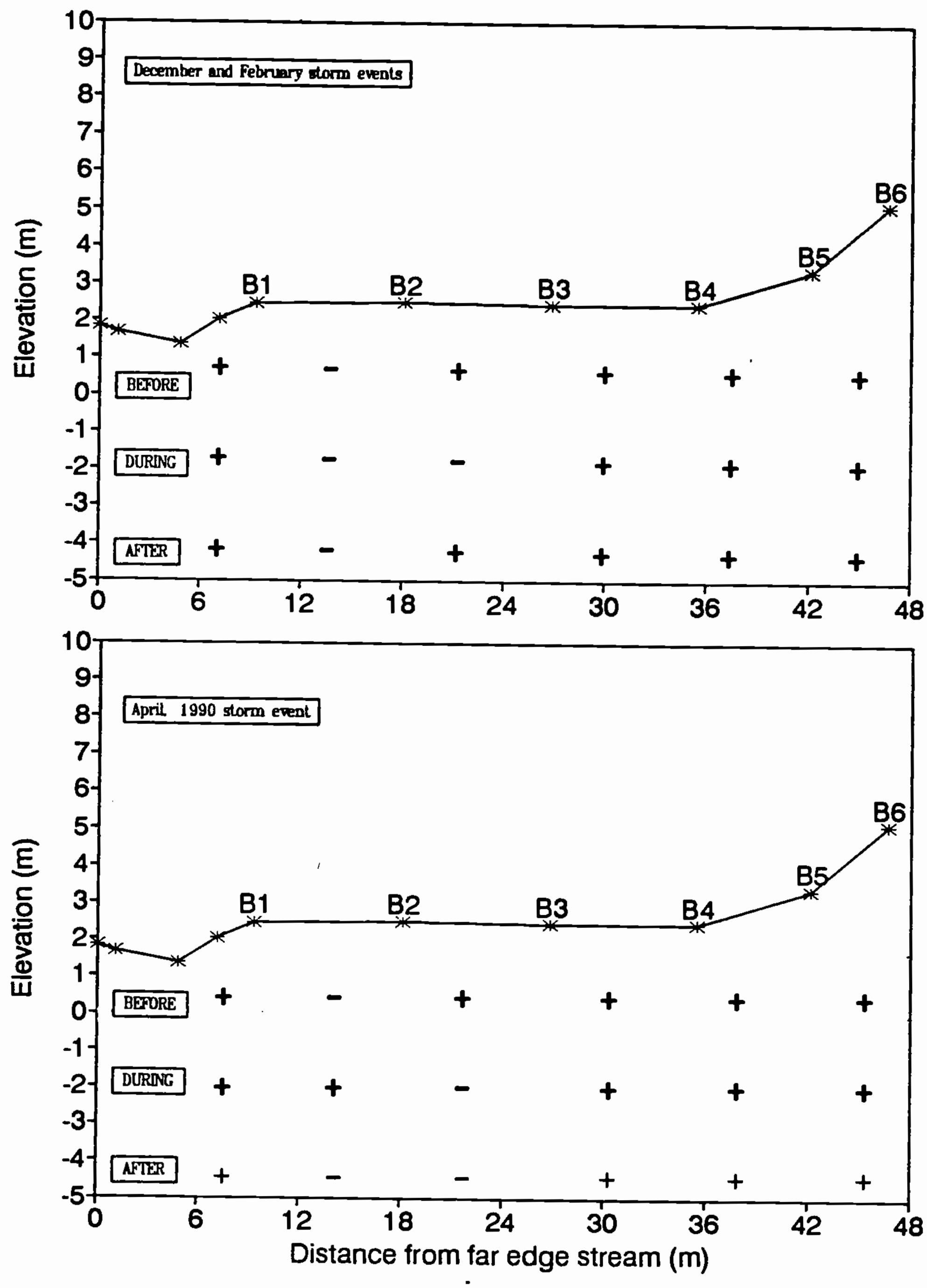

Figure 11. Direction of flow within the B transect for December, February, and April storm events. '+' indicates flow is towards the stream while '-' indicates flow is away from the stream.

directional flow pattern in April resembles those of December and February. During the occurrence of peak discharge for the April storm, the hydraulic head of B2 becomes higher than either B1 or B3, causing a ridge of higher hydraulic head that flows towards the stream, B3, and most likely downvalley. On April 29 (after the storm) there is another shift in the direction of flow and now B1 and B2 are flowing towards B3. Finally, on April 30 the directional flow pattern is identical to that of the before and after storm patterns in December and February.

These results indicate that the direction of flow within the B transect during the storm events are highly variable and depend on spatial and temporal circumstances. Similar situations may well exist throughout the remainder of the terrace. Apparently, there is not a uniform water table surface throughout the terrace that responds homogeneously. Thus, within section of terrace that appears to be relatively uniform topographically, there are zones that are more hydrologically dynamic than others depending on variations in subsurface conditions. To assume a homogenous and unidirectional flowing water table may cause other processes dependent on water table dynamics such as nutrient cycling, and stormflow and baseflow generation to be misinterpreted. These results indicate a better understanding of the hydrologic complexities of riparian areas is needed.

Question #2

Are subsurface flow velocities through a saturated soil matrix sufficient to generate stormflow in the Deer Creek study site?

This question was proposed after reviewing the literature on stormflow generation. Freeze (1972) found that subsurface velocities based on calculations using saturated K values (Equation 1) were generally too slow to generate stormflow. However, others (Beasley, 1976; Whipkey and Kirkby, 1978; Mosley, 1979) suggest that subsurface flow can have a relatively high velocity due to the presence of macropores. Thus, when subsurface flow utilizes macropores and not the soil matrix, it can be considered a feasible mechanism for stormflow generation.

To evaluate if matrix subsurface flow velocities are sufficient to generate storm peaks in the Deer Creek study site, three storm periods during the winter-spring of 1989-1990 were used to calculate travel times for water movement from the piezometer to the stream. A straight line distance from the piezometer to the stream was used to estimate minimum travel times. In addition a maximum distance based on water travelling at a 45° angle to the transect was calculated since this appeared to be a somewhat common angle from the direction of flow analysis (Question #1). Darcian flow was assumed and subsurface water velocities were calculated using Equation 1. For each piezometer the hydraulic gradient (dh/dl) was calculated to be from the piezometer to the stream. In the case of the hillslope piezometers, B5 and B6, travel time to the toeslope depressional area was calculated instead.

For all the storm events and piezometers, the minimum travel time (Table 2) was 33,000 hours or approximately 3.8 years! This minimum was for the B6 hillslope piezometer flowing into the toeslope depressional area where water has been observed flowing during periods of high precipitation. However, this calculated travel time is not sufficiently fast to account for stormflow generation. Thus, stormflow generation by soil matrix flow is not a feasible mechanism for the Deer Creek study site.

The results of this analysis appear to confirm the conclusions of other authors who have evaluated stormflow generation by subsurface mechanisms. In most cases it was discovered that subsurface velocities, based on soil matrix flow calculations from estimated values of hydraulic conductivity (K) are too slow to generate observed hydrograph peaks during storm events.

Other studies have indicated (Rahe et al., 1978; Beven and Germann, 1982; Mosley, 1982) that subsurface velocities typically occur in the 10^{-3} m/s range. A maximum velocity is in the range of 0.001 m/s and a

Table 2. Saturated hydraulic conductivity (K), travel length, maximum hydraluic gradient (HG), for December 2-5, 1889, February 6-12, 1990, and April 21-29, 1990 storm periods.

Travel path	K (m/s)	Travel length (m)	December Maximum HG	December Minimum travel time(hr)	February Maximum HG	February Minimum travel time(hr)	April Maximum HG	April Minimum travel time(hr)
B1-stream	1.5E-08	4.5	0.089	9.3E+05	0.185	4.5E+05	0.140	6.0E+05
B2-stream	3.3E-07	13.3	0.030	3.7E+05	0.034	3.3E+05	0.034	3.3E+05
B3-stream	6.0E-09	22.0	0.017	8.3E+07	0.017	5.8E+07	0.017	1.2E+08
B4-stream	2.4E-09	30.7	0.014	4.4E+08	0.014	2.3E+08	0.017	2.1E+08
B5-stream	2.4E-09	37.3	0.089	5.4E+05	0.039	1.2E+08	0.035	1.2E+08
B5-dep.	2.4E-09	4.4	0.104	4.9E+05	0.104	3.9E+04	0.070	7.4E+06
B6-stream	2.2E-07	40.9	*	*	0.044	5.3E+05	0.040	5.3E+05
B6-dep.	2.2E-07	9.0	*	*	0.134	3.7E+04	0.150	3.3E+04

* indicates missing data.

minimum in the range of 0.0001 m/s, while calculated subsurface velocities for the Deer Creek study site were from 10^{-8} to 10^{-11} m/s.

The reported maximum and minimum subsurface velocities (i.e., 0.001 and 0.0001 m/s) were used to recalculate subsurface travel times for conditions along Deer Creek to determine if reported values of macropore velocities could generate the storm peaks observed in the study area. Travel times were again calculated for a straight line travel distance to the stream and also for water travelling at a 45° angle from the orientation of the piezometer transect, in a downvalley direction. An angle of 45° was chosen after inspecting the results of the direction of flow analysis for the Deer Creek terrace, since it seemed to be a common angle.

During the December and February storm events the peaks in stage occurred 18 and 23 hours, respectively, after the onset of precipitation. A calculated subsurface travel time, based on velocities reported in the literature, for moving from furthest piezometer to the stream at a 45° angle is approximately 16 hours (Table 3), indicating that subsurface water moving at a velocity of approximately 10^{-3} m/s could partially contribute to stormflow generation. In addition, a portion of the hillslope flow could emerge at the toeslope depressional

Table 3. Subsurface travel times for the Deer Creek study site based on reported maximum and minimum subsurface velocities (Rahe et al., 1978; Beven and Germann, 1982; and Mosley, 1982).

Minimum straight line travel distance to the stream

Piezometer number	Travel distance (m)	Travel time(hr) using V_{min} (0.0001 m/s)	Travel time(hr) using V_{max} (0.001 m/s)
B1-stream	4.5	13	1
B2-stream	13.3	37	4
B3-stream	22.0	61	6
B4-stream	30.7	85	9
B5-stream	37.3	104	10
B5-depression	4.4	12	1
B6-stream	40.9	114	11
B6-depression	9.0	25	3

45° angle travel distance to the stream

Piezometer number	Travel distance (m)	Travel time(hr) using V_{min} (0.0001 m/s)	Travel time(hr) using V_{max} (0.001 m/s)
B1-stream	6.4	17	2
B2-stream	18.9	52	5
B3-stream	31.1	86	9
B4-stream	43.4	120	12
B5-stream	52.8	147	15
B5-depression	6.3	17	2
B6-stream	59.2	165	16
B6-depression	12.7	35	4

area, as observed in the field, and flow more rapidly to the stream, causing the travel time of water moving to the stream from the hillslope to be even lower. Thus, if the reported values of subsurface velocities from the literature more accurately portray the movement of water through the subsurface environment, then subsurface stormflow may occur in the Deer Creek study site.

The toeslope/hillslope environment of the study area has numerous animal burrows and a sizable amount of course fragments in the soil which is conducive to macropore development. Though the presence of a water table is reported to inhibit macropore development (Beven and Germann, 1984), the Deer Creek terrace has macropore networks that may be related to buried stream channels. During storm events piezometers that were on the hillslope (B6) and in the buried stream gravels of the terrace (B2) had hydrographs identical or nearly identical to stream hydrographs (Figure 12, 13, and 14) This situation, along with the time of travel analysis based on reported values of subsurface velocities indicates that macropore flow may play an important role in stormflow generation in Deer Creek.

The time of travel calculation used to evaluate whether soil matrix flow could generate stormflow assumed that the water at the piezometer moved directly to the stream. If translatory flow is operating, subsurface

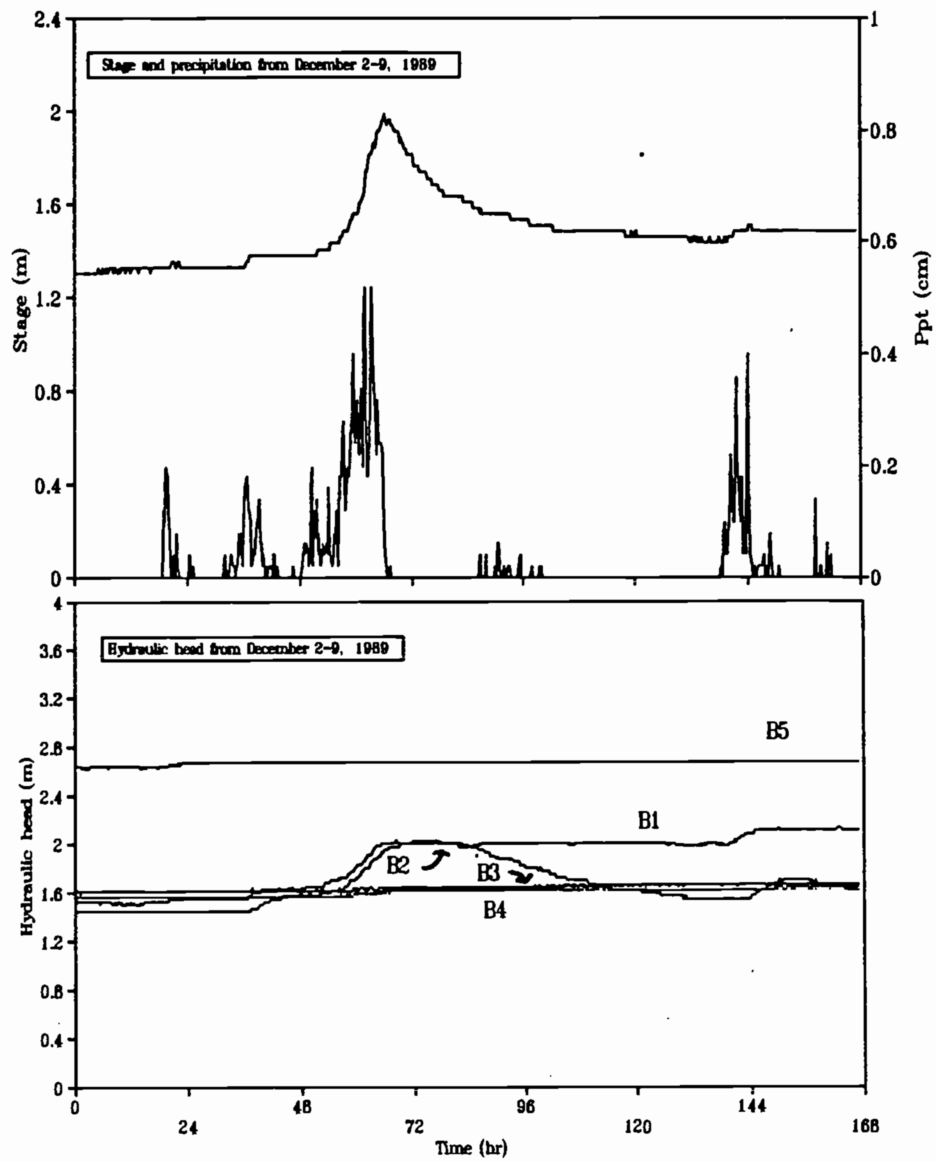

Figure 12. Stage, precipitation, and hydraulic head for December 2-9, 1989.

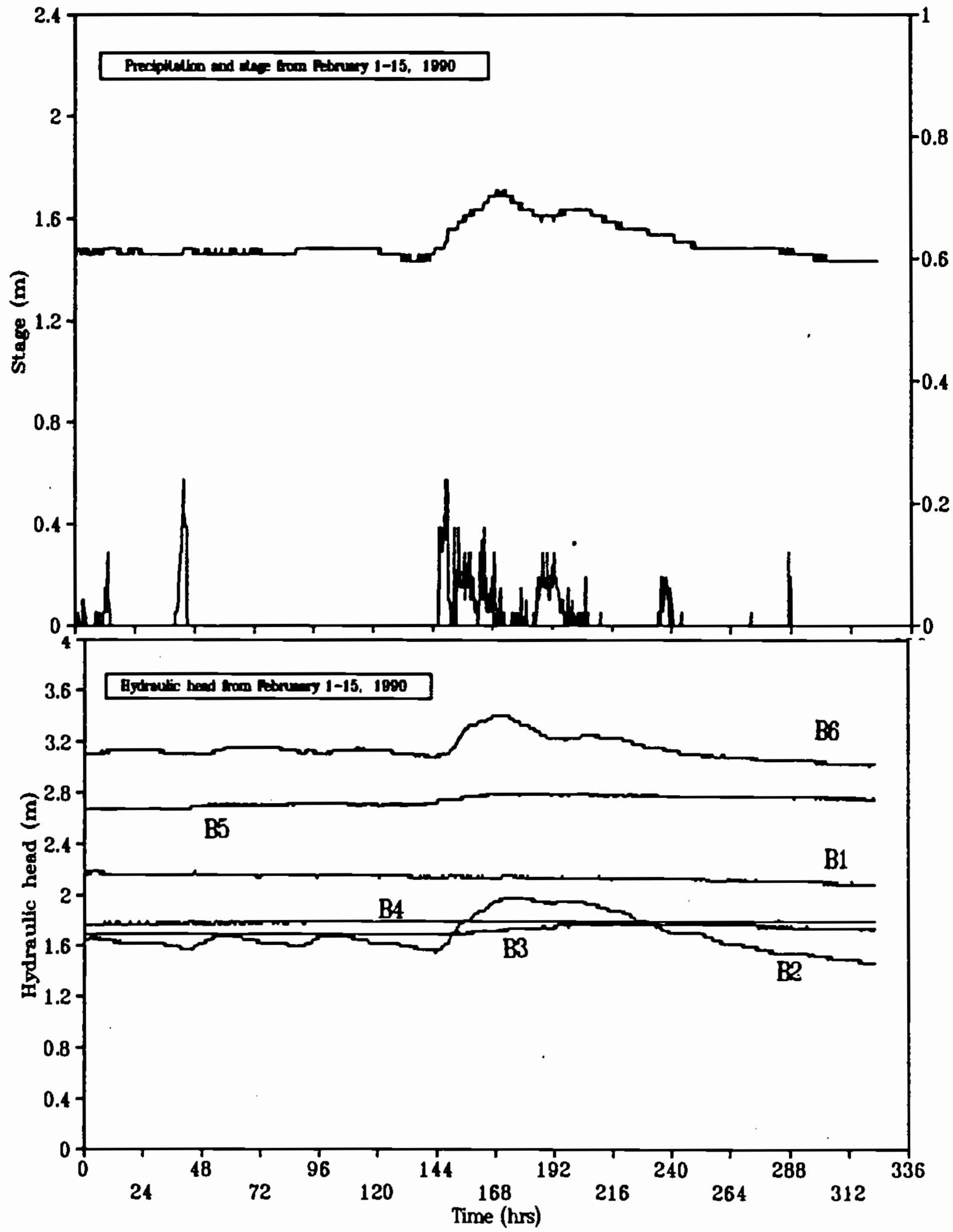

Figure 13. Stage, precipitation, and hydraulic head for February 1-15, 1990.

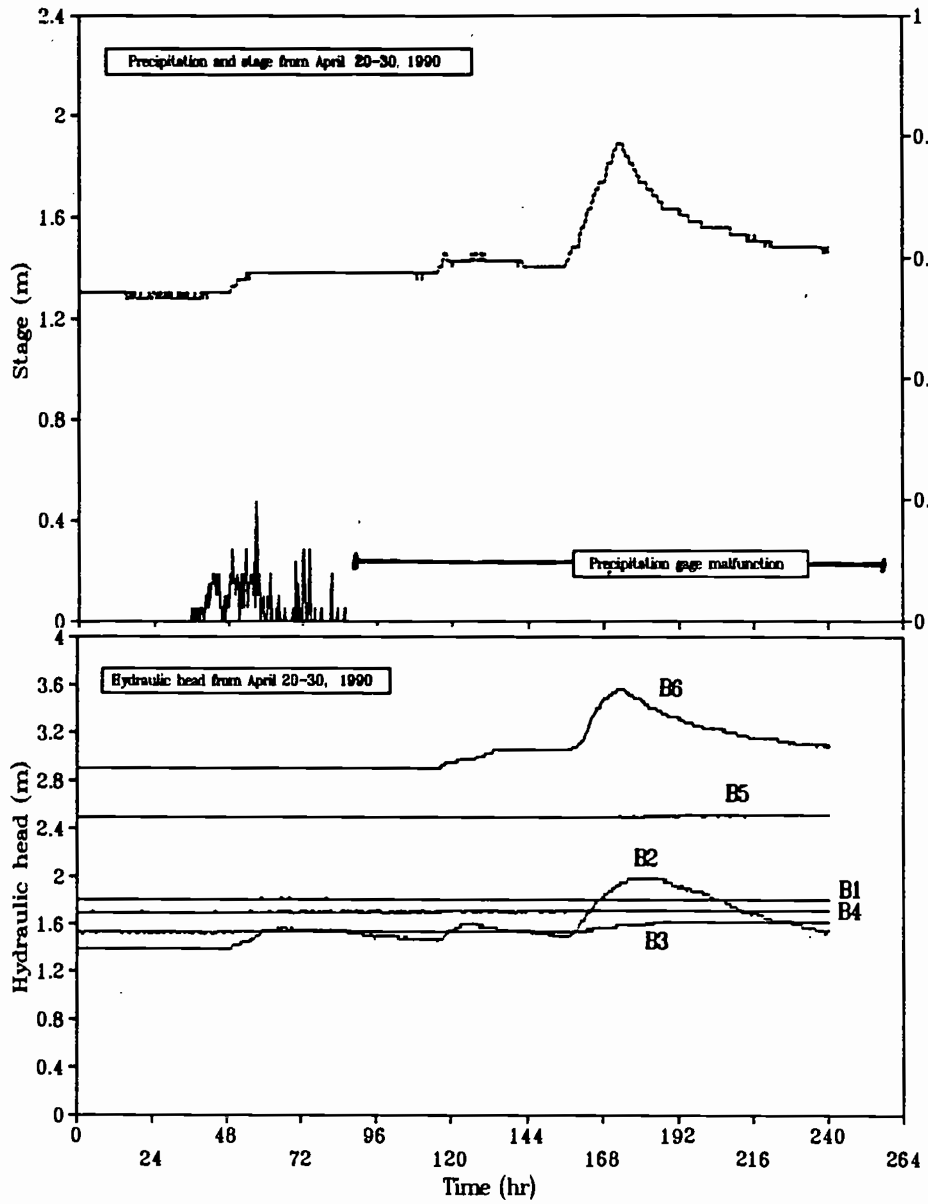

Figure 14. Stage, precipitation, and hydraulic head for April 20-30.

contributions to streamflow could still occur without the necessity of water traveling the entire distance. To evaluate this situation, another method was employed to calculate if matrix flow could generate streamflow. This method utilized the Q= -K(dh/dl)A (Equation 2) form of the Darcy equation. A steady-state period of discharge from 17:00-18:00 on December 4, 1989, prior to the hydrograph peak of the December 2-9, 1989 storm, was selected for analysis.

For this calculation:

Q = stream discharge at steady state.

K = hydraulic conductivity of the stream-adjacent piezometer, B1 = 1.5×10^{-3} m/s.

dh/dl = hydraulic gradient from the stream to the terrace = 0.08.

A = is the area of streambank above the study site stage gage contributing to streamflow = 2,399 m^2.

(A was estimated assuming an average bank height of 0.41 m and a channel length of 5853 m.

The resulting discharge is 2.9×10^{-8} cms, while the actual discharge at the time was 8.5×10^{-2} cms. These calculations indicate that, if the values of the parameters are correct, matrix flow cannot account for streamflow.

The greatest potential for error in this calculation is in the estimation of hydraulic conductivity from field methods. Error in estimates of K include: (1) the augering process used for piezometer installation caused slicksides in the soil and destroyed soil macropores, thereby reducing the K value, (2) all the assumptions of the Hvorslev method were not met, and (3) errors were made in fitting the line for the graphical portion of the Hvorslev method.

A hydraulic conductivity value was back-calculated to see if it fell within the range of K values indicated by Freeze and Cherry (1979). Using the actual stream discharge at the time, a K value of 4.4×10^{-4} m/s was calculated, which is representative of particles in the range of silty sand to clean sand. Since the streambanks along Deer Creek contain visible gravels (K= 1 to 10^{-3} m/s) and fine-textured soils, such as silty clay loams (K= 10^{-5} to 10^{-9} m/s) a K value of 10^{-4} m/s, as an average for the mixture of gravel and fine-textured soils, may be possible. Thus, without further refinement of the field technique for estimating K, the K values based on the Hvorslev method could underestimate the actual K values. If the field derived K values underestimate the actual K values by orders of magnitude, then it is possible soil matrix flow could account for streamflow during periods of winter baseflow. It appears unlikely, however, that soil matrix flow is a major contributor during periods of higher runoff.

Question #3

Does vertical infiltration from precipitation account for increased hydraulic head of the terrace piezometers during storm events?

This question was posed to determine whether the change in storage as indicated by the terrace piezometers could be accounted for by the vertical infiltration of precipitation alone, thereby discounting lateral inflow of subsurface water.

Three storm periods during December, January, and February of 1989-1990 were analyzed to address this question. If precipitation amounts during the storms exceed the change in storage of the piezometer then it is possible that vertical infiltration from precipitation could account for the change. If the depth of precipitation is less than the change in storage, then the increase in storage must also be a result of lateral influx of water from the hillslope, from another area within the terrace, or from the stream. However, a change in piezometric level does not necessarily indicate an identical change in storage. A rapid water table rise (in excess of that expected from precipitation) could be the result of the presence of a capillary fringe above the water table. If this is the case, a small amount of water can cause a large change in the water table (Novakowski and Gillham, 1988), especially in fine textured soils.

During the three storm periods precipitation was accumulated and graphed against the change in storage implied by the terrace piezometers. The change in storage, expressed as a depth of water, was calculated for each terrace piezometer by first subtracting a pre-storm hydraulic head from each storm head and then multiplying the resultant depth by the average porosity of the terrace (n=0.62) to obtain a storage depth. Initial values of hydraulic head for each of the terrace piezometers and the initial stream stage for each storm event are shown in Table 4.

In the December storm (Figure 15) there was a rapid accumulation of precipitation (15 cm in 24 hours) and a rapid rise in stage and water levels for both B1 and B2. However, water levels in B1 did not recede following cessation of precipitation as occurred with stage and the B2 piezometer. The reason for this is not entirely clear since the B1 piezometer behaved differently than either of the other stream-adjacent piezometers (A1 and C1). During the December storm, the first major storm of the winter, the B1 piezometer rose rapidly and remained high relative to A1 and C1 throughout the year. Piezometers B3 and B4 showed little response to storm precipitation.

The late January storm had a greater amount of precipitation accumulation than did the December storm but it was over a 200 hour period (Figure 16). During this

Table 4. Terrace hydraulic head (m) and stage (m) initial values for each storm period.

Piezometer	Dec 2, 1989	Jan 22, 1990	Feb 6, 1990
Stage	1.30	1.36	1.48
B1	1.52	2.24	2.13
B2	1.44	1.44	1.57
B3	1.56	1.61	1.69
B4	1.61	1.71	1.79

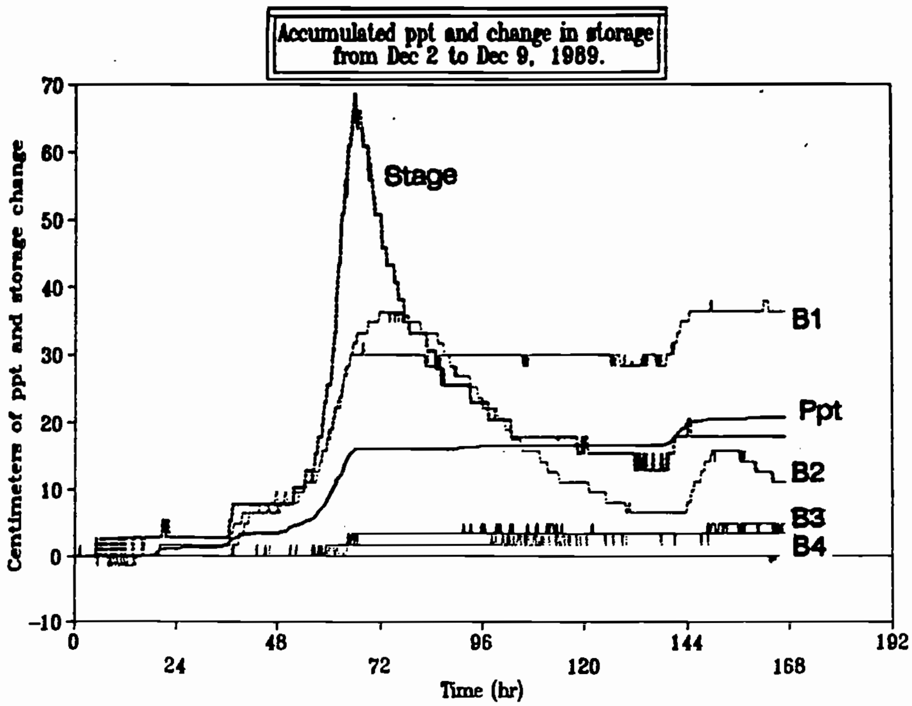

Figure 15. Accumulated precipitation and change in storage from December 2-9, 1989.

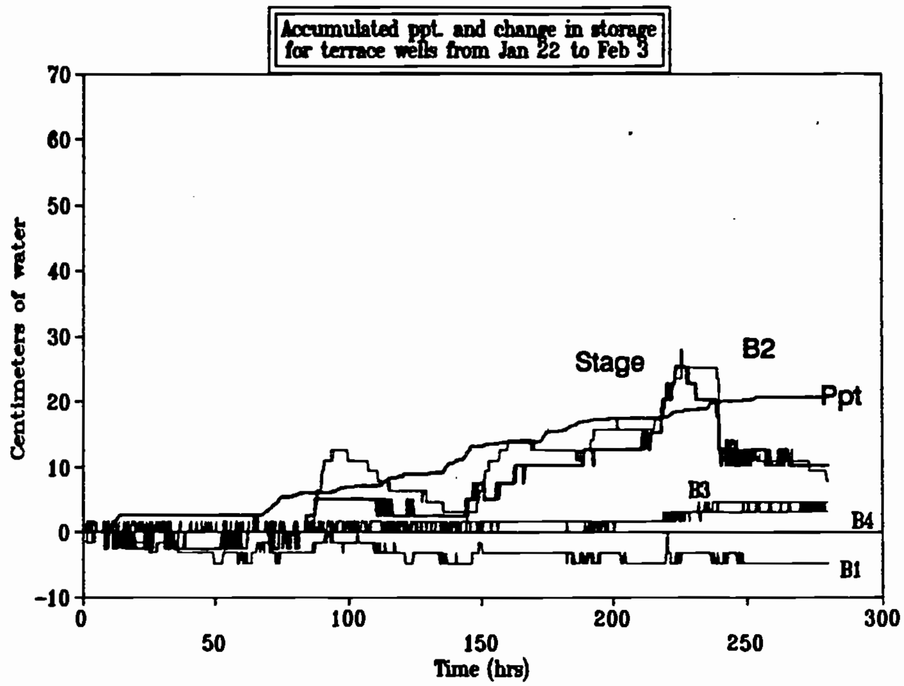

Figure 16. Accumulated precipitation and change in storage from January 22-February 3, 1990.

storm water levels in B2 parallel stage changes. In contrast, B1 shows a decrease in storage. A possible explanation for the decrease in storage in B1 is that it is discharging to the stream and perhaps into the buried channel at B2 at a faster rate than precipitation is accumulating in the profile.

The February storm is the smallest of the three with only a 12 cm of accumulated precipitation over 80 hours (Figure 17). During this storm, storage changes in B2 increased more rapidly than changes in either precipitation or stage. While the B1 piezometer showed essentially no response.

For all three storm events the B3 and B4 terrace piezometers showed a similar pattern of response which consisted of little to no change in storage. The piezometers in this area had saturated K values that were relatively low compared to the two piezometers closer to the stream. Their lack of response to storm precipitation may be due to water movement that is too slow to produce appreciable changes in storage. In addition, subsurface drainage from hillslope water may be routed through the toeslope colluvium and out onto the depressional area adjacent to the toeslope. This water is rapidly transmitted as surface runoff to the channel, and therefor is not available to recharge the hillslope-adjacent area.

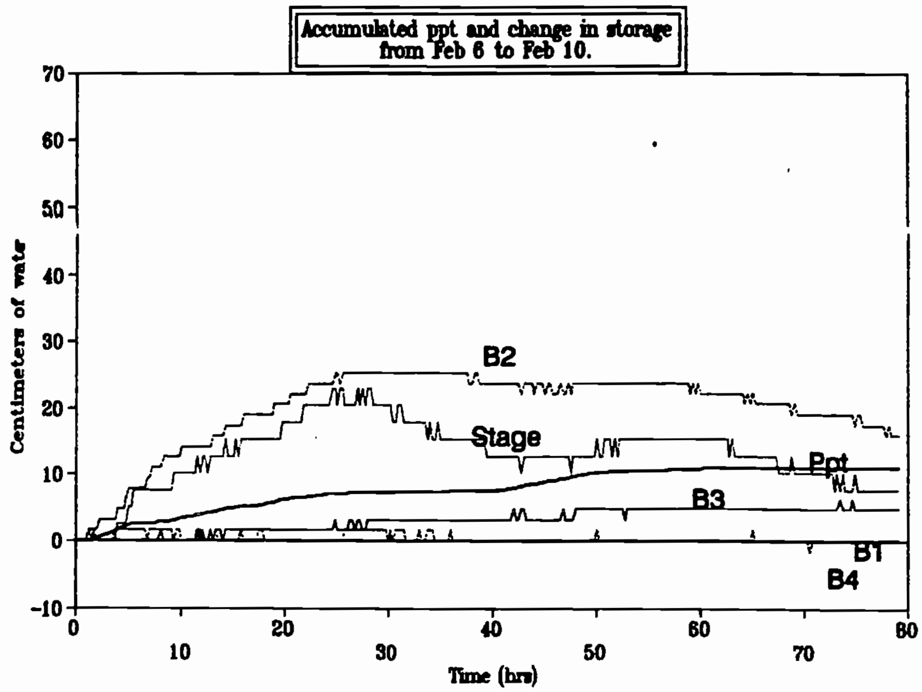

Figure 17. Accumulated precipitation and change in storage from February 6-10, 1990.

Thus, it would appear that vertical infiltration from precipitation cannot consistently account for the rapid changes in storage in the terrace piezometers.

As mentioned previously, the capillary fringe theory relates the rapid rise in the water table to small amounts of precipitation inputs to the presence of a capillary fringe. Where soil textures are fine (i.e. silts and clays) the capillary fringe can extend to the ground surface. In the Deer Creek study site the terrace soils tend to be finer textured except where there are buried stream gravels. The most rapid rise in the water table is generally seen in the piezometer (B2) located in the coarser textured soil (i.e. stream gravel). This appears to be contrary to the capillary fringe theory. The occurrence of a capillary fringe assumes a uniform soil matrix; however, where there are soil discontinuities such as sand or gravel lenses, coarse fragments, or large macropores, the capillary fringe may not be as effective in causing rapid rises in the water table.

The results seem to indicate that vertical infiltration from precipitation and the capillary fringe effect are not as critical in producing rapid changes in storage in the terrace piezometers as is the transmittal of water through the more porous matrix of the buried stream channels.

Question #4

Throughout the year does the terrace recharge the stream or does the stream recharge the terrace?

It is common for mesic streams to be characterized as effluent, i.e., stream-adjacent areas recharging the stream, as opposed to xeric streams which are characterized as influent i.e., the stream recharges the adjacent area. The intent of this question is to determine whether headwater Coast Range streams are strictly effluent along a reach.

To address this question, the hydraulic heads of stream-adjacent piezometers (A1, B1, and C1) were compared to stream stage. Stream stage at each piezometer location was calculated using the stream slope and the distance downstream of the piezometer from the stage gage. Stream stage was then subtracted from the hydraulic head of the stream adjacent piezometer. A positive result indicated that the terrace was recharging the stream while a negative value indicated the stream was recharging the terrace. The amount of time the terrace was recharging the stream for each stream adjacent piezometer was expressed as a percent.

Results indicate that there are varying recharge/discharge zones within a few meters of one another even along a fairly homogeneous appearing stream reach (Table 5). Variations in microtopography and soil

Table 5. Percent of month for the 1990 water year when terrace piezometers had hydraulic head greater than stream stage.

Month	A1	B1	C1
Oct	100	0	0
Nov	100	50*	0
Dec	100	100	0
Jan	100	100	0
Feb	67	100	0
Apr	100	100	67
May	50	100	50
June	100	100	0
July	33	100	0

*For November 25% of the month the terrace and stream were equal.

properties may be critical in determining the location of the discharge/recharge area.

Characterizing streams as influent or effluent is undoubtedly an oversimplification of natural systems. Along a single reach there may be several interspersed influent/effluent areas, and these may change seasonally. Thus, streams, and stream-adjacent areas, interact dynamically both spatially and temporally.

Question #5

Do stream adjacent terraces in small forested watersheds store sufficient water for sustaining baseflow during summer months?

This question was addressed after reviewing literature on baseflow discharge from stream-adjacent areas. Kondolf et al. (1987) and others suggest that in broad alluvial valleys the release of water from bank storage is critical in maintaining low flows. However, this question has not been addressed for small, forested headwater streams and so this question was proposed.

To determine if a terrace in a small headwater stream contains sufficient water for release to sustain or significantly augment baseflow during summer months a mass balance was done on the incoming and outgoing sources of water within the Deer Creek Watershed. To accomplish this, three assumptions were made (Rawitz et al., 1970): (1) the law of conservation of mass is valid, (2) the topographic boundaries of the watershed are also the groundwater boundaries, and (3) the watershed does not leak. Thus, for the Deer Creek study, the boundaries of the mass balance are the watershed area (3,030,000 m^2) and the terrace area (233,000 m^2). The equation used for the mass balance for the Deer Creek terrace is:

$$Ppt + Hillslope\ Q + \Delta S = Q + ET \qquad (10)$$

where Ppt is precipitation, hillslope Q is the subsurface flow from the hillslope, ΔS is change in storage, Q is stream discharge, and ET is evapotransipration.

Incoming sources of water to the stream for the mass balance include precipitation, the release of stored water from within the terrace, and hillslope subsurface discharge. For this exercise the hillslope subsurface discharge is not quantified. The outgoing sources of water include stream discharge and evapotranspiration. Each of these components were calculated on a volumetric basis for the months of June and July, 1990.

The change in storage for the terrace was first calculated starting June 12, 1990. The change in storage for the terrace was calculated by subtracting the initial piezometric level from the ground surface elevation and then subtracting the lower subsequent piezometric levels from that level and then multiplying the difference by the porosity to get a depth of water. This method may overestimate the actual volume of water released since using porosity (n) assumes that an amount of water equal to the pore space is drained. This may not be the situation, especially in fine-textured soils, where a portion of the water is retained on the soil particles and therefor not readily released. Thus, estimates of change in storage based on porosity may overestimate the actual amount of water released from storage.

To calculate an average porosity for the terrace, a bulk density (Pb) of the terrace soils was needed. Eight bulk density samples were obtained using a bulk density core sampler. The samples were then oven dried at 105° Celsius until their weight stabilized and the bulk density (Pb) was then calculated by the following equation:

$$Pb = \frac{Oven\ dried\ weight}{Original\ volume} \tag{11}$$

The porosity (n) was then calculated as follows:

$$n = 1 - \frac{Pb}{Ps} \tag{12}$$

Ps (the particle density) was assumed to be 2.65 gm/cm^3, which is the value used for most mineral soils. The average bulk density for the eight samples was 0.62 +/- 0.03 gm/cm^3.

The actual and the potential evapotranspiration values were calculated for the watershed and the terrace using the Thornthwaite method. The potential evaporation for June is 8.9 cm which is approximately equal to the actual evapotranspiration for that month. The potential evapotranspiration for July is 9.9 cm while the actual is 6.48 cm.

The results of the mass balance analysis (Table 6) indicate that the release of storage from stream-adjacent

Table 6. Mass Balance for Deer Creek Watershed.

Period	PPT entire watershed (m^3)	ET entire watershed (m^3)	Terrace Δstorage (m^3)	Q stream (m^3)	%ΔS (terrace) of ET + Q
June 12-19	30,303	62,899	3,194	40,919	3.1
June 19-26	8,601	62,899	3,427	28,547	3.8
June 26-July 3	36,970	62,899	1,445	20,926	1.7
July 3-25	21,212	144,461	11,492	58,164	5.7

terraces in the Deer Creek Watershed cannot provide sufficient water to maintain baseflow. If there was an equivalent amount of release from storage for the entire watershed the percentages approach 80-90 % of outgoing water. Given the large volumes of outgoing water from the basin in comparison to that released by the terrace there must be significant quantities of water from hillslope soils and regolith.

CONCLUSIONS

Riparian areas have been described as areas of great species diversity compared to the rest of the landscape (Gregory, in review). The results of this study indicate that riparian areas are also areas of great hydrologic diversity compared to upslope areas. In steeply sloping forested watersheds the overriding direction of water movement is downslope. In riparian areas that occupy flatter, unconstrained valley reaches the direction of water movement may not necessarily be downvalley or directly towards the stream throughout the year.

The spatial and temporal variation in riparian subsurface flow within the study site may be influenced by microtopography and the several sources of water that interact there such as lateral inflow, deep groundwater flow, stream flow and precipitation. This hydrologic spatial and temporal variation could contribute greatly to the observed plant species diversity in riparian areas. Within the riparian area there are zones where the water table is closer to the surface and likewise there are zones where the water table is far from the surface, depending on the time of year. This variation creates sites for both hydrophyllic and non-hydrophyllic plant species within the riparian area.

The spatial and temporal heterogeneity observed in the Deer Creek study is also important in other watershed characteristics such as stormflow and baseflow generation. Local variations in the subsurface environment such as the presence of buried stream channels seem to be important in stormflow generation and perhaps are important for nutrient cycling by creating zones of faster water movement through the terrace.

The simplification of hydrologic responses may be necessary for mathematical modeling streamflow from forested watersheds but may not accurately reflect the true hydrologic functioning of the 'hillslope-riparian area-stream' continuum through space and time. Thus, at least conceptually, these areas must be viewed as complex sites of several hydrologic interactions that change in response to physical and temporal circumstances.

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APPENDICES

APPENDIX A. SOIL CORE DESCRIPTIONS

All cores were obtained during September 15-20, 1989.

A1	Core depth=1.8 m. Saturation at 1.5 m.
0-32 cm.	Soil heavily compacted at surface. Loam.
32-50 cm.	Lighter color, increased fine sands.
50-70 cm.	FSL, greyish color.
70-97 cm.	Coarse fragments (CFs). Coarse SL. Brown weathering skins on the CFs.
97-141 cm.	Saturation beginning at 97 cm. Loamy sand, some CFs. Grey color.

A2	Core depth=1.8 m. Saturation at 1.3 m.
0-25 cm.	SiL to SiCl.
25-71 cm.	SiL with mottles.
71-85 cm.	SL, slight gleying and mottles.
85-108 cm.	LS, mottles.
108-128 cm.	SL, few mottles.
108-132 cm.	Gravel layer.
132-154 cm.	SL, few mottles, dark grey. Saturated.

A3	Core depth=1.8 m. Saturation at 1.7 m.
0-25 cm.	Loam or SiL. Brown.
25-45 cm.	Loam becoming more greyish. Mottles.
45-70 cm.	SiL to SiCL. Some mottling.
70-88 cm.	SL, greyish, some mottling and CFs.
88-180 cm.	SiCL. Dark grey. Many mottles from 170-180 cm.
180-185 cm.	Grey sand.

A4	Core depth=1.5 m. Saturation at 1.1 m, also buried wood.
0-45 cm.	Loam. Brown.
45-105 cm.	Loam or SiL, darker color, picking up more moisture.
105-115 cm.	SiCL, grey, wood pieces.
115-159 cm.	SiCL with some coarse sand. Grey, some mottling.

A5	Core depth=1.5 m. Saturation at 90 cm.
0-30 cm.	Loam with small CFs. Dark brown.
30-48 cm.	SL with some coarse sand and gravel. Light brown.
48-68 cm.	SiL some CFs, getting more grey.
68-98 cm.	Coarse SL, greyish brown. Saturation at 90 cm.

A6	Core depth=2.7 m. Numerous animal burrows.
0-45 cm.	SCL, numerous CFs. Mixing due to animal activity. 10 YR 4/3.
45-95 cm.	SCL, numerous CFs, some charcoal, highly mixed. 10 YR 4/4.
95-130 cm.	SCL, less mixing, becoming moist. 10 YR 4/3.
A7	Core depth=2.7 m. Numerous animal holes.
0-83 cm.	SiCL numerous CFs.
83-154 cm.	SiCL, numerous CFs, lighter color. Charcoal at 143 cm.
154-201 cm.	SiCL, numerous CFs and coarse sand, increased clay content.
B1	Core depth=1.8 m. Saturation at 1.4 m.
0-23 cm.	Loam, brown.
23-50 cm.	Loam with more fine sand, mottles in light grey-brown matrix.
50-110 cm.	SiCL, greyish, some mottles.
110-125 cm.	Loam with coarse sand mixed in, some mottling.
125-169 cm.	SiL, saturated, grey, some mottles.
B2	Core depth=1.8 m. Saturation at 1.1 m.
0-40 cm.	Loam, brown.
40-86 cm.	SL mottles, grey-brown.
86-114 cm.	SiL, grey, some mottles.
114-154 cm.	SiL, darker grey, few mottles in upper horizon. Gravel in very bottom.
B3	Core depth=1.8 m. Saturation at 1.7 m.
0-40 cm.	Loam, brown.
40-56 cm.	SiL, grey-brown.
56-110 cm.	SiL with coarse sand, grey with mottles.
110-131 cm.	SL, grey-brown, mottles.
131-144 cm.	SiL, grey with mottles.
144-174 cm.	Loam, grey to grey-brown, few mottles.
174-180 cm.	SL, liquified.
B4	Core depth=1.5 m. Moisture at 1.4 m.
0-14 cm.	Loam, brown.
14-48 cm.	SiL becoming more grey, some mottling.
48-84 cm.	SiL, dark grey to grey-brown, some mottling.
84-110 cm.	SiL, dark grey-black, some mottles.
110-132 cm.	SiCL, dark grey-brown, some blue-grey sands.
132-146 cm.	SiCL, grey-brown.

B5	Core depth=1.5 m. Saturation at 1.0 m.
0-19 cm.	Brown loam with beige sandy CFs.
19-59 cm.	SiL with gravel and coarse sand, lighter brown.
59-81 cm.	SiL with reddish sand pockets. Grey-brown, CFs.
81-96 cm.	SiL, CFs, dark grey to grey-brown.
96-109 cm.	SiL, CFs, grey-brown, some mottles, saturated.
B6	Core length=2.7 m. Some moisture at 2.1 m.
0-90 cm.	SiCL, dark brown, CFs, 10 YR 2/2.
90-181 cm.	SiCL, CFs, 7.5 YR 3/2.
181-247 cm.	SiCL, CFs, 10 YR 5/6.
247-255 cm.	Large gravel pieces, CL matrix, 10 YR 5/6.
B7	Core depth=2.7 m.
0-120 cm.	Loam, CFs, 10 YR 2/2.
120-220 cm.	Loam, many CFs, 10 YR 3/3.
220-260 cm.	Loam with CFs, 10 YR 4/3.
C1	Core depth=1.8 m. Moisture at 1.2 m.
0-20 cm.	Loam, brown.
20-41 cm.	Loam, picking up fine sand, lighter brown color.
41-57 cm.	SL, grey-brown.
57-73 cm.	SL, grey-brown, some mottles.
73-98 cm.	LS, grey-brown, some mottles.
98-125 cm.	LS, red-brown lens @98-103 cm, dark grey below.
125-133 cm.	SiL, grey-black.
C2	Core depth=1.8 m. Buried stream channel at 1.0 m. Saturation at 1.30 m.
0-42 cm.	Loam, 10 YR 3/4.
42-55 cm.	Loam with some fine sand lens, 10 YR 4/2.
55-85 cm.	SiL, some mottling, 10 YR 3/2.
85-107 cm. and	SiL, fine sand, some mottling, sand lens CFs in lower horizon, 10 YR 3/3.
107-137 cm.	Stream gravels in LS matrix, saturation at 130 cm., 10 YR 3/3.
C3	Core depth=1.8 m. Saturation at 1.50 m.
0-30 cm.	Loam, 10 YR 4/3.
30-77 cm.	SiCL, grey, mottles, 10 yr 4/2.
77-91 cm.	SL mottles, 10 yr 4/1.
91-121 cm.	SiL, mottles, 10 yr 4/2.
121-158 cm.	SiL, buried wood, some mottles in upper portion, saturation at 150 cm., 10 yr 3/1.

C4	Core depth=1.5 m. Buried wood at @ 1.2 m. Saturation at 0.5 m.
0-11 cm.	Loam, dark organic layer, 10 yr 2/2.
11-27 cm.	SiL, 10 yr 3/2.
27-52 cm.	SiCL, some mottles, 10 yr 3/2.
52-81 cm.	SL, buried wood, saturation, 10 yr 3/2.
81-111 cm.	SiL, dark grey, 10 yr 3/1.
C5	Core depth=1.5 m. Saturation at 1.4 m.
0-85 cm.	Loam, high organic matter content, CFs, some charcoal, 7.5 YR 3/4.
85-105 cm.	SiCL, tan CFs, some organic matter, dark grey, 10 YR 3/2.
105-123 cm.	SiCL, lighter grey CFs, 10 YR 3/2.
123-142 cm.	SiCL with CFs, 10 YR 4/1.
142-159 cm.	SCL, CFs, saturated, 10 YR 5/1.
C6	Core depth=2.7 m. Numerous animal burrows.
0-43 cm.	Loam, rotten wood mixed in, 10 YR 2/2.
43-65 cm.	SCL, CFs, some woody pieces, 10 YR 3/2.
65-87 cm.	SCL, some CFs, 10 YR 3/3.
87-110 cm.	SCL, CFs, 10 YR 3/4.
110-130 cm.	CL, some coarse sand, CFs, 10 YR 5/6.
130-177 cm.	CL, some coarse sand, CFs, 10 YR 6/8.
C7	Core depth=2.7 m. Numerous animal burrows.
0-119 cm.	Loam, mixing from animals, CFs, 10 YR 2/2.
119-134 cm.	Loam to CL, CFs, 10 YR 3/3.
134-178 cm.	SiL, coarse sand, CFs, some charcoal, 10 YR 5/6.
178-191 cm.	CL, small CFs, yellow orange tight clay, 10 YR 6/8.

APPENDIX B. DETAILED VEGETATION SURVEY OF THE DEER CREEK STUDY SITE

This survey was conducted on June 12, 1990 by Rob Pabst of the Forest Science Laboratory, Corvallis, OR. The transect and plot numbers correspond to those on Figure B-1. The species abbreviations are listed in Table B-1.

Transect 1-Plot 1

Species	% Cover
BRVU	25
OXOR	25
MOSI	1
STME2	2
POMU	30
DIFO	1
RUURM	4

Transect 1-Plot 2

Species	% Cover
STME2	25
DAGL	60
POTR3	5
RARE	5
TOME	2
CRDO	1
ELYMU	10

Transect 1-Plot 3

Species	% Cover
STME2	3
DAGL	10
OESA	5
MOSI	1
TOME	2
GAAPE	1
STCR	1
ELYMU	65
POTR3	5
RUSP	5

Transect 1-Plot 4

Species	% Cover
MOSI	20
VIGL	10
TOME	10
RUSP	10
GLYCE	25
POTR3	5
GAAPE	1
STME2	6
STCR	1

Transect 1-Plot 5

Species	% Cover
BRVU	15
OESA	4
STME2	20
MOSI	5
EQAR	1
STCR	2
TOME	10
VIGL	1
RUSP	40

Transect 1-Plot 6

Species	% Cover
OXOR	70
RARE	5
CHGL	10
GLYCE	1
TOME	3

Transect 1-Plot 7

Species	% Cover
BRVU	15
TOME	50
STME2	15
OXOR	2
DIFO	1
STCR	1
CHGL	5
RUSP	25

Transect 2-Plot 1

Species	% Cover
ELYMU	25
DAGL	50
HOLA	5
POTR3	5
STME2	5

Transect 2-Plot 2

Species	% Cover
HOLA	5
POTR3	50
DAGL	10
MELIC	5
EQAR	3
STME2	10
RULA2	2

Transect 2-Plot 3

Species	% Cover
URDI	6
STME2	20
DAGL	10
GAAPE	1
TOME	1
HOLA	2

Transect 2-Plot 4

Species	% Cover
OESA	3
POTR3	50
TOME	5
STME2	10
STCR	1
MOSI	1

Transect 2-Plot 5

Species	% Cover
MOSI	3
TOME	25
DAGL	1
BRVU	5
STME2	2
VIGL	2
DIFO	3
RUSP	75
STCR	1

Transect 2-Plot 6

Species	% Cover
LYAM	25
GAAPE	1
OXOR	40
STME2	8
RARE	2
TOME	4
BRVU	3
RUSP	5
GLYCE	5

Transect 2-Plot 7

Species	% Cover
MEBU	4
DIFO	2
OXOR	50
MOSI	2
CADE	2
TOME	6
RUSP	10
POMU	10
STCR	1
VIGL	1

Transect 3-Plot 1

Species	% Cover
ELYMU	30
TOME	3
STME2	8
POTR3	2
VIGL	2
STCR	1
RARE	1
RUSP	35

Transect 3-Plot 2

Species	% Cover
SARA	5
RUSP	2
JUNCU	1
CADE	5
TOME	10
MOSI	1
ELYMU	50
POTR3	5
BRVU	5
ATFI	2

Transect 3-Plot 3

Species	% Cover
TOME	60
VIGL	2
STME2	20
GAAPE	1
POTR3	2
ATFI	1
HOLA	2
RUSP	70
BRVU	1

Transect 3-Plot 4

Species	% Cover
EQAR	4
SARA	5
STME2	10
BRVU	10
TOME	35
STCR	5
RUSP	30
VIGL	1

Transect 3-Plot 5

Species	% Cover
TOME	60
STME2	15
GAAPE	4
RUSP	90
MOSI	1
BRVU	10
STCR	5
VIGL	2

Transect 3-Plot 6

Species	% Cover
CHGL	3
GAAPE	2
TOME	70
GLYCE	5
STME2	15
CRDO	3
OXOR	5
VIGL	1
MOSI	1
STCR	1

Transect 3-Plot 7

Species	% Cover
STME2	40
OXOR	40
RUSP	30
BRVU	2
MEBU	4
POTR3	1
POMU	25
TOME	3
STCR	1
GAAPE	1

Transect 4-Plot 1

Species	% Cover
ELYMU	8
BRVU	10
STME2	4
RARE	2
VIGL	3
STCR	1
ATFI	5
TOME	5
RUSP	35

Transect 4-Plot 2

Species	% Cover
ATFI	15
CAOB	10
RUSP	25
VIGL	5
TOME	2
MOSI	1
BRVU	1

Transect 4-Plot 3

Species	% Cover
CAOB	40
VIGL	2
RUSP	90

Transect 4-Plot 4

Species	% Cover
CAOB	80
STME2	5
RUSP	8
TOME	1
VIGL	1

Transect 4-Plot 5

Species	% Cover
CAOB	95
STME2	4

Transect 4-Plot 6

Species	% Cover
BLSP	4
OXOR	4
RARE	4
MOSI	2
DIFO	3
STME2	5
CADE	10
VIGL	1
TOME	1
POMU	5
BRVU	2
FESTU	5

Transect 4-Plot 7

Species	% Cover
RUSP	5
SARA	2
BRVU	35
DIFO	1
OXOR	8
MOSI	2
GAAPE	1
VIGL	1
STCR	1
STME2	3
FESTU	1

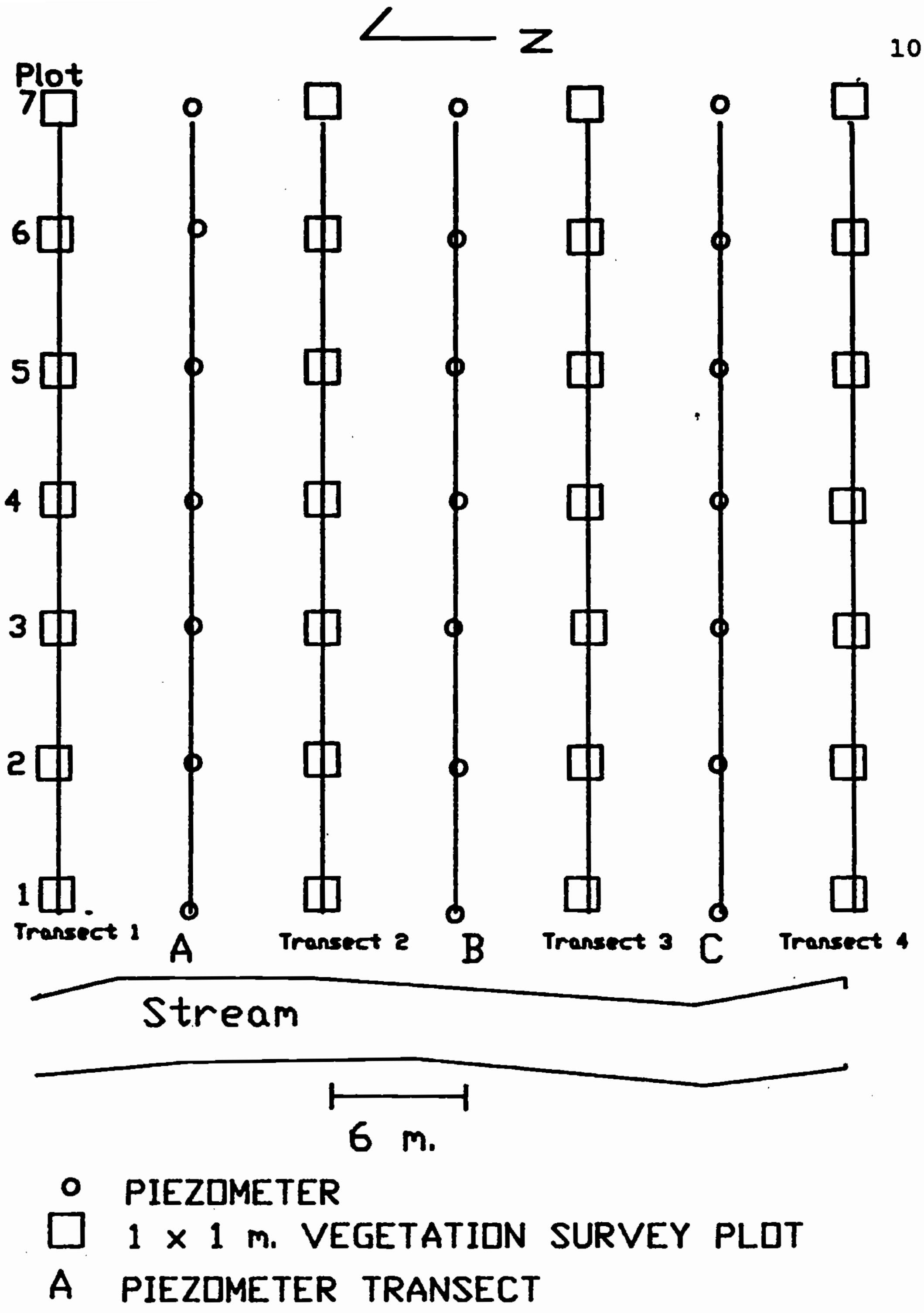


Figure B-1. Vegetation survey plot location.

Table B-1. Deer Creek study site plant species list.

HERBS

Species name	Abbreviation	Common name
Lysichitum americanum	LYAM	skunk cabbage
Stellaria crispa	STCR	
Equisetum arvense	EQAR	horsetail
Dicentra formosa	DIFO	bleeding heart
Staychs mexicana	STME2	
Oxalidaceae (Oxalis)	OXOR	Oregon oxalis
Montia siberica	MOSI	Indian lettuce
Ranunculus repens var. repens	RARE	creeping buttercup
Galium aparine var. echinospermum	GAAPE	goosegrass, bedstraw
Chrysoplenium glecomaefolium	CHGL	golden saxifrage
Tolmiea menziesii	TOME	pig-a-back plant
Oenanthe sarmentosa	OESA	wild parsley
Urtica dioica	URDI	stinging nettle
Viola glabella	VIGL	yellow violet

SEDGES,GRASSES AND RUSHES

Species name	Abbreviation	Common name
Carex deweyana	CADE	
Carex obnupta	CAOB	
Elymus sp.	ELYMU	
Festuca sp.	FESTU	
Glyceria sp.	GLYCE	
Melica sp.	MELIC	
Bromus vulgaris	BRVU	Columbia brome
Dactylis glomerata	DAGL	ochardgrass
Holcus lanatus	HOLA	velvet grass
Melica bulbosas	MEBU	onion grass
Poa trivialis	POTR3	roughstalk bluegrass

FERNS

Species name	Abbreviation	Common name
Athyrium filix-femina	ATFI	lady fern
Blechnum spicant	BLSP	deer fern
Polystichum munitum	POMU	sword fern

SHRUBS

Species name	Abbreviation	Common name
Sambucus racemosa var. arborescens	SARA	red elderberry
Crataegus gouglassii var. suksdorfii	CRDO	hawthorne

Rubus laciniatus	RULA2	evergreen blackberry
Rubus spectabilis	RUSP	salmonberry
Rubus ursinus var. macropetalus	RUURM	trailing blackberry

Hydrostatic Water Depth & Temperature Probes

The Model 6508 Hydrostatic Water Depth Probes are low cost, solid state probes designed to provide accurate long term measurement of water depths from 0 to 20 metres and water temperature (optional) from 8.9 to 54°C. They are sealed and factory calibrated to standard ranges, temperature compensated and interchangeable.

The probe uses the hydrostatic pressure of water to measure water depths in a variety of environments. (The hydrostatic pressure of water is dependent on the depth at which the water pressure is measured and it is by this relationship that the probe calculates depth.)

USE - Ideal for drainage, bore hole and river height recording projects. This type of probe can also be used to measure water with dissolved solids (i.e., brine) and other liquids. Four models are available: each measures to a different maximum depth.

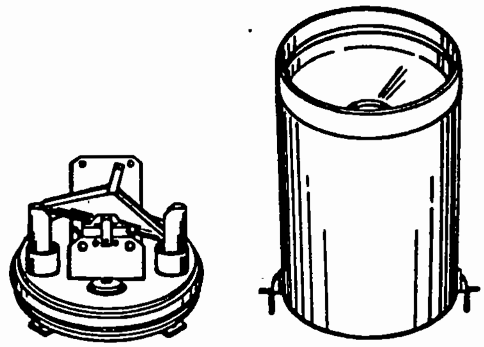
SPECIFICATIONS

Material:	High impact PVC (UV resistant) fitted with chromed brass nose cone
Size:	25mm diameter, 180mm long
Weight:	200 grams (excl. cable)
Electronics:	Integrated amplifier & correction circuitry
Sensor:	Semi-conductor strain gauge element
Accuracy:	±1% over 0-50°C
Resolution:	0.4% of range (-1 to 40°C)
Overpressure:	up to 100% of rated pressure
Power:	5V DC, 4mA (4% of battery life)
Cable:	4 core vented cable with shield Length: depth capacity plus 5m, max 50m
Connector:	25 pin 'D' type supplied
Range:	from 0 to 20 metres, varies by model
Channel Usage:	1 analog channel for depth 1 analog channel for temperature (if required)

C-1 Pressure transducer.

STARLOG PRODUCTS 1989

Model 6506A
Tipping Bucket Rainfall Gauge



The Model 6505A Tipping Bucket Rainfall Gauge collects rainfall using a funnel receiver of 203mm diameter. The rain is then strained by a metal gauze before being passed to the metallic tipping bucket measuring system. The bucket tips when a specified volume of rain is collected. A reed switch detects the tipping action and sends a signal which is counted by the Logger. This cycle continues while rain falls.

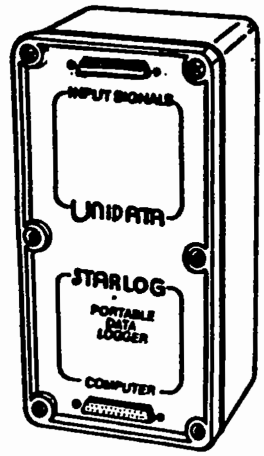
The amount of precipitation which tips the bucket can be adjusted from 0.1 to 0.5mm. It is preset to 0.2mm.

SPECIFICATIONS

Material:	Painted, polished stainless steel, cast alloy base
Funnel:	203mm diameter
Signal Element:	Sealed reed switch
Cable:	PVC 2-wire, shielded, 5 metres
Weight:	3 kg
Sensitivity:	adjustable from 0.1 to 0.5mm
Range:	0-200mm of rain per hour
Accuracy:	4% up to 200mm/hr
Channel Usage:	4-bit counter channel

C-2. Tipping bucket rain gage.

Model 6003B **Portable Data Logger**



Model 6003B Portable Data Logger

SPECIFICATIONS

Material:	Grey, high impact PVC
Dimensions:	211mm H x 108mm W x 81mm D
Weight:	800 grams excluding battery pack
Operating Temp.:	-20°C to 60°C, humidity immune
Scan Rate:	from 1/4 to 15 seconds
Log Interval:	from 1 s to 168 hrs (7 days)
Memory:	Low power CMOS, 32 or 64 Kbytes
Buffers:	up to 16, circular or linear
Available Voltage:	5V DC regulated at a maximum of 50mA to power instruments, output pulse length up to 250ms
Program Runtime:	12 to 250 milliseconds
Conversion:	first conversion 10 ms after powerup 8-bit successive approximation
Computer:	Serial, half duplex, RS-232 TTL levels, 300/1200/9600 baud 8 Data + 1 Stop bit
Connectors:	25 pin 'D' type plug and socket
Processor:	8 bit single chip MPU type 8748
Clock:	Battery powered, crystal regulated Accuracy ±10 seconds/month Date/Time Calender Year/Month/Day/Hour/Minute/Second

Channel Specifications

Analog:	8 inputs, 8-bit resolution 0 to 2.55V Input impedance (logger active): >1M Ohm Load impedance (logger inactive)	
	for signals <500mV:	>1M Ohm
	for signals >500mV:	10k Ohm
	Source drive impedance:*	<10k Ohm
	10mV per bit resolution	
Counters:	4 4-bit or 2 8-bit resolution pulse inputs 100k Ohm pull up to 5V suitable for potential free contacts 4-bit counter: counts up to 16 pulses per scan 8-bit counter: counts up to 3825 pulses per scan with prescale	
Serial:	1 input only, 1 bi-directional up to 16 channels, 2 bytes each	

*recommended

C-3. Data logger.

APPENDIX D. SATURATED HYDRAULIC CONDUCTIVITY VALUES (K) FOR THE DEER CREEK STUDY SITE.

PIEZOMETER NUMBER	TIME LAG (sec)	F(m=7.3) (m)	K(m=7.3) (m/s)
A1	26700	13.38	1.4E-07
A2	6300000	13.48	6.0E-10
A3	1900000	13.48	2.0E-09
A4	39000	13.48	9.6E-08
A5	4100000	13.48	9.2E-10
A6	22000	13.48	1.7E-07
B1	243000	13.48	1.5E-08
B2	11400	13.48	3.3E-07
B3	624000	13.48	6.0E-09
B4	1550000	13.48	2.4E-09
B5	1540000	13.48	2.4E-09
C1	2200000	13.48	1.7E-09
C2	173	13.48	2.2E-05
C3	72000	13.48	5.2E-08
C4	52000	13.48	7.2E-08
C5	6500000	13.48	5.8E-10
C6	4600	13.48	8.2E-07

Source: Salminen, 1990.