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THROUGHFLOW VARIABILITY ON A FORESTED HILLSLOPE
UNDERLAIN BY COMPACTED GLACIAL TILL

by

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B.Sc. Simon Fraser University, 1993

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ABSTRACT

This research examined the spatial variability of throughflow in a shallow forested soil in terms of the controls exerted by topography, soil properties, and the shape of the water table. A 10 m wide plot was instrumented with nine troughs installed at a road cut, and hydraulic head was measured at 37 piezometers installed to the compacted till surface immediately upslope of the road cut. Detailed surveys of surface and subsurface (i.e., till) topography and water balance calculations provided information on contributing areas to individual troughs and the entire plot.

Contributing areas draining to individual troughs varied substantially depending on whether surface or subsurface topography, or water balance analysis was used. Contributing area estimates based on water balance calculations were generally greater than topographically-derived estimates. The higher estimates based on water balance calculations may be related to the poor survey resolution of fine-scale details of the compacted till surface or lateral shunting of subsurface flow via macropores, or a combination of the two processes.

Temporal variations in the observed spatial distribution of trough throughflow could be explained by changes in water table configuration upslope of the troughs and activation of discrete macropores which shunt water laterally across the slope. At low flows, when the water table conformed closely to the compacted till topography, the observed distribution of trough throughflow conformed best to the distribution of trough contributing areas computed from subsurface topography. At moderate to high flows,

shifts in the trough throughflow distribution due to changes in the orientation of flow tubes to individual troughs were best predicted by the Dupuit flow model. Although surface topography provided a reasonable proxy for water table configuration at moderate to high flows, lateral shunting of water by discrete macropores overwhelmed topographic control on throughflow pathways, at least at scales of 1-2 m.

Effective hydraulic conductivity (K_{eff}) at individual troughs (trough scale) and troughs 4-9 combined (plot scale) were calculated from Darcy's Law. Variation in K_{eff} decreased one order of magnitude between the trough and plot scales. Furthermore, plot scale estimates of K_{eff} compare well with previous estimates of K_{eff} made within the UBC Research Forest at similar scales, or larger.

Determination of the vertical variation in K_{eff} for individual troughs was confounded by shifting flow pathways. However, the vertical variation of K_{eff} at the plot scale was reasonably approximated by a linear relation of saturated thickness above compacted till. Deviations from the linear relation at low saturated thicknesses may be partially attributed to preferential flow via a permeable organic root mat observed along the upper surface of the compacted till.

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CHAPTER 1

INTRODUCTION

Throughflow is water that infiltrates into the soil and moves laterally downslope either as unsaturated or saturated flow (Kirkby and Chorley, 1967). The general condition favouring the generation of throughflow is a marked contrast in vertical hydraulic conductivities which promotes the development of a perched water table and flow parallel to the upper surface of the basal confining layer. The conditions required for throughflow generation are met in many temperate forested environments including Canada (Chamberlin, 1972), United States (Whipkey, 1965), New Zealand (Mosley, 1979, 1982), and Japan (Tanaka et al., 1988; Sidle et al., 1995).

In undisturbed catchments of temperate forests, throughflow is an important mechanism for stormflow generation (Cheng, 1988; Bonell, 1993). Typically, shallow (ca. 1 m), coarse-textured soils overlie thick glacial till deposits, or bedrock, of reduced permeability. Forest soils are extremely heterogeneous and contain a dense network of tree roots, which yield high effective hydraulic conductivities. The heterogeneous properties of soils, in combination with complex terrain, promote spatially variable throughflow (e.g., Nagpal and deVries, 1976; Thompson, 1993). This thesis examines the spatial variability of throughflow in an attempt to elucidate controls on its distribution.

The next section outlines the significance of throughflow variability to field and modelling studies. Section 1.2 outlines a conceptual framework for understanding

throughflow variability and reviews several models describing throughflow. Sections 1.3 and 1.4 review previous empirical studies of throughflow variability and hydraulic conductivity, respectively, which provide the basis for the thesis objectives in section 1.5.

1.1 SIGNIFICANCE OF THROUGHFLOW VARIABILITY

Numerous field studies over the last 40 years have examined the mechanisms (e.g., macropore vs. matrix flow) and chemistry (e.g., old vs. new water) of throughflow in order to resolve catchment-scale hydrologic response (e.g., Whipkey, 1965; Mosley, 1979; Sklash et al., 1986; McDonnell, 1990). Due to practical limitations, many of these studies have focused on the measurement of flow from narrow troughs, typically 1-3 m wide. Several researchers have extrapolated from these small scale measurements to estimate catchment subsurface flow with little regard for the inherent spatial variability of throughflow which may exist within the catchment (Weyman, 1970; McDonnell, 1990). Although this approach may be justified as no other information was available, it may lead to grossly incorrect estimates. Weyman (1970), for example, estimated that a mean peak throughflow discharge of 1.32 L/min per metre of streambank was required to produce the observed 720 L/min increase in stream discharge along a 540 m long reach. However, peak discharge observed from instrumented soil plots was only 0.185 L/min per metre of streambank (Weyman, 1970).

At present, little progress has occurred towards linking small scale field measurements of subsurface flow to catchment scale estimates. Anderson and Burt (1978), Mosley (1979), and Huff et al. (1982) documented the effects of convergent

hollows in generating enhanced subsurface flow contributions to stream reaches. At the catchment scale, surface topography may correctly describe throughflow pathways. However, controls on throughflow pathways at the hillslope scale may not always be described by surface topography. Field studies by Pilgrim and Huff (1978), Pilgrim et al. (1978), Woods and Rowe (1996), and Freer et al. (1997) clearly demonstrated the complexity of flow processes which may occur across apparently planar hillslope segments.

Recent studies of the spatial variability of throughflow at the hillslope scale have sought to explain throughflow variability in terms of upslope contributing areas based on either surface or subsurface (e.g., bedrock) topography (Woods and Rowe, 1996; Freer et al., 1997). However, basing analysis on static contributing areas neglects the dynamic nature of controls on the rate and distribution of flow issued from hillslopes, such as the lateral extent and thickness of the saturated wedge (Weyman, 1973; Anderson and Burt, 1978) and the effects of macropores (Beven and Germann, 1982).

One of the primary aims of research into the spatial variability of throughflow is to gain an understanding of scales which are representative of their surroundings (e.g., Woods and Rowe, 1996). Given the heterogeneous nature of throughflow movement on forested hillslopes, it is important to define spatial scales at which hydrologic behaviour may be described without detailed knowledge of flow processes occurring at smaller scales. Wood et al. (1988) and Freeze and Cherry (1979) defined this threshold spatial scale as the Representative Elementary Area (REA) and Representative Elementary Volume (REV) (Figure 1.1).

Figure 1.1 Conceptual relation between hydraulic parameter (e.g., hydraulic conductivity) and spatial scale. The REA/REV represents a threshold scale at which small scale heterogeneities are averaged out and may be considered homogeneous. Figure modified after Jenssen (1990).

1.2 MODELS OF THROUGHFLOW VARIABILITY

1.2.1 Conceptual framework for modelling saturated throughflow

Darcy's Law provides a conceptual framework for understanding saturated throughflow movement. Darcy's Law is an empirical law which describes movement of water within the unsaturated and saturated zones under steady state or transient conditions in isotropic or anisotropic media. The three-dimensional form of Darcy's Law is

$$v_x = -K_x \frac{\partial h}{\partial x} \quad (1.1a)$$

$$v_y = -K_y \frac{\partial h}{\partial y} \quad (1.1b)$$

$$v_z = -K_z \frac{\partial h}{\partial z} \quad (1.1c)$$

where v_x , v_y , and v_z represent the specific discharge (LT^{-1}) in each of the three directions; K_x , K_y , and K_z are the directional components of hydraulic conductivity (LT^{-1}); and h is total hydraulic head (L).

Although Darcy's Law provides the basic framework for most hydrologic models to describe throughflow variability, its application typically requires simplifying assumptions regarding spatial variation of hydraulic conductivity and hydraulic gradient. The following section reviews throughflow models of varying complexity and highlights model assumptions.

1.2.2 Review of throughflow models

1.2.2.1 Three-dimensional physically-based models

A physically-based description of hillslope hydrology is based on the combination of Darcy's Law and the law of conservation of mass, commonly referred to as Richard's equation:

$$\frac{\partial}{\partial x} \left(K_x \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_y \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_z \frac{\partial h}{\partial z} \right) = \frac{\partial \theta}{\partial t} \quad (1.2)$$

where $\frac{\partial \theta}{\partial t}$ is the rate of change of moisture content with time. Richard's equation (1.2)

describes throughflow in three-dimensions and allows hydraulic conductivities and hydraulic gradients to vary in space and time. Numerical solution of Richard's equation is accomplished using a finite difference or finite element scheme which represents the catchment as a discrete set of nodes with either regular (finite difference) or irregular spacing (finite element). An advantage of modelling flow using this approach is that the dynamic nature of hydraulic gradients can be accommodated. However, the solution of Richard's equation requires accurate definition of boundary conditions and collection of necessary input data and can be computationally demanding. Most three-dimensional models have been developed to solve flow systems on ideal or theoretical hillslopes (e.g., Binley et al., 1989).

1.2.2.2 Two-dimensional physically based models

Richard's equation can be simplified to two horizontal dimensions by assuming Dupuit-

Forchheimer flow (a.k.a. 'Dupuit flow'). Dupuit flow assumes that flow is parallel to the impermeable layer and the hydraulic gradient is equal to the slope of the water table throughout the depth of the saturated zone.

Jackson and Cundy (1992) used a two-dimensional finite difference scheme based on Dupuit flow to model saturated zone thickness in a first-order catchment in western Oregon. Although their model predicted observed piezometric response reasonably well, it was computationally-intensive, requiring a 10 s time step for numerical stability. Furthermore, no direct measurements of throughflow were made to assess the model's predictions of the spatial variability of throughflow.

Modelling the spatial variability of hydraulic conductivity for input into two-dimensional models may not be feasible for many practical applications. Numerical studies often assume hydraulic conductivity is homogeneous and isotropic, typically using a depth-integrated value of hydraulic conductivity based on soil cores (e.g., Jackson and Cundy, 1992).

1.2.2.3 One-dimensional topographically-driven flow

Richard's equation may be simplified to one dimension by further assuming that topography controls throughflow pathways. The assumption of topographically-driven flow treats saturated throughflow as following streamlines which cross elevation contours at right angles (Hillel and Hornberger, 1979; Fayer and Hillel, 1982). Lateral flow (i.e., perpendicular to fall line) is assumed to be zero. These models discretize the catchment into a set of flow tubes defined by neighbouring streamlines, which do not vary through

time. The flow tubes are further discretized into a series of elements representing position in the soil profile and position on the slope profile, at which vertical recharge or evaporation and lateral saturated flow can be estimated. Early models assumed a constant flow tube width over the length of the slope profile. As a result, many of the early topographically-driven models neglected flow convergence in complex topography (Hillel and Hornberger, 1979). More sophisticated models have been developed to incorporate topographic convergence by assuming a variable flow tube width (e.g., Prevost et al., 1990; Anderson and Kemp, 1991).

Hydraulic conductivity can be specified explicitly within each element. However, most studies assume hydraulic conductivity is homogeneous and isotropic for simplification and for lack of appropriate field data (Fayer and Hillel, 1982; Prevost et al., 1990).

1.2.2.4 Quasi-steady state topographically-driven models

Throughflow models can be simplified further by invoking the assumption that saturated throughflow is in quasi-steady state with recharge to the upslope surface contributing area. That is,

$$q_i = ra_i \quad (1.3)$$

where q_i is the saturated throughflow discharge from contour segment i [L^3T^{-1}]; r is the spatially uniform recharge rate [LT^{-1}]; and a_i is the upslope contributing area draining to contour segment i [L^2]. By assuming that the water table is parallel to the soil surface, downslope flow per unit contour length can be calculated as

$$q_i = T_i \tan\beta \quad (1.4)$$

where T_i is the local soil transmissivity of the saturated zone [L^2T^{-1}], and $\tan\beta$ is the local surface slope. Local soil transmissivity is defined as

$$T_i = \int_{-\infty}^0 K(z) dz \quad (1.5)$$

where z is the depth to the water table relative to the local ground surface [L], and $K(z)$ is the value of hydraulic conductivity at depth z [LT^{-1}].

Equations (1.3) and (1.4) provide the basis for the models TOPMODEL (Beven and Kirkby, 1979) and TOPOG (O'Loughlin, 1986). Transmissivity is usually assumed to be spatially homogeneous. Therefore, flow pathways are derived from the geometrical features of the catchment (i.e., DEMs of surface topography). The relative simplicity of TOPMODEL and TOPOG has led them to become popular means to model flow pathways in catchments.

In the original version of TOPMODEL (Beven and Kirkby, 1979), the vertical hydraulic conductivity profile was modelled as an exponential function of saturated depth

$$K(z_i) = K_o e^{-fz_i} \quad (1.5)$$

where z_i is the depth of the water table relative to the local ground surface, $K(z_i)$ is the value of hydraulic conductivity at z_i , K_o is the hydraulic conductivity at the soil surface, and f is a parameter describing the rate of decline of hydraulic conductivity with depth [L^{-1}]. More recently, Ambroise et al. (1996) suggested that the exponential form previously assumed may not be appropriate for predicting base flow recession in forested catchments. They proposed the following linear and constant forms to describe the

vertical hydraulic conductivity profile:

$$K(z_i) = K_o \left(1 - \frac{z_i}{D_i} \right) \quad (1.6)$$

$$K(z_i) = K_o \quad (1.7)$$

where D_i is the local soil depth above an impermeable layer. An important implication for modelling throughflow using equations 1.6 and 1.7 is that hydraulic conductivity is assumed to equal zero at or below the base of the soil profile.

1.3 EMPIRICAL STUDIES OF THROUGHFLOW VARIABILITY

Quasi-steady state, topographically-driven models (e.g., TOPMODEL, TOPOG) have become popular means to model water and solute movement in forested catchments due to their relatively simple data requirements. However, there has been relatively little field research to support the model assumptions. Previous field studies by Anderson and Burt (1978) and Huff et al. (1982) documented the enhancement of subsurface flow inputs to stream reaches from convergent hollows. However, studies of throughflow variability in Maimai catchment, New Zealand (Woods and Rowe, 1996) and Panola Mountain Research Watershed (McDonnell et al., 1996; Freer et al., 1997) suggest that surface topography may not accurately predict the spatial variation of throughflow at the hillslope scale.

Woods and Rowe (1996) monitored the variability of throughflow over a 110.5 day period on a forested hillslope in Maimai catchment using thirty throughflow troughs (1.7 m long). Trough outflow generally corresponded with upslope surface contributing

area. However, the observed flow at several troughs was less than expected based on surface topography. McDonnell (1997) attempted to explain the spatial variability of flow at the Maimai hillslope in terms of bedrock (i.e., subsurface) contributing areas since saturation of the soil profile occurs from the bedrock surface up into the soil horizons. Although slight differences in trough contributing areas between surface and bedrock topographies were observed, detailed knowledge of the bedrock surface did not improve estimation of the flow distribution across the hillslope. Freer et al. (1997) suggested that at Maimai, where bedrock and surface topographies are highly correlated, it is reasonable to assume that hydraulic gradients are parallel to surface topographic gradients at the hillslope scale.

Although both surface and bedrock topographies explained the spatial distribution of throughflow across the Maimai hillslope to some degree, the spatial distribution varied with total trough flow. The observed temporal variation in throughflow distribution contradicts the quasi-steady state assumption underlying TOPMODEL and TOPOG (equation 1.3). Woods et al. (1997) attempted to explain the observed throughflow variability at Maimai under the paradigm of topographically-driven flow by assuming throughflow variability was related to spatial variations in recharge rate.

McDonnell et al. (1996) and Freer et al. (1997) instrumented a hillslope in Panola Mountain Research Watershed with ten 2 m long troughs. They observed that the spatial distribution of throughflow across the hillslope was better correlated with the bedrock topographic index than surface topographic index (Freer et al., 1997). Therefore, Freer et al. (1997) suggested that the bedrock surface may be more appropriate to describe

throughflow variability on hillslopes where the surface and bedrock topographies are poorly correlated.

Although both the Maimai and Panola studies have provided valuable information on the observed throughflow variability which may exist across a hillslope, these studies did not incorporate measurements of upslope saturated wedge to explain the observed spatial variability. Several researchers have noted that the configuration of the saturated wedge is dynamic, controlling the rate and distribution of flow issued from a hillside (Weyman, 1973; Anderson and Burt, 1978; Burt and Butcher, 1985). Therefore, it may be incorrect to assume that the direction of throughflow remains perpendicular to topographic contours over all time periods. Hinton et al. (1993) observed that contributing areas based on surface topography differed by as much as 57% from contributing areas based on groundwater divides in a small forested catchment in central Ontario.

Variations in water table configuration imply that contributing areas draining narrow contour segments (e.g., troughs or soil pits) are dynamic. Attempts to delineate contributing areas on the basis of either surface or subsurface topographies (e.g., Peters et al., 1995; Woods and Rowe, 1996; Freer et al., 1997) may not truly represent the area draining to a contour segment at any given time period. Furthermore, comparison of runoff per unit contributing area between hillslope and catchment scales (e.g., Sidle et al., 1995) may be uncertain unless the dynamics of contributing area to the hillslope can be defined. A water balance approach may provide one method to estimate an effective contributing area draining narrow contour segments. Troendle and Reuss (1997) used the

water balance approach to estimate contributing areas draining two contour segments of 12 m and 36 m long. However, this technique does not provide any information on the spatial location of contributing areas.

The dynamic nature of the saturated wedge suggests that neither surface nor subsurface topography may completely describe throughflow pathways at the hillslope scale. Instead, topographic controls on throughflow pathways likely depend on local saturated zone thickness. Where soils are nearly saturated to the surface, surface topography is a good approximation to local hydraulic gradients (Dunne et al., 1975). However, where (or when) the saturated thickness overlying the basal confining layer is relatively thin, subsurface topography may be more appropriate to estimate hydraulic gradients. The dynamic nature of the saturated wedge may therefore have a significant influence on the spatial pattern of throughflow, with bedrock and surface topography representing two end-members in describing flow pathways. At present, no studies have examined the dynamic configuration of the saturated wedge with respect to the two end-members.

Soil heterogeneity may also have a significant influence on the observed throughflow variability across a hillslope. Freer et al. (1997) found that macropores contributed approximately 40% of the total runoff from the Panola hillslope. At Maimai, Woods and Rowe (1996) observed a high degree of soil heterogeneity at trough faces (pipe flow observed at nine of the thirty troughs); however, they failed to link these descriptions with observations of trough throughflow variability. Given the high degree of soil heterogeneity observed at trough faces by Woods and Rowe, it is feasible that soil

heterogeneity may over-ride topographic control of throughflow pathways, at scales less than about 10 m. Flow via macropores may have led to the poor relations between throughflow and upslope contributing area observed by Woods and Rowe (1996).

1.4 EMPIRICAL STUDIES OF HYDRAULIC CONDUCTIVITY IN FORESTED SOILS

Considerable research has been conducted in the past three decades to document the vertical and lateral heterogeneity of hydraulic conductivity in forest soils (Harr, 1977; Elsenbeer et al., 1992; Buttle and House, 1997). Vertical soil heterogeneity is typically ascribed to reduced biological, physical, and chemical weathering with depth, or the occurrence of a basal confining layer such as bedrock, glacial till, or hardpan. Lateral soil heterogeneity has often been overlooked even though several field observations of macropore flow in forested environments have documented its existence (Nagpal and deVries, 1976; Mosley, 1979, 1982; Beven and Germann, 1982). Megahan and Clayton (1983) used salt tracers to document throughflow pathways in forest soils of the Idaho batholith. Not only was the spatial distribution of tracer highly localized, but the point of first detection did not correspond to the most direct route based on surface topography.

Since the Darcian approach to describing water flow through porous media replaces individual grains and pores with a representative continuum, it is necessary that soil samples are large enough that small-scale heterogeneities (e.g., macropores) are averaged out. The scale at which small scale heterogeneities are averaged out is often referred to as the representative elementary volume (REV). Bouma (1983) suggested that

an appropriate REV for a clay soil with continuous macropores was approximately 10^5 cm^3 . However, typical methods of measuring hydraulic conductivity (e.g., slug test, cylinder infiltrometer) sample soil volumes of approximately 10^2 to 10^6 cm^3 (Jenssen, 1990). The effect of scale on hydraulic conductivity estimates has been rarely examined. However, Megahan and Clayton (1983) and Utting (1979) noted that estimates based on tracer injection or plot outflow measurements can be up to two orders of magnitude greater than laboratory-derived estimates.

The discrepancy between laboratory and plot scale estimates of hydraulic conductivity may also be due to the preferred orientation of root channels downslope (Eis, 1974; Noguchi et al., 1997). Furthermore, several studies have observed a permeable root mat or weathered zone along the upper surface of the basal confining layer (deVries and Chow, 1978; Utting, 1979; Tischer, 1986; Wels et al., 1991; Peters et al., 1995). These preferred pathways oriented downslope may not be incorporated into laboratory-derived estimates of hydraulic conductivity which typically measure the vertical (e.g., cylinder infiltrometer) or a combination of vertical and horizontal components (e.g., auger hole method, slug tests). As a result, many researchers have likely underestimated lateral downslope fluxes using values of hydraulic conductivity based on vertical infiltration techniques (Harr, 1977; Anderson and Burt, 1978; McDonnell, 1990).

1.5 RESEARCH OBJECTIVES

The previous literature review has highlighted the lack of field studies examining the spatial variability of throughflow, especially in relation to the configuration of the saturated wedge, and measurement of hydraulic conductivity in forested catchments. The following four research objectives will be addressed in this thesis:

- (1) To compare contributing areas draining narrow contour segments based on topographic analysis and water balance calculations
- (2) To examine the spatial variability of throughflow in terms of water table configuration and soil anisotropy
- (3) To evaluate the ability of commonly-used modelling assumptions to predict throughflow variability
- (4) To investigate the variability of effective hydraulic conductivity, particularly in relation to spatial scale and its vertical variation within the soil profile.

Four chapters comprise the remainder of this thesis. Chapter 2 provides details of methodology and data analysis. Chapter 3 presents results, and Chapter 4 provides interpretation and discussion of results pertaining to the research objectives. Chapter 5 summarizes the main conclusions and makes recommendations for future research.

CHAPTER 2

METHODS

This chapter describes the biophysical characteristics of the University of British Columbia Research Forest and the experimental hillslope in detail. Details on site instrumentation, methods of data collection, sampling design, and data analysis are subsequently discussed. The chapter concludes with an examination of the study period to put the study into a temporal context.

2.1 STUDY AREA

2.1.1 Location and Physiography

The study was carried out in the University of British Columbia Research Forest, located approximately 50 km east of Vancouver, near the town of Maple Ridge (Figure 2.1). The Research Forest is situated along the southern flanks of the Coast Mountain Range. The southern half of the Forest consists of gently rolling terrain with a few hilly granitic-cored uplands. Terrain through the northern half is mountainous, commonly accented by steep-flanked granitic knolls and bedrock outcrops (Klinka and Krajina, 1986). Elevations range from a few metres above sea level near Pitt Lake to slightly over 1,000 m along the eastern border of the Research Forest. Figure 2.2 indicates locations of the present study and previous hillslope hydrological studies carried out in the Research Forest.

Figure 2.1 Location of UBC Research Forest and proximity to Vancouver and Coast Mountains. Shaded portion of Coast Mountains represents elevations of 450 m or higher.

Figure 2.2 Detailed map of UBC Research Forest indicating locations of Research Forest headquarters and study site. Circled numbers indicate locations of previous hillslope hydrological studies completed in Research Forest.

2.1.2 Climate

The climate of the Research Forest is classified as Pacific Marine Humid. Monthly mean daily temperatures, based on 1961-90 climate normals measured at the Atmospheric Environment Service (AES) climate station located at the Research Forest headquarters, range from 1.9°C in January to 17°C in August (Environment Canada, 1991). Total annual precipitation is 2184 mm, of which the bulk (75%) is produced during the winter months (October to April) by frontal and cyclonic storms (Figure 2.3). Owing to the relatively low elevation of the study area, in combination with the moderating effects of the Pacific Ocean, snowfall comprises less than 5% of the total annual precipitation (Figure 2.3).

Figure 2.3 Climate normals (1961-1990) for mean daily temperature and precipitation at the UBC Research Forest headquarters.

2.1.3 Geology

Bedrock geology of the Research Forest has been mapped primarily as quartz diorite, diorite, and granodiorite belonging to the Coast Crystalline Complex (Roddick, 1966). Most exposed bedrock is generally unweathered due to rapid uplift and recent glaciation in the Coast Mountain Range, and can be extremely variable in its degree of fracturing. Utting (1979) noted that the fracturing was oriented randomly, with fracture apertures of 2-3 cm and spacing of roughly 10 m.

Surficial geology of the Research Forest is primarily glacial in origin, deposited during the Wisconsin Glaciation which occurred approximately 10,500 to 26,000 years BP (Hicock and Armstrong, 1981). Vashon till, deposited during the maximum extent of the Cordilleran ice sheet during the Fraser Glaciation (ca. 14,500 years BP) is extensive throughout most of the Research Forest (Hicock and Armstrong, 1985). Vashon till is extremely compact in many areas, possibly suggesting that it is a basal till. Colluvium is common in the mountainous northern half of the Research Forest, while glaciofluvial and glaciolacustrine sediments are common at low elevations in the southern half of the Research Forest.

2.1.4 Pedology

Soils have been primarily mapped as coarse-textured humic-ferric podzols (Klinka, 1976). Soil depths range up to 2 m, with compacted till or granitic bedrock found at an average depth of 1 m (Utting, 1979). Soils are typically sandy-loam to loamy-sand in texture and have a high coarse fragment content. Saturated hydraulic conductivities are

typically 10^{-4} to 10^{-3} m/s in the soil, based on pump test and vertical infiltration methods (Cheng, 1988). Hydraulic conductivities in the underlying compacted till are approximately 10^{-7} m/s, based on slug tests conducted by Utting (1979).

2.1.5 Hydrology

Peak stormflow from forested catchments within the Research Forest is generated from long-duration cyclonic and frontal storms occurring in the winter months (October to April) (Cheng, 1988). Annual runoff is approximately 60% of annual precipitation (Feller and Kimmins, 1979).

Rainfall intensities rarely exceed the infiltrability of forest soils (Cheng, 1988). Hortonian overland flow is rarely observed except on severely disturbed sites, such as forest roads. In undisturbed forested environments, runoff pathways are primarily via subsurface stormflow and saturation overland flow. The heterogeneous nature of forest soils causes a significant portion of stormflow to follow high conductivity pathways, effectively by-passing the soil matrix. Rapid, concentrated flows from macropores have been observed at several locations within the Research Forest (Nagpal and deVries, 1976; Feller and Kimmins, 1979; Utting, 1979; Tischer, 1986). Water table response on hillslopes is generally rapid, fluctuating up to 70 cm over a single storm (Thompson, 1993).

2.1.6 Vegetation

The Research Forest is located within the Coastal Western Hemlock biogeoclimatic zone (Klinka and Krajina, 1986). Climax vegetation is Western Hemlock (*Tsuga*

heterophylla), but Douglas-fir (*Pseudotsuga menziesii*), and Western Red Cedar (*Thuja plicata*) are also common. Stand history throughout the Research Forest is complex due to forest fires and extensive clear-cutting which occurred during the late 1800s and early 1900s (Klinka and Krajina, 1986).

2.2 FIELD SITE

2.2.1 Location and physiography

The research site is a small forested hillslope, approximately 10 m wide by 60 m long, located approximately 1.5 km east of the Research Forest headquarters (Figure 2.2; Figure 2.4). The site is accessible by vehicle from forest road A40, which contours across the base of the hillslope (Figure 2.4). Soils, vegetation, and geologic materials present at the site are representative of physical and biological conditions occurring on similar hillslopes in southwestern British Columbia. As a result, hydrologic processes characterized at the research site should be representative of similar-sized hillslopes in southwestern British Columbia.

The study site is generally southeast facing with an average slope of approximately 30% (Figure 2.4). Elevations range from approximately 230 to 250 masl. Surface topography across the study site is relatively uniform with no major regions of convergence or divergence. However, microtopography (i.e., scales ranging from 1 to 5 m) is complex due to historic tree-throw occurring on site (Figure 2.5).

The study site was sub-divided into two main units: “plot” represents a small

Figure 2.4 Overview map of research site and field instrumentation. Approximate surface contributing area to plot and hillslope illustrated by dotted and dashed lines respectively. Contour interval is 0.5 m.

hillslope segment which drains to a series of troughs, whereas “hillslope” refers to a larger hillslope segment draining to a nearby road culvert (Figure 2.4). This designation will remain consistent throughout the remaining chapters of this thesis.

2.2.2 Pedology

Soils may be classified as both humic-ferric podzols and orthic-dystric Brunisols, depending on stratification of the B horizon. The forest floor is comprised of an extremely permeable layer of needles, twigs, and logs in various states of decay. A thick litter layer, often up to 0.3 m, surrounds stumps of old-growth trees felled in the 1920s. Underlying the litter layer is a discontinuous Ae horizon, approximately 0.05 m thick.

The B horizon is typically 0.3 to 0.9 m thick and has a high coarse fragment content (up to 50%). Soils are slightly acidic (pH = 5.5 to 6.0) and are dark yellowish-brown (10YR 4/4) in colour when dry. Roots are present throughout the entire B horizon with a slight decrease in root densities with depth.

A thin (ca. 1-2 cm), permeable root mat is often observed along the upper surface of the compacted till layer where the transition from the soil to the unweathered compacted till is abrupt (Utting, 1979). The compacted till layer apparently causes downward-migrating roots to splay out laterally along the upper surface of the compacted till, providing a connected pathway for subsurface flow downslope. The root mat layer remains moist through much of the winter, drying out only during extended periods of no precipitation. A permeable root mat has been observed at other research sites in southwestern British Columbia where soils are underlain by compact till

Figure 2.5 Typical forest floor microtopography due to historic tree-throw. Flagging tape attached to sticks in foreground represent locations of auger holes dug to survey topography of compacted till (section 2.4.2).

(Chamberlin, 1972; deVries and Chow, 1978; Utting, 1979; Tischer, 1986). The spatial extent of the root mat is unknown since most observations were conducted along road cut segments. Displaced organic fragments from auger holes drilled on site (section 2.4.2) confirmed the presence of a root mat at the base of the soil profile. These observations suggest that the root mat may extend up to 20 to 40 m upslope of the road cut at the study site.

2.2.3 Plot geology

Soils are underlain primarily by compact Vashon till. However, granitic bedrock also underlies soils, especially in the upper sections of the hillslope. Topography of the compacted till layer does not necessarily follow surface topography (Figure 2.6).

The till is heterogeneous with physical properties of colour and degree of compaction varying over the scale of metres. Unweathered till has a light grey (5Y 7/1) colour when dry and is extremely compact. Weathered till is typically brownish-yellow (10YR 4/4) in colour when dry and is either fractured into compact, weathered fragments or is relatively soft and massive in nature.

Approximate thickness of the compacted till unit is not precisely known. Based on observations of granitic outcrops immediately east of the study site along forest road A40, it appears that the till thickness may range between 1 and 3 m. This observation is consistent with Utting's (1979) findings at a nearby site within the Research Forest (Figure 2.2).

Figure 2.6 Photograph illustrating plot face at road cut. Note differences in surface and compacted till topographies.

2.2.4 Vegetation

The study site is vegetated by a relatively homogeneous stand of Douglas fir (*Pseudotsuga menziesii*), Western Hemlock (*Tsuga heterophylla*), and Western red cedar (*Thuja plicata*) approximately 60 to 70 years in age. The stand regenerated naturally from a forest fire which occurred in 1933. Remnant stumps from logging activities which occurred in the 1920s are present throughout the understory. Understory vegetation is sparse, but consists of huckleberry (*Vaccinium parvifolium*), vine maple (*Acer circinatum*), and swordfern (*Polystichum munitum*).

2.3 FIELD MEASUREMENTS

2.3.1 Plot throughflow

Throughflow issuing from the plot was measured by a series of nine concrete troughs installed along the exposed road cut. The troughs sampled throughflow continuously across the plot, except for a small gap (ca. 1 m) between troughs 1 and 2 (Figure 2.7). A 0.5 m diameter boulder within the till made it impossible to construct a trough in this segment. Discharge from troughs 1 and 2 drained westward to an unnamed tributary, while troughs 3 to 9 drained eastward to a nearby culvert draining the hillslope (see section 2.3.2).

Troughs were installed immediately below the soil-till contact exposed along the road cut. A small ledge 0.05 to 0.1 m wide was chiseled into the compacted till to serve as a base on which concrete was poured. The heterogeneous nature of the till made it

Figure 2.7 Detailed map of field instrumentation. Arrows indicate directions of flow in drainage ditch (shaded).
Contour interval 0.5 metres.

impossible to construct identical troughs across the road cut face. Troughs ranged in width from 0.97 m (trough 7) to 1.57 m (trough 2) (Table 2.1). A plastic funnel set into the concrete at the lowest point along the trough concentrated flow to a single point for measurement (Figure 2.8). Troughs were designed to accommodate flows up to 200 mL/s. During peak stormflow periods, however, throughflow discharge at trough 1 exceeded 200 mL/s due to the activation of a macropore (ca. 2-3 cm diameter) located within the soil profile. To accurately measure discharge at this trough, flow from the macropore was shunted past the trough using a beveled cylinder (7 cm diameter) pressed into the soil column surrounding the macropore. An aluminum tray mounted beneath the beveled cylinder shunted the flow past the trough and was measured separately (Figure 2.9). During all measurement periods, tarps were erected over the road cut to prevent rain from entering the troughs (Figure 2.10).

Throughflow discharge at each trough was measured manually with a graduated cylinder and stopwatch. The size of graduated cylinder used to measure discharge depended on a visual approximation of the flow rate at each trough. Given a maximum possible error of ± 0.2 s in operating the stopwatch and measurement errors using graduated cylinders (± 0.25 mL for 25 mL graduated cylinder to ± 2.5 mL for 500 mL graduated cylinder), maximum error in discharge measurement is less than $\pm 10\%$. A typical measurement error for trough discharge was approximately ± 3 to $\pm 6\%$.

Table 2.1 Summary of trough and plot dimensions

	T1	T2	T3	T4	T5	T6	T7	T8	T9	Plot
Width (m)	1.14	1.57	1.14	1.36	1.01	1.18	0.97	0.99	1.16	10.52

Figure 2.8 Photograph of trough 8 showing details of trough construction. Plastic funnel set into concrete concentrated flow to a single point for measurement. Note: the large black hole above the funnel was created by displacing a rock from the compacted till layer.

Figure 2.9 Photograph illustrating trough 1 and aluminum tray to shunt macropore flow past trough for separate measurement.

Figure 2.10 Photograph illustrating tarps erected over plot face to prevent rainfall from directly entering troughs during periods of throughflow measurement.

2.3.2 Hillslope throughflow

Discharge at a nearby road culvert (0.4 m diameter) was monitored to provide estimates of throughflow issued by the hillslope (Figure 2.7). The contributing areas of troughs 3 to 9 are nested within the hillslope as they drain to the road culvert (Figure 2.7). In addition to the hillslope, the culvert drains a 50 m long road segment (ca. 500 m² in area) which generates Hortonian overland flow during moderate rainfall intensities.

Contributing area of the hillslope, excluding the road segment, is estimated to be approximately 2000 m² based on surface topographic surveys. Although the subsurface contributing area of the hillslope is unknown, it is assumed to be similar to estimates based on surface topography. This assumption was based on the fact that the eastern portion of the hillslope is bounded by bedrock ridge which extends to the forest road (Figure 2.4). The western hillslope divide, defined from surface and subsurface topographic surveys (section 2.4), was reasonably similar for both topographies.

Discharge was measured manually at the culvert outlet using a graduated 4 L plastic bucket and a stopwatch. During periods of high culvert discharge (> 2000 mL/s), a 20 L plastic bucket was used to capture culvert flow. Given maximum possible errors in recording time (± 0.2 s) and errors in volume measurements (± 100 mL), maximum errors in culvert discharge are approximately $\pm 20\%$.

2.3.3 Hydraulic head

Hydraulic head was monitored across the lower 10 m of the plot by 37 piezometers installed to the soil-till contact (Figure 2.7). Average spacing of piezometers is

approximately 2–4 m. Piezometers were installed over a greater width than the contour length sampled by the throughflow troughs to ensure that flow tubes to each trough could be adequately defined.

Piezometers were constructed of 1.5 cm I.D. PVC tubing. Holes (5 mm diameter) were drilled in the lower 10 to 20 cm of the pipe, and screened using nylon mesh. Installation was in a 50 mm diameter hand-augered hole dug to the soil-till contact. Coarse sand ($K_{\text{sat}} \cong 2 \times 10^{-3} \text{ ms}^{-1}$) was packed around the screened area and a layer of bentonite clay or concrete slurry packed around the piezometer above the coarse sand pack sealed the piezometer intake. The remainder of the augered hole was backfilled with excavated soil.

A water table well constructed of 4 cm I.D. ABS tubing was installed near the base of the plot and instrumented with a pressure transducer. The pressure transducer was connected to a Campbell Scientific CR21x data logger which recorded average water table levels every 10 minutes.

A field survey conducted in April, 1997, using a total station provided both vertical and horizontal control on the positions of piezometers. Ground elevations adjacent to the piezometers on the eastern side along the same local contour were surveyed. The length of piezometer above ground was added to the surveyed ground elevation to obtain the elevation to the top rim of the piezometer.

Depth to the free water surface in each piezometer was measured from the top of the piezometer using a ranging rod, which consisted of a length of 0.6 cm I.D. acrylic tubing with a metric tape mounted to the outside of the tube. The water level in each

piezometer was determined by blowing air into the acrylic tube while lowering it into the piezometer and listening for bubbles. Depths were recorded to the nearest millimetre. However, true depths using this method are probably accurate to within ± 5 mm.

If the throughflow system can be approximated by Dupuit-Forchheimer flow (section 1.2.2.2), then hydraulic head measured by wells or piezometers may underestimate the true water table elevation (Figure 2.11). Errors between “measured” and “true” local water table elevation vary depending on the length of piezometer intake, total saturated thickness, and slope of impermeable layer. A summary of possible errors in saturated thickness are provided in Table 2.2 for piezometers with 10 cm and 20 cm screened intakes and the water table well. In general, the difference between measured and predicted is small ($< 8\%$ of true saturated thickness). The influence of this error on the predicted water table configuration is discussed further in section 3.4.1.

Table 2.2 Predicted difference between “measured” and “true” depth of saturated zone above till assuming Dupuit flow⁺

Saturated thickness normal to till (cm)		Piezometer (10 cm intake)	Piezometer (20 cm intake)	Well
10	True	10.44	10.44	10.44
	Measured	9.99	10.01	10.01
	Difference	0.45	0.43	0.43
20	True	20.88	20.88	20.88
	Measured	19.57	19.98	20.02
	Difference	1.31	0.90	0.46
60	True	62.64	62.64	62.64
	Measured	57.88	58.30	60.06
	Difference	4.76	4.34	2.58

⁺ Assuming slope of the basal confining layer is 30%.

Figure 2.11 Sketch illustrating difference between "measured" and "true" water table elevations at piezometer locations.

2.3.4 Precipitation

A tipping bucket rain gauge was installed in a nearby clearing to record the timing of precipitation inputs to the study hillslope (Figure 2.7). The tipping bucket was connected to the CR21x data logger which recorded precipitation totals every 10 minutes.

Daily precipitation recorded at the study site was highly correlated with daily precipitation recorded at the AES station located at the Research Forest headquarters, approximately 1.5 km away ($R^2 = 0.97$, $n = 55$). Linear regression between the two stations gave an equation of $y = 0.87x - 0.37$, with the study site data as the dependent variable. The high degree of correlation between the two precipitation gauges implies that return periods measured at the AES station should be similar to those expected at the research plot. Maximum rainfall intensity-duration-frequencies of storms measured at the AES station during the study period are summarized in section 2.7.

2.3.5 Sampling design

Since only a single water table well and the tipping bucket rain gauge were instrumented for continuous monitoring of hillslope throughflow conditions, a defined procedure for measuring field instrumentation was established. Discharges at the culvert outlet and nine troughs were measured prior to measuring water levels in piezometers. After piezometers were measured, discharges at the culvert outlet and trough locations were re-measured. Changes in trough and plot discharges before and after piezometer measurements provided a means to assess whether near-steady-state or transient flow conditions were occurring. A complete set of discharge measurements (nine troughs plus culvert outlet)

took approximately 10 to 15 minutes to complete, while a complete set of piezometer measurements took approximately 20 to 40 minutes to complete. During periods of intensive monitoring of hillslope throughflow conditions, measurements occurred approximately every 2 hours for periods of up to 36 hours.

2.4 FIELD SURVEYS

2.4.1 Surface topography

Surface topography was surveyed using a total station. A total of 916 points were surveyed, representing an average density of 0.10 m^{-2} (Figure 2.12). Survey points extended slightly beyond the approximate catchment boundary of the hillslope to adequately define surface contributing areas to the plot and hillslope. Based on repeated measurements on known points from different instrument locations, average errors are less than $\pm 10 \text{ cm}$ for horizontal position and $\pm 1 \text{ cm}$ for vertical position.

Isolated topographic highs (e.g., glacial erratics) or lows (e.g., pits created by tree-throw; Figure 2.5) were not believed to play a major role in governing throughflow pathways. Rather than accentuate the topographic variability, care was taken to choose survey points such that they smoothed topographic anomalies.

2.4.2 Subsurface topography

A detailed survey of subsurface topography was carried out only for the contributing area to the plot since it was not feasible to conduct a detailed survey of subsurface topography

Figure 2.12 Location of survey data points for surface topography. Contours of surface topography provided as background. Contour interval is 1 m.

over the larger hillslope segment. Subsurface topography was defined by augering down to the impermeable horizon (i.e. compacted till or bedrock). A total of 206 auger holes was dug across the plot, representing an average survey point density of 0.09 m^{-2} (Figure 2.13). A further 28 survey points were gathered from a survey of the soil-till contact along the exposed road cut at the base of the hillslope. Soil depths ranged from 28 to 173 cm, with an average of 102 cm.

Auger holes were drilled using a combination of hand and power augering techniques. Initially, a two-person Groundhog power auger mounted with a 7 cm diameter bit was used to dig the majority of each hole. A hand auger (5 cm diameter bit) was then used to finish the hole and determine the depth to the impermeable horizon. However, digging using the power auger was found to be, at times, as difficult as hand augering. Subsequently, most of the auger holes were dug using the hand auger.

Once the depth to impermeable horizon was determined by augering, a piece of flagging tape was tied around the auger shaft at the soil surface (i.e., top of the litter layer) and the auger was removed from the hole. The depth was measured from the flagging tape to the end of the auger using a metric carpenter's tape. Following completion of the ground survey, a total station was used to survey points adjacent to each auger hole to provide horizontal and vertical control. The depth to the impermeable horizon was subtracted from the surface elevation surveyed along the same local contour adjacent to the auger hole.

Determining the “true” depth to the impermeable horizon depended on the material underlying the soil B horizon. Where till comprised soil parent materials, the

Figure 2.13 Location of surveyed data points for subsurface topography. Contours of surface topography provided as background. Contour interval is 1 m.

impermeable horizon was taken as the depth at which the auger was forced to refusal and compact, unweathered till was observed on the end of the auger bit.

Where soils were underlain by granitic bedrock, the depth to the impermeable horizon was taken as the maximum depth that the auger could be drilled. Since it was impossible to determine whether the observed bedrock horizon was truly the impermeable horizon or just a large boulder, several test holes were augered in the close proximity. The deepest of the test holes was used to represent the true elevation of the impermeable horizon. In this study, this method is not believed to affect subsurface contributing areas for two reasons. Firstly, bedrock underlies the soil B horizon only in the upper sections of the study hillslope. Therefore, large errors in depth to the impermeable horizon will not have a major influence on downslope contributing areas calculated for the troughs. Secondly, a compacted till horizon was observed at depth in the majority of auger holes (175 out of 206). Therefore, only a small number of holes may be in error.

Vertical error in estimating the elevation of the till surface is approximately ± 10 cm, on average, based on the difference between auger hole depth and soil pit depth dug adjacent to nine auger holes. Errors in estimating horizontal position are similar to those found for surface topography.

2.5 DATA ANALYSIS

2.5.1 Estimation of contributing areas

2.5.1.1 Estimates based on digital elevation models of topography

Trough and plot contributing areas were estimated from contour maps generated from surface and subsurface topographic surveys. Irregularly-spaced survey data were converted to a regularly-spaced grid (0.2 m spacing) and contoured at a 0.2 m interval using SURFER (Golden Software, 1995). As the choice of interpolation method affects elevation interpolated at grid nodes, contour maps were produced using three common interpolation methods: kriging (linear semivariogram), minimum curvature, and triangulation interpolation.

Flow lines were drawn on contour plots “by eye” to represent flow divides between adjacent troughs. A manual method of flow path determination was chosen over computerized methods (e.g. single- or multi-flow path algorithms) for two reasons. Firstly, research has not resolved which algorithm best simulates subsurface flow pathways on forested hillslopes. Secondly, trough divides do not necessarily coincide with grid nodes. Therefore, some subjectiveness would remain in allocating flow from a grid cell to adjacent troughs.

Four replicate contour maps for each interpolation method were produced to estimate the error associated with manually drawn flow lines. Each replicate map was offset by one-quarter of the contour interval (10 cm) to provide a slightly different representation of the hillslope. In summary, 12 contour plots (3 interpolation x 4

replicates) were used to estimate surface and subsurface contributing areas to each trough.

2.5.1.2 Estimates based on water balance calculations

Contributing areas were also estimated by computing a water balance for storms in which throughflow was intensively monitored. A simple water balance can be expressed as:

$$R_i = (P - E - \Delta S)A_{\text{eff}i} \quad (2.1)$$

where R_i is total runoff volume measured from contour segment i [L^3], P is incident precipitation [L]; E is evaporative losses [L]; ΔS is change in soil moisture storage, and $A_{\text{eff}i}$ is the effective contributing area draining contour segment i [L^2].

Since no direct measurements of interception and evaporation by the forest canopy were made during the storm periods, the net precipitation available for runoff (i.e., $P - E - \Delta S$) was assumed to be equal to the precipitation measured at the tipping bucket gauge on-site. This assumption will likely under-estimate contributing areas computed using the outflow method.

The observed runoff volume was computed from the time-integrated discharge over the storm period until post-storm plot discharge was similar to pre-storm levels. This technique minimized changes in hillslope storage. Observed throughflow discharges at troughs and plot were assumed to increase or decrease linearly between measurement periods.

2.5.2 Spatial variability of throughflow

2.5.2.1 Trough-scale throughflow proportions

Discharges measured at individual troughs were compared to the total trough discharge to examine the spatial variability of throughflow across the hillslope. Since a complete set of discharge measurements (nine troughs, plus culvert outlet) took approximately 10-15 minutes to complete (section 2.3.5), it was assumed that no significant changes in discharge occurred at any of the troughs during the measurement period.

Spatial variability of trough throughflow was analysed by comparing the relative proportions of flow carried by each of the nine troughs at different time periods. The relative proportion of flow carried by trough i at time t was calculated as

$$P_{it} = \frac{q_{it}}{Q_t} \quad (2.2)$$

where P_{it} is the proportion of total trough flow carried by trough i at time t , q_{it} is the discharge observed at trough i at time t , and $Q_t = \sum_{i=1}^9 q_{it}$.

2.5.2.2 Plot-scale throughflow proportions

Plot discharge was compared to hillslope discharge measured at the culvert outlet to examine how throughflow proportions vary at larger spatial scales. Since troughs 1 and 2 did not drain to the culvert outlet (Figure 2.7), their discharge was added to the measured culvert discharge.

Since culvert discharge incorporates Hortonian overland flow from the road

surface during periods of intense rainfall, comparisons of throughflow proportions were only conducted for measurements in which no precipitation was recorded during the previous hour. Assuming a maximum travel length of 25 m and a minimum flow velocity of 1 to 3 cm/s (Dunne and Leopold, 1978), then 1 hr provides adequate time for Hortonian overland flow-generated runoff to exit via the culvert outlet.

2.5.3 Water table configuration

Contour plots of the water table immediately upslope from the troughs were generated for measurement periods in which plot discharge did not vary by greater than 10% before and after piezometer measurements. Since the throughflow system may have been in a state of transient response during the measurement period, this method of data screening provides a reasonable approximation of steady-state conditions.

Irregularly-spaced piezometer data were converted to a regularly-spaced grid (0.10 m spacing) using kriging (linear semivariogram) and contoured at a 0.10 m interval with SURFER. No adjustments to original piezometric data were made to reflect the true local water table elevation assuming Dupuit flow (section 3.3.1). Flow tubes to individual troughs were manually drawn onto contour plots by assuming flow is orthogonal to water table contours.

Changes in water table configuration were quantitatively assessed by comparing measured hydraulic head to the surface and subsurface elevations at piezometer locations. The vertical separation between the water table and the elevation of the ground or till surface was calculated at each piezometer location for a given measurement period. The

vertical separation with respect to surface and subsurface topographies were calculated as:

$$\alpha_{it} = h_{it} - z_{si} \quad (2.3)$$

$$\beta_{it} = h_{it} - z_{ci} \quad (2.4)$$

where h_{it} is the height of water table at piezometer i at time t , z_{si} and z_{ci} are respectively the ground surface and till surface elevation at piezometer i , α_{it} is the vertical separation between the water table and ground surface at time t , and β_{it} is the vertical separation between the water table and basal confining layer at time t .

The standard deviation of estimates of α_{it} and β_{it} at selected piezometer locations was used to assess the fit of the water table with either surface or subsurface topography at different flow periods. A standard deviation of zero would indicate that the water table is parallel to the given topography. Since the number of piezometers registering a response varied with the flow, a subset of 15 piezometers was selected which exhibited response over a wide range of flow conditions (P-3 to P-8, P-10 to P-12, P-17 to P-21, and P-35; Figure 2.7). The subset of piezometers was also chosen so that all standard deviations computed between time periods were computed from a standard sample of piezometers.

2.5.4 Trough throughflow distribution compared to theoretical models

The observed throughflow distribution among troughs was compared to predictions from two models: (1) quasi-steady state topographically-driven flow with spatially-uniform recharge and (2) Dupuit flow.

Under the assumption of quasi-steady state, topographically-driven flow with spatially-uniform recharge, the distribution of throughflow should conform to the distribution of upslope surface (or subsurface) contributing area (equation 1.3). Therefore, the predicted throughflow proportion at any trough may be calculated as

$$f_{it} = \frac{a_i}{\sum_{i=1}^9 a_i} \quad (2.5)$$

where f_{it} is the predicted throughflow proportion at trough i at time t , and a_i is the upslope surface (or subsurface) contributing area to trough i .

Assuming soil hydraulic conductivity is homogeneous and isotropic, predicted throughflow proportions based on Dupuit flow may be related to the geometry of the upslope saturated zone draining to individual troughs

$$f_{it} = \frac{w_{it} d_{it} \sin \theta_{it}}{\sum_{i=1}^9 w_{it} d_{it} \sin \theta_{it}} \quad (2.6)$$

where w_{it} is the flow tube width, d_{it} is the average saturated thickness above the basal confining layer, and $\sin \theta_{it}$ is the local slope of the water table at trough i at time t . Flow tubes were interpreted manually from computer-interpolated contour maps of the water table surface (section 2.5.4). Care was taken to locate measurements of w_{it} and d_{it} between piezometers measuring a response and at least 2 m upslope of the road cut to minimize draw-down of the water table. Since computer-interpolated contour maps of the water table surface were only generated for periods in which plot throughflow was in near-steady state, not all sets of piezometric measurements were eligible for this analysis.

Both models were compared to observed throughflow distributions at different time periods using the following error index:

$$EI_t = \sum_{i=1}^9 (P_{it} - f_{it})^2 \quad (2.7)$$

where EI_t is the error index representing the similarity between observed and predicted throughflow distributions across the instrumented plot at time t . The greater the value of EI_t , the greater the discrepancy between observed and theoretical distributions of throughflow.

2.5.5 Effective hydraulic conductivity

Effective hydraulic conductivity (K_{eff}) is an integrated measure of the combined flow through macropores and soil matrix. Assuming that the soil is homogeneous and isotropic and that the hydraulic gradient can be reasonably approximated by the water table surface, then the effective hydraulic conductivity can be back-calculated from Darcy's Law as:

$$K_{effi} = \frac{Q_{it}}{w_{it} d_{it} \sin \theta_{it}} \quad (2.8)$$

where K_{effi} and Q_{it} are respectively the effective hydraulic conductivity and measured discharge at trough i at time t , and w_{it} , d_{it} , and $\sin \theta_{it}$ are respectively the width, mean depth, and mean hydraulic gradient determined from a cross-section of the flow tube draining to trough i at time t . Since the form of Darcy's Law in equation 2.5 assumes steady-state flow conditions, K_{eff} was only calculated for troughs in which trough and plot discharge did not vary by greater than 10% before and after piezometer measurements.

Stream lines to each trough were hand-drawn onto kriged contour maps of water table configuration assuming flow was perpendicular to water table contours.

Representative cross-sections across flow tubes were located approximately 2-4 m upslope of the road cut in order to minimize draw-down of the water table. Cross-sections were located close to, or between, piezometers registering a response. Estimates of K_{eff} were not calculated at troughs where piezometers upslope did not exhibit any response.

Kriged contour maps of the saturated thickness above compacted till were used to estimate the mean thickness of saturated subsurface flow at each cross-section. Saturated cross-sectional area for flow was then estimated by multiplying the average saturated thickness by the width of the cross-section.

Effective hydraulic conductivity was also estimated for troughs 4 to 9 lumped together to provide estimates of K_{eff} integrated over a larger contour segment (6.67 m; Table 2.1). The larger contour segment provided a more stable cross-section at which to determine saturated cross-sectional area for flow. Since the water table configuration directly upslope of troughs 4 to 9 exhibited regions of slight convergence and divergence, the cross-section was oriented roughly perpendicular to the average flow direction. Average saturated thickness above compacted till along the cross-section was calculated digitally using the “slice” procedure provided in SURFER.

2.6 OVERVIEW OF THE STUDY PERIOD

The study period extended from January to April, 1997. During this period, three storm events were monitored intensively (Table 2.3). Several inter-storm periods were also

monitored on a less-frequent interval to provide information on low flow conditions.

Maximum rainfall intensities recorded at the AES station at the Research Forest headquarters for the 1-, 6-, 12-, and 24-hour durations are summarized in Table 2.4 and compared to IDF curves for the station in Figure 2.14. The January 29-31 storm had a 6-hour rainfall intensity, with a return period of over 8-years, based on rainfall measured at the AES gauge at the Research Forest headquarters.

The study period was unusually wet when compared to long-term climate observations at the Vancouver International Airport which has been in continuous operation since 1937. Total precipitation measured during January and March 1997 was respectively 161% and 221% above normal while February 1997 was 83% of normal (Table 2.5). Total precipitation measured during the winter of 1996/97 and for the month of March 1997 was the wettest on record at Vancouver International Airport. Total rainfall on March 18, 1997, was the second greatest one day rainfall on record at Vancouver International Airport. Total precipitation measured during the month of January 1997 was the third wettest on record.

Table 2.3 Overview of major storms intensively monitored

Storm event	Data from study site		Data from AES station			
					Antecedent rainfall	
	Max. plot discharge (mL/s)	Max. saturated thickness at well (cm)	Total rainfall (mm)	Approx. duration (hr)	2-day (mm)	5-day (mm)
Jan. 29-31	1220*	65.4	82.0	20	27	27
Feb. 18-19	87.5	29.2	21.3	42	10.4	71.4
Mar. 17-20	677	63.0	193	70	21.8	23

* Maximum plot discharge for January storm was estimated as flow to trough 1 was exceeded during hydrograph peak. Trough 1 peak discharge estimated from non-linear relation (see section 3.1.3).

Table 2.4 Maximum rainfall intensities measured at the Research Forest headquarters during the period of study

Rainfall Duration (hr)	Maximum Intensity (mm/hr)	Date of Occurrence	Approx. return period based on IDF curves (yr)
1	11.6	Jan. 30	2
6	8.30	Jan. 29-30	8
12	6.27	Jan. 29-30	6
24	4.17	Mar. 19-20	3

Table 2.5 Total monthly precipitation recorded during study period and climate normals (1961-1990) at Vancouver International Airport

Month(s)	Total precipitation (mm)		Percentage of normal precipitation
	Normal	1997	
January	149.8	241.3	161 %
February	123.6	103.0	83 %
March	108.8	240.0	221 %
October - April	921.3	1443.4	157 %

Figure 2.14 Intensity-duration-frequency (IDF) curves for UBC Research Forest headquarters, 1963 to 1990. Maximum recorded rainfall intensities during study period for 1, 6, 12, and 24 hour durations labelled by open circles.

CHAPTER 3

RESULTS

This chapter starts with an examination of estimated contributing areas for individual troughs and the plot in section 3.1. Observed throughflow variability at trough and plot scale is then discussed in section 3.2. Water table configuration upslope of the troughs is then discussed in section 3.3, followed by an evaluation of model predictions of the observed throughflow variability in section 3.4. The chapter closes with an examination of the lateral and vertical variability in estimates of effective hydraulic conductivity in section 3.5.

3.1 CONTRIBUTING AREA ESTIMATES

3.1.1 Estimates of contributing areas based on surface topography

Surface contributing areas to each of the nine troughs are summarized in Table 3.1.

Coefficients of variation (C.V.) for contributing area estimates within each interpolation method averaged from 7% to 12% (maximum 24%). Variability in trough contributing area estimates between trials was lowest using the kriging interpolation method, while the minimum curvature and triangulation interpolation methods exhibited slightly higher variability.

Surface contributing area fractions (f_i) were generally similar between interpolation methods. Mean estimates of plot contributing area based on surface topography were similar between the three interpolation methods, ranging from 415 to

Table 3.1 Estimates of trough contributing areas based on surface topography using kriging, minimum curvature, and triangulation interpolation methods. Residual area represents unsampled hillslope segment between troughs 1 and 2 and is not included in estimates of plot contributing area or trough contributing area fractions (f_i). C.V. represents coefficient of variation.

Interpolation Method	Trial	T1	Residual area	T2	T3	Trough contributing areas (m ²)						Plot contributing area (m ²)
						T4	T5	T6	T7	T8	T9	
Kriging	1	13.35	11.68	323.47	6.60	4.39	4.79	10.08	34.59	7.90	9.55	414.72
	2	12.49	11.60	329.12	6.53	5.49	4.54	9.80	33.50	6.48	8.32	416.27
	3	11.91	12.11	330.81	5.40	5.19	4.47	9.06	31.10	9.48	7.49	414.91
	4	12.65	11.87	328.06	6.05	5.27	4.66	9.89	34.97	6.34	7.12	415.01
	mean	12.60	11.82	327.87	6.15	5.09	4.62	9.71	33.54	7.55	8.12	415.23
	C.V.	4.7%	1.9%	1.0%	9.0%	9.4%	3.0%	4.6%	5.2%	19.4%	13.3%	0.2%
	f_i	3.0%		79.0%	1.5%	1.2%	1.1%	2.3%	8.1%	1.8%	2.0%	
Minimum curvature	1	25.10	13.70	311.74	5.70	3.73	5.30	7.99	3.02	55.85	7.59	426.02
	2	26.87	10.93	322.86	4.28	4.30	6.31	6.59	2.88	46.54	5.54	426.17
	3	27.22	9.11	322.90	5.20	5.20	4.18	8.07	2.47	50.01	5.54	430.79
	4	26.42	12.70	308.40	5.38	3.97	4.90	8.47	4.19	54.04	6.05	421.82
	mean	26.40	11.61	316.48	5.14	4.30	5.17	7.78	3.14	51.61	6.18	426.20
	C.V.	3.5%	17.4%	2.4%	11.9%	15.0%	17.2%	10.5%	23.5%	8.1%	15.7%	0.9%
	f_i	6.2%		74.3%	1.2%	1.0%	1.2%	1.8%	0.7%	12.1%	1.5%	
Triangulation	1	24.55	6.21	335.09	6.04	4.35	5.10	9.23	4.97	51.56	6.38	447.27
	2	20.30	7.09	337.73	7.00	6.62	3.70	8.56	3.54	49.10	6.03	442.58
	3	19.06	6.81	333.49	6.70	6.66	4.06	9.03	2.98	50.95	5.80	438.73
	4	17.30	8.23	337.34	6.76	5.48	5.68	8.00	3.31	52.30	5.93	442.10
	mean	20.30	7.09	335.91	6.63	5.78	4.64	8.71	3.70	50.98	6.04	442.67
	C.V.	15.2%	12.0%	0.6%	6.2%	19.0%	19.7%	6.3%	23.7%	2.7%	4.1%	0.8%
	f_i	4.6%		75.9%	1.5%	1.3%	1.0%	2.0%	0.8%	11.5%	1.4%	

443 m². All three methods indicated that trough 2 drained the largest upslope area (74-79% of total plot area), while troughs 3, 4, and 5 drained the smallest areas (Table 3.1). Mean estimates of the surface contributing area to the residual hillslope segment between troughs 1 and 2 ranged from 7.1 to 11.8 m², less than 3% of the plot contributing area. Significant differences were observed in contributing area estimates for troughs 7 and 8. Both the minimum curvature and triangulation interpolation methods provided similar contributing area estimates for trough 7 and 8, with trough 8 having the larger of the two areas. However, the kriging interpolation method estimated the opposite, with trough 7 having the larger of the two areas (Table 3.1).

3.1.2 Estimates of contributing areas based on subsurface topography

Estimates of subsurface trough contributing areas based on kriging, minimum curvature, and triangulation interpolation methods are summarized in Table 3.2. Coefficients of variation (C.V.) for contributing area estimates within each interpolation method averaged from 9% to 12% (maximum 23%). Variability in trough contributing area estimates between trials was lowest using the kriging interpolation method, while estimates based on the minimum curvature and triangulation interpolation methods exhibited slightly higher variability.

Estimates of trough contributing area fractions (f_i) based on subsurface topography were reasonably similar among the three interpolation methods (Table 3.2). However, substantial differences in contributing areas to troughs 6, 8, and 9 were observed between the three interpolation methods. All three interpolation methods

Table 3.2 Estimates of trough contributing areas based on subsurface topography using kriging, minimum curvature, and triangulation interpolation methods. Residual area represents unsampled hillslope segment between troughs 1 and 2 and is not included in estimates of plot contributing area or trough contributing area fractions (f_i). C.V. represents coefficient of variation.

Interpolation Method	Trial	T1	Residual area	T2	T3	T4	T5	T6	T7	T8	T9	Plot contributing area (m ²)
Kriging	1	7.41	8.46	2.61	0.88	1.23	1.80	64.26	3.08	124.97	140.74	346.98
	2	8.23	9.12	1.88	0.81	1.23	2.10	73.88	3.96	139.82	120.90	352.81
	3	7.28	9.92	2.30	0.82	1.31	1.80	68.23	2.54	113.59	155.06	352.91
	4	8.12	10.48	2.09	0.77	1.35	1.70	70.35	3.16	120.05	145.02	352.60
	mean	7.76	9.50	2.22	0.82	1.28	1.85	69.18	3.18	124.61	140.43	351.32
	C.V.	6.3%	9.3%	14.0%	5.8%	4.6%	9.5%	5.8%	18.5%	9.0%	10.2%	0.8%
	f_i	2.2%		0.6%	0.2%	0.4%	0.5%	19.7%	0.9%	35.5%	40.0%	
Minimum curvature	1	7.49	9.03	1.61	0.64	1.14	2.12	221.18	3.68	16.28	136.99	391.12
	2	8.17	9.24	1.16	0.60	1.22	2.22	223.02	6.08	9.30	134.56	386.33
	3	8.43	8.53	1.60	0.55	1.35	2.10	225.48	4.64	15.95	131.37	391.47
	4	9.15	9.17	1.94	0.63	1.52	2.11	218.48	4.09	15.57	139.60	393.09
	mean	8.31	8.99	1.58	0.61	1.31	2.14	222.04	4.62	14.28	135.63	390.50
	C.V.	8.3%	3.6%	20.3%	6.3%	12.5%	2.7%	1.3%	22.7%	23.3%	2.6%	0.75%
	f_i	2.1%		0.4%	0.2%	0.3%	0.5%	56.9%	1.2%	3.7%	34.7%	
Triangulation	1	4.90	13.35	2.24	0.72	1.23	1.82	186.36	3.90	16.49	90.09	307.74
	2	5.57	11.26	2.28	0.90	1.11	1.98	182.99	3.97	19.02	86.07	303.89
	3	4.85	12.04	2.91	0.86	1.05	2.39	180.27	6.04	14.95	108.32	321.62
	4	4.36	12.04	2.52	0.84	1.02	1.72	188.15	7.26	15.16	92.40	313.43
	mean	4.92	12.17	2.49	0.83	1.10	1.98	184.44	5.29	16.41	94.22	311.67
	C.V.	10.1%	7.1%	12.4%	9.3%	8.4%	14.9%	1.9%	31.1%	11.4%	10.4%	2.47%
	f_i	1.6%		0.8%	0.3%	0.4%	0.6%	59.2%	1.7%	5.3%	30.2%	

indicated that troughs 6 to 9 drain the majority of the plot contributing area (approximately 95% of total plot area).

Mean estimates of plot contributing area ranged from 311 to 390 m², depending on the interpolation method used (Table 3.2). The greater variability in mean plot contributing area estimates based on subsurface topography, as compared to surface topography, may be related to the nature of spacing of survey points. Although the overall survey point spacing was reasonably similar for both surface and subsurface topographies (0.10 and 0.09 m⁻² respectively; sections 2.4.1 and 2.4.2), a higher density of points were surveyed in the plot contributing area for surface topography. Mean estimates of the residual hillslope segment between troughs 1 and 2 were small, ranging from 8.99 to 12.17 m², comprising less than 3% of the plot area.

3.1.3 Contributing area estimates based on water balance

Outflow and precipitation measurements made during the January 29-31, 1997 storm were used to estimate trough and plot contributing areas based on the water balance. Figure 3.1 illustrates the rainfall hyetograph, plot hydrograph, and associated discharge measurements for the storm. Total rainfall recorded for the storm at the tipping bucket gauge on-site was 82 mm (Figure 3.1; Table 2.3).

Total estimated outflow volume at individual troughs and the plot was integrated at a 10 minute interval from January 29 at 11:10 hrs to January 31 at 15:10 hrs. This time period corresponded to the period of intensive discharge measurement (Figure 3.1). Discharge was linearly interpolated between measurement periods, except for the period

Figure 3.1 Rainfall hyetograph and estimated plot hydrograph for January 29-31, 1997 storm. Actual discharge measurements are labelled with open diamonds. Trough 1 discharges were estimated for the hydrograph peak using a non-linear function of saturated thickness at water table well.

from January 30 16:00 to January 31 12:00 hrs. This period was interpolated using relations between trough discharge and saturated thickness at the water table well, which was continuously recorded.

During the hydrograph peak, discharge at trough 1 exceeded the trough's capacity since the aluminum tray for shunting macropore flow past the trough had not yet been constructed (section 2.3.1). Missing values of trough 1 discharge during this period were reconstructed using two methods to provide a range in estimates of runoff volume. The first method assumed that trough 1 discharge during the hydrograph peak was equal to the approximate discharge capacity of the trough (ca. 200 mL/s; section 2.3.1), providing a minimum estimate of total outflow volume from trough 1. The second technique predicted trough 1 discharge (Q_{T1}) as a non-linear function of saturated thickness (z) observed at the water table well (Figure 3.2). Non-linear regression gave an equation of $Q_{T1} = -9.99 + 3.18 \times 10^{-8} z^{5.69}$ ($n = 22$; $R^2 = 0.98$) for saturated thicknesses greater than 30 cm.

Total estimated volume of runoff and estimated effective contributing areas for troughs and plot are summarized in Table 3.3. Estimated trough contributing areas based on outflow were generally greater than estimates based on either surface or subsurface topography (Table 3.3). The best estimate of plot contributing area based on outflow (i.e., 531 m²) was 24% and 51% greater than estimates based on surface or subsurface topography, respectively (Table 3.3).

It is likely that the estimates of contributing areas provided in Table 3.3 were minimum estimates for two reasons. Firstly, post-storm discharge and average saturated

thickness exceeded pre-storm levels (Table 3.4). Therefore, the change in soil moisture storage is likely positive. Secondly, it was assumed that interception and evaporation by the tree canopy was negligible. Net precipitation available for runoff was assumed to be equal to the total precipitation recorded by the tipping bucket gauge instrumented on site in a nearby clearing.

Figure 3.2 Discharge at trough 1 in relation to saturated zone thickness measured at the water table well. Non-linear relation derived from measurements made during Jan. 29-31 and Mar. 17-20, 1997 storms.

Table 3.3 Estimated contributing areas based on observed and predicted outflow for January 29-31, 1997 storm

Trough number	Observed volume of outflow (m ³)	Effective contributing area (m ²)	Surface contributing area (m ²) [‡]	Subsurface contributing area (m ²) [‡]
T1	(9.0) 14.6 *	(110) 178 *	19.8	7.0
T2	3.7	46	327	2.1
T3	0.3	3	6.0	0.8
T4	0.6	7	5.1	1.2
T5	1.2	15	4.8	2.0
T6	7.2	88	8.7	159
T7	5.3	65	13.5	4.4
T8	5.1	62	36.7	51.8
T9	5.4	66	6.8	123
Plot	(38.0) 43.6 *	(463) 531 *	428	351

* Minimum estimates of outflow volume and effective contributing area for trough 1 and plot shown in brackets

‡ Estimates of contributing area based on averages of all three interpolations from Tables 3.1 and 3.2

Table 3.4 Comparison of pre- and post-storm discharges at troughs and plot

Trough	Pre-storm discharge* (mL/s)	Post-storm discharge** (mL/s)
T1	0.402	0.471
T2	1.41	4.07
T3	0.0	0.32
T4	1.33	2.10
T5	0.456	6.03
T6	4.64	12.2
T7	2.47	13.2
T8	1.96	11.2
T9	8.99	18.3
Plot	21.8	67.8
Avg. sat. thickness [‡] (cm)	9.0	13.9

* Pre-storm discharges measured on Jan. 29/97 11:10 hrs.

** Post-storm discharges measured on Jan. 31/97 15:10 hrs.

‡ Based on piezometers registering a water table response

3.2 THROUGHFLOW VARIABILITY

3.2.1 Trough throughflow variability

Maximum discharges recorded at individual troughs ranged from 4.55 to over 336 mL/s (Table 3.5). Maximum discharges for all troughs except trough 1 (T1) were recorded on January 30, 1997 between 3:00 and 4:00 PST. Trough 1 became inoperable during this period as discharge exceeded the trough's capacity. Maximum discharge for trough 1 was recorded on March 18, 1997 at 13:26 PST.

The proportion of total throughflow carried by individual troughs exhibited three general trends with increased plot discharge (Figure 3.3): decreasing flow proportion with increased plot flow (T4, T9); peak flow proportion at moderate plot flow (T2, T3, T5-8); and peak throughflow proportions at high flow (T1). Trough throughflow proportions generally became less variable and assumed a more constant proportion as plot flow increased (Figure 3.3).

Table 3.5 Maximum trough discharges recorded during the study period

Trough	T1	T2	T3	T4	T5	T6	T7	T8	T9
Max. discharge (mL/s)	336	138	4.55	6.94	28.8	119	76.3	98.1	63.2

Figure 3.3 Trough throughflow proportions at measured values of plot discharge.
Note different scales for y-axis.

Mean trough throughflow proportions were summarized into three categories on the basis of plot discharge: low (0-149 mL/s), moderate (150-349 mL/s), and high (>350 mL/s) flows (Figure 3.4). At low flows, plot throughflow was dominated by troughs 6 to 9 (Figure 3.4). The proportion of flow issued by these troughs (ca. 80-90%) was similar to the estimated subsurface contributing area fraction estimated for the troughs (ca. 96%; Table 3.2). During extremely low flows, however, trough 9 accounted for up to 80% of the flow issued from the plot (Figure 3.3). Visual survey of the road cut face at trough 9 revealed that the trough appeared to be located at the base of a till depression. Furthermore, a large macropore located at the soil-till contact often conveyed flow (> 10 mL/s) at trough 9 when adjacent troughs (i.e. troughs 7 and 8) were limited to a slow drip (ca. < 5 mL/s). Flow convergence due to the combined effect of macropores and till topography may account for trough 9's dominance at very low flows.

As plot discharge increased, throughflow proportions to troughs 3 to 9 steadily decreased while throughflow proportions to trough 1 increased (Figures 3.3, 3.4). At high flows, trough 1 conveyed the majority of flow issued by the plot. Visual observation of trough 1 at moderate to high flows indicated the activation of a discrete macropore located approximately 30-40 cm above the soil-till contact (Figure 2.8). When the saturated thickness was less than 30 cm, flow at trough 1 was less than 5 mL/s (Figure 3.2). However, once the saturated thickness exceeded approximately 30-40 cm, flow at the trough increased dramatically. The macropore at trough 1 was apparently activated when the water table rose high enough to submerge a section of the macropore system, enabling water entry.

Figure 3.4 Mean trough throughflow proportions for low (0-149 mL/s), moderate (150-349 mL/s), and high (>350 mL/s) plot flows.

3.2.3 Plot throughflow

The proportion of hillslope flow issued by the plot was monitored for periods in which the one-hour antecedent precipitation was zero (section 2.5.2.2). As a result of this criterion, most of the data points eligible for the analysis were measured during the falling limb of the hydrograph. Since troughs 1 and 2 drained to an unnamed tributary to the west of the plot, their discharges were added to culvert discharges to estimate hillslope flow. Discharge from the unsampled contour segment between troughs 1 and 2 was assumed to be negligible (Figure 2.7). Hillslope contributing area was estimated to be approximately 2000 m² (section 2.3.2).

At flows less than 500 mL/s, the proportion of hillslope flow transmitted by the plot ranged from 30% to 60% (Figure 3.5). At higher flows, the proportion of hillslope flow issued by the plot ranged from approximately 20% to 30%, which is similar to the ratio of contributing areas based on surface and subsurface topographies and plot outflow estimates (Table 3.6).

Table 3.6 Comparison of the ratio of plot to hillslope contributing areas based on surface and subsurface topography and plot outflow estimates

Method		A _{plot} (m ²)	A _{plot} :A _{hill} [*]
Surface topography	Kriging	415	20.8 %
	Min. Curv.	426	21.3 %
	Trangulation	443	22.2 %
Subsurface topography	Kriging	351	17.6 %
	Min. Curv.	391	19.6 %
	Triangulation	312	15.6 %
Water balance		463 - 531	23.1 - 26.6 % [‡]

^{*} Subsurface contributing area of hillslope was assumed to be approximately 2000 m² (section 2.3.2)

[‡] Based on range of plot contributing area estimates

Figure 3.5 Proportion of hillslope discharge carried by the plot, as a function of total hillslope discharge. Data represent measurement periods in which 1-hour antecedent precipitation was zero to minimize Hortonian overland flow contributions from the road.

3.3 WATER TABLE CONFIGURATION

3.3.1 Piezometric response

Water levels recorded in piezometers varied dramatically during the study period.

Maximum saturated thickness observed at the water table well was 65.4 cm, recorded on January 30, 1997 at 3:20 PST. All piezometers registered a water level response at least once during the study period. However, piezometers P-24 and P-32 (Figure 2.7) remained dry over much of the study period except during peak flow during the January 29-30 storm.

As mentioned in section 2.3.3, piezometers underestimate the local water table elevation if Dupuit-flow is assumed. Figure 3.6 illustrates the slight differences in water table configuration using corrected and original values of water table elevation measured during a period of high plot discharge (576 mL/s). Differences were greatest upslope of troughs 6 to 9 where the saturated thickness was greatest. Little change was observed in the water table configuration directly upslope of troughs 2 to 5 where saturated thicknesses were generally minimal. The similarity between uncorrected and corrected water table configurations at high flows suggests that flow tubes to individual troughs would likely remain similar over all flows. Subsequent analyses of saturated wedge configuration were therefore based on uncorrected piezometer measurements.

3.3.2 Water table configuration compared to topography

As observed from Figure 3.7, interpreted flow tubes to individual troughs varied substantially at different magnitudes of plot flow. At low flows, the water table

Figure 3.6 Differences in water table configuration using uncorrected and corrected water table elevations at piezometer locations. Date of measurement: March 18, 1997 at approximately 9:50 hrs. Plot discharge is 576 mL/s.

A) Jan. 31/97 Plot flow = 64.7 mL/s

B) Jan. 30/97 Plot flow = 193 mL/s

C) Jan. 30/97 Plot flow = 374 mL/s

D) Mar. 18/97 Plot flow = 576 mL/s

Legend

- Trough divides
- Piezometers (response)
- Streamlines
- × Piezometers (no response)

Figure 3.7 Water table configuration at selected values of plot flow. Water table contour interval is 0.10 m

configuration emulated the topography of the compacted till layer. Regions of convergence are observed to trough 9, while regions of divergence are located at troughs 2 to 8 (Figure 3.7a). Between the low and moderate flow periods (Figure 3.7a,b), water table elevation increased substantially in the region directly upslope of troughs 6 to 9. This increase caused hydraulic gradients to become more uniform in magnitude and direction across the hillslope, thereby reducing flow convergence to trough 9. As plot flow increased, flow lines generally became oriented perpendicular to surface topographic contours. Flow divergence in the region directly upslope of troughs 3 to 5 was apparent over all flow conditions, reflecting subsurface topographic divergence.

Figure 3.8 shows a lateral cross-section across the base of the plot (Figure 3.8). At low flows (22.1 and 69.9 mL/s), the water table appeared to conform well with subsurface topography at the cross-section. At higher flows (193, 309, and 677 mL/s), increases in saturated zone thickness were concentrated in depressions in the till surface which caused the water table to conform more closely with surface topography.

Figure 3.9 shows that, at low flows, the standard deviation with respect to subsurface topography was lower than with respect to surface topography, indicating that the water table conformed more closely with subsurface topography. At higher flows (>100-200 mL/s), however, the standard deviation was lower with respect to surface topography. Therefore, the water table became more parallel with the surface topography at higher flows.

Figure 3.12 Lateral cross-section (A-A') of hillslope illustrating shift in water table configuration with increasing plot discharge. Inset map illustrates location of cross-section; contours indicate surface topography. Vertical exaggeration of cross-section is 3.7x.

Figure 3.9 Standard deviation of the vertical separation between water table and surface or subsurface topography at selected piezometer locations.

3.4 EVALUATION OF MODEL PREDICTIONS OF THROUGHFLOW VARIABILITY

3.4.1 Predictions based on quasi-steady state topographically-driven flow

Figures 3.10 and 3.11 compare observed trough throughflow to predictions based on surface and subsurface topography, respectively. Since trough contributing area fractions depended on interpolation method used (sections 3.1.1, 3.1.2), results using all three interpolation methods are shown in Figures 3.10 and 3.11.

The high error index observed at low plot flows (<20 mL/s) indicates that the observed throughflow distribution conformed poorly to the predicted distribution based on surface contributing areas (Figure 3.10). As plot flow increased, the error index dropped for plot flows greater than approximately 30 mL/s, then increased slightly as flow increased above 100 mL/s (Figure 3.10).

The observed trend in the error index with respect to subsurface topography was similar to that for surface topography (Figures 3.10, 3.11). However, error index values were lower over all observed flows. The error index increased for flows less than 20 mL/s, indicating that the observed throughflow distribution conformed poorly to the predicted distribution based on subsurface contributing areas (Figure 3.11). The error index dropped to low values for plot flows of approximately 20-100 mL/s and steadily increased for plot flows greater than 100 mL/s (Figure 3.11). At high flows, values of the error index approached those based on surface contributing areas.

Figure 3.10 Error index based on surface topography

Figure 3.11 Error index based on subsurface topography

3.4.2 Predictions based on Dupuit flow

Similarity between observed and predicted throughflow distributions based on Dupuit flow is presented in Figure 3.12. Only data points for periods when near-steady state conditions prevailed were included in the analysis (section 2.5.4).

The error index with respect to Dupuit flow was lowest at plot discharges of approximately 100-300 mL/s (Figure 3.12). At plot flows less than 100 mL/s, and greater than 300 mL/s, the error index increased. The high error index at flows less than 100 mL/s was probably related to increased uncertainty associated with estimating saturated cross-sectional area and hydraulic gradient from interpolated piezometer data at low plot flows or preferential flow via the permeable root mat located along the upper surface of the compacted till.

Figure 3.12 Error index based on Dupuit flow.

3.4.3 Summary

The predictive accuracy of topographically-driven flow and Dupuit flow models is summarized in Figure 3.13. The error index computed using surface contributing areas is based on kriging, while the error index computed using subsurface contributing areas is based on minimum curvature. These interpolation methods provided the lowest error index with respect to each topography (Figures 3.10, 3.11).

The observed throughflow distributions conformed best to the Dupuit flow model, except at flows less than 100 mL/s, when the observed throughflow distribution conformed best to the distribution of subsurface contributing areas (Figure 3.13). Over all observed plot flows, the topographically-driven flow model based on surface topography provided the worst fit with the observed throughflow distribution.

All three models exhibited an increase in the error index as plot flow increased greater than approximately 150 mL/s. The increasing disagreement between predicted and observed throughflow distributions is related to the activation of the macropore at trough 1 at moderate to high flows (Figure 3.2).

Figure 3.13 Comparison of error indices for topographically-driven flow and Dupuit flow models. Note that error indices for topographically-driven flow are based on kriging (surface topography) and minimum curvature (subsurface topography) interpolations.

3.5 EFFECTIVE HYDRAULIC CONDUCTIVITY

3.5.1 Estimates of K_{eff} at the trough scale

Effective hydraulic conductivity (K_{eff}) was estimated at each of the nine troughs for selected times representing a range in plot discharge (Figure 3.14). Estimates of K_{eff} were only calculated for troughs for which the upslope saturated zone was defined by piezometric measurements and measured discharge conformed to the near-steady state criterion outlined in section 2.5.5.

Trough estimates of K_{eff} generally exhibited a positive trend with increased flow (Figure 3.14). Variability in estimates of K_{eff} ranged up to two orders of magnitude with a minimum variability for plot discharges between 90 and 200 mL/s (Figure 3.14). The increased variability in K_{eff} at low flows may reflect increased error in estimating saturated cross-sectional area from piezometer measurements. However, the increased variability at low plot flows may also reflect preferential flow via the permeable root mat along the upper till surface.

As plot discharge increased above 200 mL/s, K_{eff} estimates increased in variability (Figure 3.14). Values of K_{eff} greater than 1×10^{-3} m/s at high flows were primarily derived from troughs 1 and 2 (Figure 3.14). At least some of the increased variability at high flows may be associated with preferential flow via macropores which became activated once the water table rose above a threshold elevation (e.g., trough 1; Figure 3.2).

Figure 3.14 Variability in trough estimates of effective hydraulic conductivity (K_{eff}) at different plot flows.

Relations between K_{eff} and saturated thickness for each of the nine troughs are shown in Figure 3.15. However, these relations are confounded by variations in flow tube width to individual troughs, especially at plot discharges less than 200 mL/s (Figure 3.16). Therefore, the vertical variation in K_{eff} for individual troughs reflects both lateral and vertical variability. The large variability in flow tube width at flows less than 200 mL/s roughly corresponds with the main changes to water table configuration observed in Figure 3.7.

Troughs 4, 5 and 6 exhibited little variation in flow tube width with increased plot discharge and may be good candidates to examine vertical variation in K_{eff} (Figure 3.16). Relations between K_{eff} and saturated thickness for these troughs were steadily decreasing (trough 4), steadily decreasing, then steady (trough 5), or slightly increasing (trough 6) (Figure 3.15). However, no consistent relation was observed between troughs.

Figure 3.15 Variation in effective hydraulic conductivity (K_{eff}) with mean saturated thickness above compacted till at individual troughs.

Figure 3.16 Flow tube widths draining to troughs as a function of plot discharge.

3.5.2 Estimates of K_{eff} at the plot scale

Plot-scale estimates of K_{eff} were estimated for the flow tube draining to the combined troughs 4-9. As seen in Figure 3.17, K_{eff} ranged from 9×10^{-5} to 4×10^{-4} m/s; which was considerably less than the observed variability at the trough scale (Figure 3.14).

The vertical variation in K_{eff} was generally linear for mean saturated thicknesses of greater than approximately 0.10 m (Figure 3.17). For saturated thicknesses greater than 0.10 m, the vertical variation in K_{eff} was best described by the linear relation, $Y = 7.32 \times 10^{-4}X + 1.04 \times 10^{-5}$ ($R^2 = 0.91$; $n = 21$; S.E.E. = 2.6×10^{-5}), where K_{eff} and mean saturated thickness are the dependent and independent variables, respectively.

Plot-scale estimates of K_{eff} increased as saturated thicknesses decreased below 0.10 m (Figure 3.17). The increase in K_{eff} may be partly due to a bias in estimating saturated cross-sectional area either through reduced flow tube width or reduced saturated thickness. However, it may also have reflected increased flow dominance via the permeable organic root mat at the soil-till contact.

Figure 3.17 Variation in K_{eff} ($\diamond$) and flow tube width (o) with mean saturated thickness for troughs 4 to 9 combined. Linear relation valid for saturated thicknesses greater than approximately 0.10 m.

CHAPTER 4

DISCUSSION

The research objectives posed in chapter one are discussed in sections 4.1 to 4.5. Sections 4.6 and 4.7 address implications for field and modelling studies, respectively.

4.1 COMPARISON OF SURFACE AND SUBSURFACE CONTRIBUTING AREA ESTIMATES

Estimates of plot contributing area were reasonably similar between surface and subsurface topography. Therefore, it appears that detailed knowledge of subsurface topography may only provide additional information at scales less than 5-10 m. This finding concurs with Woods and Rowe (1997), who observed similarities between surface and subsurface contributing area estimates at the Maimai hillslope when troughs were lumped in groups of four or five (scales of approximately 8-10 m). Woods and Rowe further argued that the major additional information provided by detailed survey of the bedrock surface at Maimai may only be at the scale of one or two troughs (i.e., contour lengths of 2-4 m).

Contributing areas estimated from digital elevation models (DEMs) of surface or subsurface topography in the present study may be susceptible to three sources of error: 'measurement' error, reflecting error in surveyed elevation and position, 'model' error, reflecting the choice of interpolation method used to interpolate a regular grid from irregularly-spaced surveyed data points, and 'human' error, involved in manually

interpreting contributing areas from contour maps.

Measurement error in surveyed data points can cause re-allocation of considerable area to adjacent troughs, especially if the points in error are located near the base of the hillslope. Woods and Rowe (1997) noted that estimates of trough contributing area at Maimai underwent a factor-of-five change in area when the original survey points were changed slightly, although they did not provide any details on the extent of their sensitivity analysis. Although a sensitivity analysis was not conducted in the present study, average errors in survey measurements based on repeated measurements of known points were less than ± 10 cm for horizontal position and less than ± 1 cm for vertical position (section 2.4.1). Measurement error in estimating the vertical position of the compacted till surface may be much greater (ca. ± 10 cm) due to augering (section 2.4.2). Therefore, survey error was likely more substantial for the DEM of subsurface topography.

Since the choice of interpolation method affects the fit of the mathematical surface to the irregularly-spaced survey points, slight differences in elevations interpolated at regularly-spaced grid nodes may result. In the present study, the observed differences in surface contributing area to troughs 7 and 8 illustrate how choice of interpolation method may influence contributing area estimates (Table 3.1). Depending on interpolation method, upslope surface contributing area was allocated in favour of trough 7 (i.e., kriging interpolation) or trough 8 (i.e., minimum curvature and triangulation).

The third source of error which may influence contributing area estimates is a

result of hand-drawing flow lines to individual troughs. Replicate measurements of trough contributing areas for each interpolation method suggested that the error involved in manually interpreting contributing areas was small (average C.V. for each interpolation method was approximately 10%; Tables 3.1, 3.2).

Effective plot contributing area based on water balance calculations was substantially greater than estimates based on either surface or subsurface topography. However, the probable bias in net precipitation available for runoff and changes in soil moisture storage suggests that contributing areas based on the water balance were a minimum estimate. The much greater plot contributing area based on the water balance could be explained by the inability of the subsurface topographic survey to resolve fine scale details of the compacted till surface, and/or lateral shunting of subsurface flow via macropores from areas outside the topographic boundaries (e.g., Tsuboyama et al., 1994).

4.2 SPATIAL VARIABILITY OF THROUGHFLOW AT TROUGH AND PLOT SCALES

Trough throughflow proportions exhibited a consistent shift in distribution with increased flow. Woods and Rowe (1996) also found a consistent relation between trough throughflow proportion and flow. However, they lacked information regarding the configuration of the saturated wedge to explain the observed relations.

The shift in the spatial distribution of trough throughflow observed in the present study may be partially explained by changes in the configuration of the water table. At low flows, the configuration of the water table conformed closely to subsurface

topography. Therefore, troughs draining zones of subsurface topographic convergence (e.g., troughs 6 to 9) dominated the observed throughflow distribution. As plot flow increased, piezometric levels in subsurface topographic depressions rose disproportionately compared to zones of subsurface divergence. The increased piezometric levels in depressions acted to reduce flow convergence to troughs instrumented at the base of subsurface topographic depressions and cause flow lines to become more perpendicular to surface elevation contours. The observation that trough throughflow proportions decreased in variability with increased flow suggests that throughflow pathways (i.e., water table configuration) became more stable at higher flows.

The shift in the spatial distribution of trough throughflow was also influenced by soil heterogeneity. Visual observation of trough 1 confirmed the activation of a soil macropore which dominated the total throughflow issued by the plot at high flows. The observation of dramatic increases in trough 1 discharge once a threshold saturated thickness was attained suggests that throughflow variability was controlled by the position of the water table. The activation of the macropore at trough 1 requires the development of lateral hydraulic gradients to feed the macropore. However, the average spacing of piezometers in the present study was not detailed enough to provide evidence of lateral hydraulic gradients. Woods and Rowe (1996) also inferred the establishment of lateral hydraulic gradients at moderate to high flows at the Maimai hillslope to explain the similarity in discharges between adjacent troughs, even though upslope surface contributing areas were substantially different. However, Woods and Rowe did not make

any measurements of the upslope saturated zone. The establishment of lateral hydraulic gradients due to the activation of macropore systems is likely very common on temperate forested hillslopes. Therefore, it may be difficult to elucidate topographic control on flow pathways for contour segments less than 10 m.

Woods and Rowe (1996) and Woods et al. (1997) attempted to explain the observed throughflow variability at Maimai hillslope in terms of the spatial pattern of soil moisture deficit, which was hypothesized to influence the spatial pattern of recharge. Although throughflow variability observed in the present study does not contradict Woods and Rowe's hypothesis, the variability can be explained simply by changes in water table configuration and flow through macropores. Indeed, the apparent spatial variability of recharge may be an artifact of the likely incorrect assumption that flow lines do not vary through time.

Although macropores exerted an important control on the spatial variability of throughflow in the present study, it is not known whether the observed macropore dominance of the trough throughflow distribution at high flows is representative for similar-sized forested hillslopes. Studies by Freer et al. (1997) at Panola Mountain Research Watershed indicated that macropore flow may dominate the observed trough throughflow distribution sampled from hillslopes approximately 20 m wide. Therefore, it may be reasonable to suggest that results from the present study are representative of similar-sized forested hillslopes. Detailed measurement of the density, orientation, and connectivity of macropore systems (e.g., Noguchi et al., 1997), in combination with measurements of throughflow, may provide information on representative scales at which

macropore systems can be effectively averaged out. However, representative scales may be difficult to define if the connectivity of macropore systems vary with water content or antecedent wetness.

The proportion of hillslope flow transmitted by the plot was generally greater than or equal to the fraction of plot:hillslope contributing area based on surface and subsurface topography, and water balance analysis (Figure 3.5; Table 3.6). At low flows, the increased proportion of hillslope flow issued by the plot suggests that the plot drained a disproportionate area of the hillslope. Possibly, the fine scale features of subsurface topography were not resolved by the auger survey or flow via the permeable organic root mat caused the plot to drain a much larger area than predicted. Hillslope discharge measured at the culvert outlet may have also been underpredicted at low flows, since a greater proportion of ditch flow may have been lost beneath the road bed or through coarse aggregate fill surrounding the culvert. As flow increased, flow likely became more uniform across the hillslope leading to the similarity between throughflow proportions and contributing area fractions.

4.3 EVALUATION OF MODEL PREDICTIONS OF THROUGHFLOW VARIABILITY

Error indices indicated that the Dupuit flow model best predicted the observed spatial distribution of throughflow among troughs except at low flows (<100 mL/s). It is likely that the Dupuit flow model was most effective in predicting throughflow variability because the model accounts for changes in water table configuration through time.

Topographically-driven models based on unslope contributing area do not adequately describe throughflow variability except at low flows, when the water table configuration conformed closely to subsurface topography.

All three models invoke the assumption that soils are homogeneous and isotropic. In reality, however, forest soils are heterogeneous and anisotropic. At very low plot flows (< 20 mL/s), the increase in error index observed for the models based on surface and subsurface contributing areas may be related to preferential flow via the permeable organic root mat to trough 9. Error indices for all three models also increased with plot flow above approximately 100 mL/s. The increased disagreement between observed and predicted throughflow distributions was likely related to the activation of the macropore system at trough 1.

4.4 SPATIAL VARIABILITY IN EFFECTIVE HYDRAULIC CONDUCTIVITY

The decreased variability in estimates of effective hydraulic conductivity between trough and plot scales suggests that a representative scale at which the effects of macropore systems is averaged out may be at scales of at least 5 to 10 m. Previous estimates of effective conductivity by Utting (1979) and Tischer (1986) at other sites within the Research Forest compare well with results from the present study (Figure 2.2; Table 4.1).

Utting's (1979) calculation of plot-scale K_{eff} is likely a minimum estimate since he assumed a maximum value of saturated thickness above compacted till based on recorded measurements from approximately 6 in-plot piezometers distributed across the

lower portion of the hillslope. The range in Tischer's (1986) estimates of plot-scale K_{eff} relate to uncertainties in estimating effective porosity at her study site. However, the other input parameters used to compute K_{eff} using the kinematic wave equation (i.e., slope of initial hydrograph rise, $\partial Q/\partial t$, and recharge rate) were similar among all three trials conducted. Both Utting and Tischer assumed that the hydraulic gradient was equal to the average surface slope, rather than calculating it from contour maps of water table configuration.

Estimates of plot-scale K_{eff} in the present study may also be a minimum estimate for hillslope scales of approximately 10 m since the analysis did not incorporate the macropore system at trough 1. Incorporating the trough 1 into plot-scale estimates of would likely increase the resulting magnitude of K_{eff} reported in the present study at higher flows.

Table 4.1 Comparison of plot scale estimates of K_{eff} measured in Research Forest

Author	Method	Plot Width (m)	Saturated thickness (m)	K_{eff} (m/s)
Utting (1979)	Plot outflow	30	0.5	0.8×10^{-4}
Tischer (1986)	Kinematic wave equation	7	0.18 - 0.22	1.6×10^{-4} - 3.2×10^{-4}
This Study	Plot outflow	7	0.09 - 0.45	1.0×10^{-4} - 3.5×10^{-4}

4.5 VERTICAL VARIATION IN EFFECTIVE HYDRAULIC CONDUCTIVITY

Relations between effective hydraulic conductivity and saturated depth at the trough scale were confounded by the fact that flow tubes to troughs changed through time. Therefore, K_{eff} estimates are based on slightly different soil cross-sections. Relations between K_{eff} and saturated thickness were not consistent even for troughs which had relatively stable flow tubes (i.e., troughs 4-6 combined). It appears that meaningful relations between K_{eff} and saturated thickness may only become apparent when troughs are grouped together at scales of approximately 5-10 m. A possible solution to examine the vertical variation in K_{eff} for narrow contour segments would be to install parallel concrete walls extending several metres upslope from the trough and to soil-till (or bedrock) contact. The concrete walls would control flow tube width draining a given contour segment. However, the walls may break continuity of macropore systems which propagate laterally across the hillslope.

Plot-scale estimates of K_{eff} , based on troughs 4 to 9 combined, generally increased linearly with increasing saturated thickness (Figure 3.18). The observed linear relation conforms with the model (equation 1.6) proposed by Ambroise et al. (1996). However, observed K_{eff} values diverged from the linear relation at saturated thicknesses less than 0.10 m. The increased value of K_{eff} at saturated thicknesses less than approximately 0.10 m may reflect increased flow dominance via the permeable organic root mat located along the upper till surface, or errors associated with estimating saturated cross-sectional area

above till at low flows.

Although the effects of trough 1 were not incorporated in plot scale observations of the vertical soil heterogeneity, the relation between K_{eff} and saturated thickness will likely remain linear for saturated thicknesses less than 30 cm observed at the water table well. However, the relation will likely become exponential for saturated thicknesses greater than 30 cm due to the dramatic increase in discharge at trough 1.

4.6 IMPLICATIONS FOR FIELD STUDIES

The present study has revealed that it may be difficult to define contributing areas draining to contour segments of 1 to 10 m in length. Even when digital elevation models of either surface or subsurface topography can be defined accurately, inferred flow paths may not be representative of the true throughflow pathways. The present study indicates that throughflow pathways, at scales of approximately 1-10 m, may vary depending on the configuration of the water table and soil anisotropy. Field studies which attempt to make comparisons of runoff per unit contributing area from instrumented hillslope segments (e.g., Sidle et al., 1995; Woods and Rowe, 1996), or attempt to make inferences regarding the path length and residence time of water on hillslopes, need to acknowledge uncertainties due to the dynamics of the flow system and effects of soil anisotropy.

In light of the dynamic nature of the flow system at small scales, water balance calculations may prove useful to estimate contributing areas draining narrow contour segments. Troendle and Reuss (1997) found that areas contributing to contour segments of 12 and 36 m, based on water balance calculations, were much greater than estimates

based on surface topography. Although the water balance technique provides information on effective contributing area, it does not provide information on the spatial extent of contributing area or changes through time (e.g., as a result of increasing antecedent wetness).

One possible solution for field studies to minimize variations in upslope contributing area may be to locate troughs across the base of subsurface convergent zones (e.g., Woods and Rowe, 1996). Locating troughs across subsurface topographic depressions would minimize changes in contributing area related to changes in water table configuration (e.g., Figure 3.7). However, this method still ignores soil anisotropy which may shunt flow laterally across subsurface topographic divides once the water table saturates a portion of the macropore system.

Most hillslope hydrologic studies measure flow from troughs, typically 1-3 m wide. However, flow tubes to narrow contour length may not remain stable due to changes in water table configuration and shunting of flow by discrete macropores. Therefore, extrapolating throughflow measurements from narrow contour segments in space or time is difficult. Hillslope scales of approximately 10 m or greater may provide a greater understanding of flow at the catchment scale since the effects of soil anisotropy and water table configuration on flow pathways may be greatly reduced. Of 19 hillslope hydrological studies examined by the author, only five utilized troughs greater than 10 m in length (i.e., Dunne and Black, 1970; Beasley, 1976; Utting, 1979; Wilson et al., 1990; and Troendle and Reuss, 1997).

4.7 IMPLICATIONS FOR MODELLING

The present study indicates that variability of throughflow, water table configuration, and effective hydraulic conductivity at scales of approximately 10 m or less can be quite dramatic. Considering that many topographically-driven flow models utilize DEM grid sizes of 10 m, or greater, it may be inappropriate to assume that hydrologic processes are homogeneous at the sub-grid scale.

The common assumption of many topographically-driven models (e.g., TOPMODEL) that hydraulic gradients are parallel to either surface or subsurface topography over all flows is not correct. The present study indicates that the water table configuration, and therefore hydraulic gradient, varies with local saturated thickness. If hydraulic conductivity is assumed to be homogeneous and isotropic, then models based on surface topography may adequately model throughflow variability at high flows, when the water table is most parallel with surface topography. However, models based on surface topography may provide a poor representation of throughflow variability at low flows, when the water table is most parallel with subsurface topography. Since saturated thickness depends on hillslope position and antecedent wetness, it may be impossible for either surface or subsurface topography to completely describe throughflow pathways. The present study indicates that models based on Dupuit flow may best describe throughflow pathways since they incorporate variations in water table configuration.

This study suggests that a representative scale at which soils may be considered homogeneous and isotropic is at least 5-10 m. At this scale, the vertical variation in hydraulic conductivity may be best described as a linear function of saturated thickness.

Many previous modelling studies have assumed that hydraulic conductivity is vertically homogeneous and therefore independent of saturated thickness (e.g., Jackson and Cundy, 1992; Ambroise et al., 1996; Woods et al., 1997). However, differences in model representations of the vertical variation in hydraulic conductivity may have important implications for routing water and chemicals downslope. Models which assume that hydraulic conductivity is vertically homogeneous will predict a constant travel or contact time of water and chemicals through soil layers. However, models which assume that hydraulic conductivity varies linearly or exponentially will predict lower travel times at high flow and hence lower concentrations of soil-derived solutes. Differences in model representations of the vertical variation in hydraulic conductivity will cause significant differences in the predicted rate and timing of water or chemicals issued from hillslopes.

Although current models typically characterize the vertical variation in hydraulic conductivity as constant or linear or exponentially increasing, none have incorporated the presence of a high conductivity zone at the base of the soil profile to represent the permeable organic root mat. The root mat may act to rapidly transmit flow and downslope, even when the perched saturated zone is thin. Field experiments by Peters et al. (1995) on the Canadian Shield suggest that 100 % of throughflow may travel via a thin weathered zone at the soil-bedrock interface. Further research is required to appropriately model vertical soil heterogeneity where a highly conductive zone at the soil-bedrock (or till) interface is observed.

CHAPTER 5

CONCLUSIONS

5.1 SUMMARY OF MAIN FINDINGS

5.1.1 Contributing areas draining narrow contour segments

Estimated contributing areas draining narrow contour segments may vary substantially depending on whether surface or subsurface topography or water balance analysis is used. Contributing area estimates based on water balance calculations were generally greater than topographically-derived estimates. The higher estimates based on water balance calculations may be related to the poor survey resolution of fine-scale details of the compacted till surface or lateral shunting of subsurface flow via macropores, or a combination of the two processes.

Variations in flow pathways due to changes in the configuration of the water table and lateral shunting of flow through macropores suggest that it may be impossible to define contributing areas to narrow contour segments from topography. The water balance approach may provide an alternative method for estimating the effective upslope contributing area to narrow contour segments, but does not provide information on its spatial extent.

5.1.2 Spatial and temporal variability of throughflow

At low flows, the spatial distribution of trough throughflow closely followed the distribution of subsurface contributing areas, when the water table was most parallel to

the compacted till layer. At higher flows, the spatial distribution of throughflow shifted, partly as a result of changes in the configuration of the water table which caused changes in flow pathways to individual troughs. At high flows, the water table became more similar to surface topography. However, lateral shunting of water by discrete macropores at high flows overwhelmed topographic control on the trough throughflow distribution.

5.1.3 Models of throughflow variability

At low flows, the observed throughflow distribution conformed best to the distribution of upslope contributing area based on subsurface topography. During low flows, water table configuration upslope of the troughs closely paralleled subsurface topography. At higher flows, water table configuration became more parallel to surface topography. However, the observed trough throughflow distribution did not conform well with the distribution of upslope contributing area based on surface topography. Lateral shunting of water by discrete macropores (e.g., trough 1) at moderate to high flows overwhelmed topographic control on the throughflow distribution. Although none of the models tested accounted for the effects of soil anisotropy, the Dupuit flow model provided the best predictions of throughflow variability at moderate to high flows since it accounted for temporal changes in water table configuration.

5.1.4 Variability of effective hydraulic conductivity

At the trough scale (ca 1 m), estimates of K_{eff} varied over two orders of magnitude. However, for hillslope widths ranging from 5 to 30 m, the results of the present study,

and those of Utting (1979) and Tischer (1986), indicate a range in K_{eff} of less than an order of magnitude. Therefore, a representative scale at which macropore systems are averaged out may be at hillslope widths of approximately 10 m for the coarse-textured soils commonly encountered within the UBC Research Forest.

Vertical variation in K_{eff} for individual troughs was confounded by shifting flow pathways. However, when several troughs were lumped together, K_{eff} could be approximated as a linear function of saturated thickness. Observed deviations from the linear relation at low (<0.10 m) saturated thicknesses may be attributed to increased flow dominance via the permeable organic root mat at the base of the soil profile or increased error in estimating mean saturated thickness from piezometer measurements.

5.2 RECOMMENDATIONS FOR FUTURE RESEARCH

Detailed knowledge of surface or subsurface topography may not provide increased understanding of throughflow pathways, especially at small scales. An important question still to be addressed is whether a relevant spatial scale exists at which the effects of macropore systems may be effectively averaged out. Detailed measurement of the density, orientation, and connectivity of macropore systems (e.g., Noguchi et al., 1997), in combination with measurements of throughflow, may allow determination of representative scales at which macropore systems can be effectively averaged out. The present study indicates that this scale may be at least 5 to 10 m, suggesting that description of throughflow pathways based on topography may be optimal at grid scales of at least 5 to 10 m. However, a conflicting consideration is the need to resolve relevant topographic

features controlling throughflow convergence. Thompson (1993), for example, found that a DEM grid size of no greater than 10 m was required to capture the relevant topographic features controlling throughflow convergence at a nearby site in the Research Forest (Figure 2.2). Further research combining topographic analyses and detailed hillslope flow measurements are required to determine if an optimal scale exists that captures relevant topographic information controlling flow convergence yet averages the effects of macropore systems.

Our current understanding of the temporal dynamics of macropore systems is poorly understood. The present study indicates that macropore systems (e.g., trough 1) are important in controlling the hydraulic properties of forest soils and conveying a dominant portion of flow issued from the plot. However, plot scale soil hydraulic properties would likely change if the macropore system were disturbed or disrupted (e.g., natural deterioration or forest harvesting activities). If a representative scale can be defined at which macropore systems are averaged out, then future research should address how much time is required for the plot to hydrologically recover following disturbance (i.e., K_{eff} before disturbance = K_{eff} following disturbance).

The presence of a highly permeable layer (root mat) along the upper surface of the basal confining layer has been noted in many areas of the UBC Research Forest (e.g., Nagpal and deVries, 1976; Utting, 1979; Tischer, 1986). However, present knowledge on the distribution and genesis of the root mat is still poorly understood. The permeable root mat may transmit a significant fraction of flow downslope during low flow conditions, and may influence the vertical heterogeneity within the soil profile. Detailed investigation

of the distribution of the root mat may provide information on its hydrologic role in forested catchments in southwestern British Columbia.

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