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Runoff generation in a steep, soil-mantled landscape

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[1] Scale and slope dependence of hydrologic response are investigated for two channel network source areas (unchanneled valleys) in the Oregon Coast Range. Observations of response to both natural and applied precipitation reveal that runoff occurred as subsurface flow in which water passed through partially saturated soil, into the shallow fractured bedrock, to emerge as subsurface partial source areas near the channel head. The two dominant approaches to modeling subsurface flow in steep topography, routing of Darcy or fracture flow and the hydrologic similarity approximation of TOPMODEL, respectively predict either a strong slope dependence or no slope dependence to timescales of subsurface runoff generation. Compilation of data from our Coos Bay study sites with observations reported previously elsewhere indicates weak area dependence but no slope dependence in the lag-to-peak and discharge recession constants. This finding supports the interpretation that patterns of antecedent soil moisture and vadose zone characteristics control response times of runoff generation by subsurface storm flow. As slope should influence lateral flow routing once subsurface saturation develops, we conclude that the hydrologic response of steep catchments appears to be insensitive to slope because the controlling timescale is that of the vertical unsaturated flow. *INDEX TERMS:* 1860 Hydrology: Runoff and streamflow; *KEYWORDS:* runoff, streamflow, source area

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1. Introduction

[2] The question of whether subsurface flow can transmit water at rates sufficient to contribute to storm flow has received much attention over the last thirty years. Based on field observations, *Dunne and Black* [1970] concluded that subsurface flow contributed insignificantly to storm runoff in a small, low-gradient catchment. *Freeze* [1972, 1974], using a physically based numerical simulation, showed that storm runoff by subsurface flow requires steep, convex slopes and high saturated hydraulic conductivities; *Beven* [1981] later showed that the required conductivities were consistent with field measurements. Field studies on steep slopes with highly conductive soils document runoff by subsurface storm flow (SSSF) during storm events [*Harr*, 1977; *Yee and Harr*, 1977; *Pierson*, 1980; *Tsukamoto and Ohta*, 1988; *Onda*, 1994; *Montgomery et al.*, 1997] and the proposed mechanisms to explain stream discharge response to storms in steep soil-mantled landscapes include lateral throughflow [*Loudermilk*, 1934; *Hursh*, 1936], lateral unsaturated flow [*Harr*, 1977], macropore flow [e.g., *Bouma et al.*, 1977; *DeVries and Chow*, 1978; *Mosely*, 1979, 1982; *Beven and Germann*, 1982; *McDonnell*, 1990], displacement of old water at the base of a slope by upslope additions of new water [*Hewlett and Hibbert*, 1967; *Martinec*, 1975], and local runoff from streamside areas with high initial soil moisture [*Newbury et al.*, 1969; *Sklash et al.*, 1976; *Sklash and Farvolden*, 1979]. Relatively few monitoring studies of steep hillslopes have directly addressed runoff generation processes [e.g., *Harr*, 1977; *Mosely*, 1979; *McDonnell*, 1990; *Anderson et al.*, 1997], as most focused on piezometric response due to interest in debris flow initiation [e.g., *Swanston*, 1970; *Pierson*, 1980; *Sidle*, 1984; *Petch*, 1988; *Johnson and Sitar*, 1990; *Montgomery et al.*, 1997]. Although mechanisms for generating SSSF remain controversial, runoff generation by SSSF is typically assumed to arise as a perched water table forms at a conductivity barrier at depth, inducing lateral flow to a channel at the base of the slope. In this conceptual framework, storm flow response of mountain streams is thought to be controlled by rapid flow through steep slopes or by shorter travel distances to channels in steeper and therefore more highly dissected terrain.

[3] Rainfall-runoff models treat hillslope runoff production by subsurface storm flow by either explicitly tracking flow routing through hillslopes or through hydrologic similarity arguments [*Wigmosta and Lettenmaier*, 1999]. Flow routing models based on Darcy's law, in which the head gradient parallels the topographic slope, predict that velocity is proportional to slope, which implies a slope dependence to both the lag between rainfall and runoff and to rates of discharge recession [*Zecharias and Brutsaert*, 1988]. In contrast, the hydrologic similarity argument that underlies the widely used TOPMODEL [*Beven and Kirkby*, 1979] does not explicitly account for flow routing through

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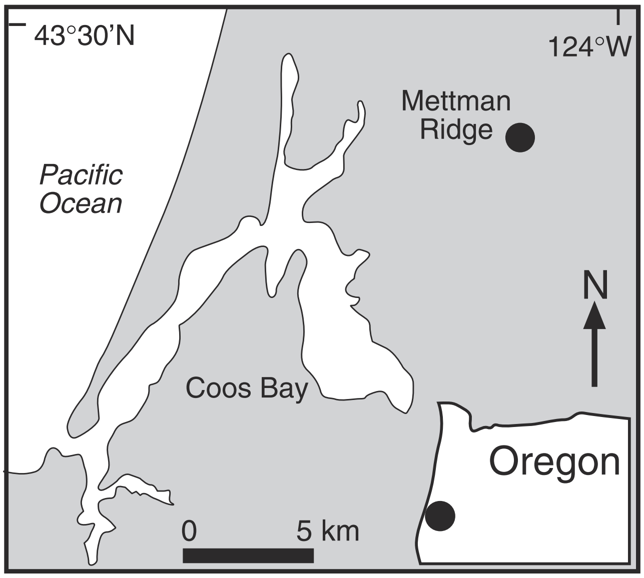

Figure 1. Location map for the Mettman Ridge study area.

hillslopes. Instead, the hydrologic similarity approach assumes instantaneous subsurface hydrologic response, the magnitude of which is based on the topographic index (whether the original ln*A*/tanθ, or the more physically correct ln*A*/sinθ) where A is drainage area per unit contour width and θ is the local slope). This assumption results in no slope dependence to the timing of subsurface runoff generation, although there is a strong slope-dependence to both the location and routing of runoff by overland flow.

[4] Here we discuss hydrograph characteristics for natural storms and sprinkling experiments at two intensively monitored and very steep catchments. In comparing these results to data from other, lower-gradient sites around the world we find the lack of a slope dependence implicit in TOPMODEL better describes the available data. This finding supports the interpretation that runoff generation by subsurface storm flow is dominated by localized subsurface partial source areas fed by vertical unsaturated flow.

2. Study Area

[5] The study area consists of two unchanneled catchments (CB1 and CB2) along Mettman Ridge, roughly 15 km north of Coos Bay in the Oregon Coast Range (Figure 1). In this area, the Coast Range exhibits steep, finely dissected hillslopes; shallow landsliding and debris flows are a dominant geomorphic process in headwater channels. The unchanneled valleys, or hollows, at the head of the channel network undergo a cycle of periodic infilling with colluvium and excavation by debris flows [*Dietrich and Dunne*, 1978; *Dietrich et al.*, 1986], with recurrence intervals greater than several thousand years [e.g., *Reneau and Dietrich*, 1990, 1991]. Highly conductive soils overlying shallow fractured bedrock produce rapid increases in pore pressure in topographic hollows during storms [e.g., *Harr*, 1977; *Pierson*, 1980] and shallow landsliding is a primary mechanism of channel initiation [*Montgomery and Dietrich*, 1988].

[6] The study area includes two small catchments selected after an exhaustive search to find a small, steep channel-head source area where we could conduct sprinkling experiments. The two study catchments are typical in size and slope for source-area basins along Mettman Ridge (Figure 2). The smaller CB1 catchment is an 860 m^2 unchanneled valley with an average slope of 43°; the larger CB2 catchment is a 3,270 m^2 basin with an average slope of 40° adjacent to CB1. Installation of an extensive network of trails, stairs, walkways and platforms at CB1 provided safe work areas and minimized disruption of the colluvial soil during installation of an extensive array of instrumentation [*Anderson et al.*, 1997; *Montgomery et al.*, 1997; *Torres et al.*, 1998].

[7] Both CB1 and CB2 were clear-cut logged in 1987, and replanted with Douglas fir (*Psuedotsuga menziesii*) seedlings in 1989. Soil and bedrock properties are typical of the Oregon Coast Range [*Harr and Yee*, 1975; *Schroeder and Alto*, 1983], with a thin but highly variable soil profile on interfluves [*Schmidt*, 1999; *Heimsath et al.*, 2001] and thicker soils (about 1.4 to 2.0 m) along hollows [*Montgomery et al.*, 1997]. Bedrock in the area consists of relatively undeformed Eocene sandstone [*Beaulieu and Hughes*, 1975] that dips 8° to 17° into the slope. The near-surface sandstone is variably fractured and weathered, although exposures in channel ways lower on the slope reveal massive, relatively impermeable bedrock. The overlying soils are organic rich, low density (1.2 gm/cm^3), high porosity (≈50%), stony sand loams [*Montgomery et al.*, 1997; *Torres et al.*, 1998]. Burrow holes and decayed root

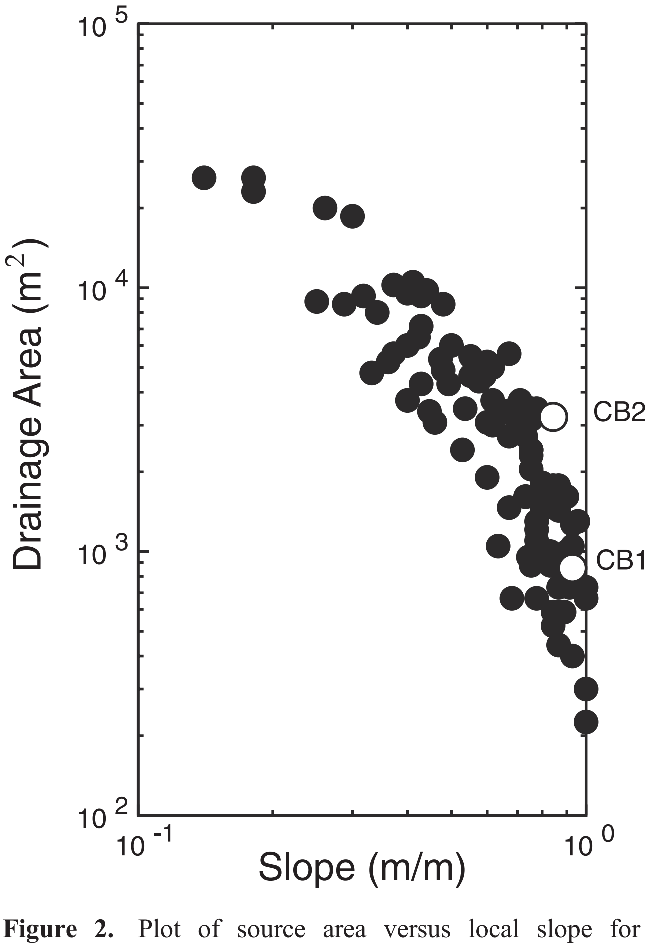

Figure 2. Plot of source area versus local slope for channel heads near Coos Bay. Data are from *Montgomery and Dietrich* [1988, 1992].

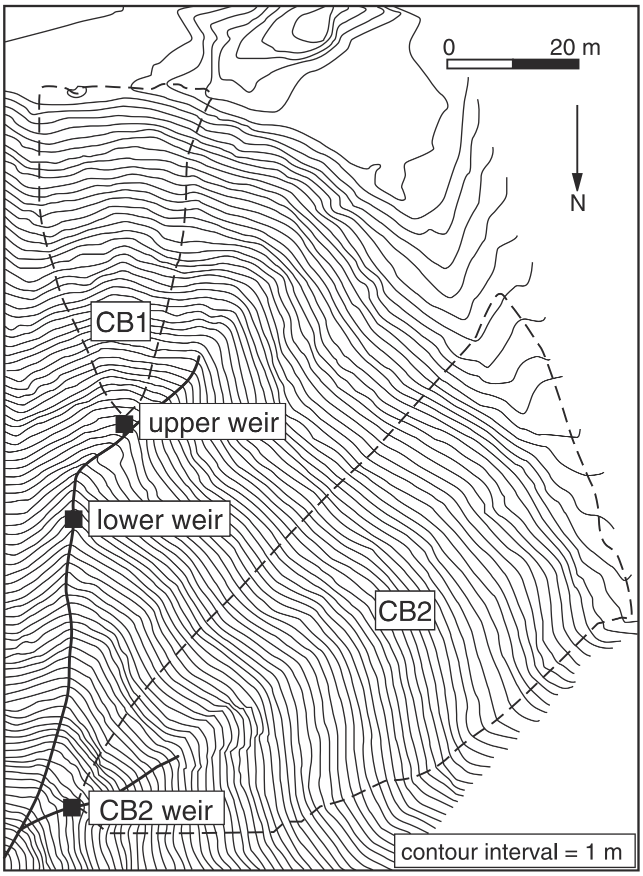

Figure 3. Map showing catchment areas for CB1 and CB2 and the location of the upper, lower, and CB2 weirs (solid squares). Dashed lines indicate ephemeral channels; contour interval is 1 m.

cavities are common. Saturated hydraulic conductivity in the soils ranges from 3.7×10^{-4} to 5.2×10^{-7} m s^{-1}, with a mean of 1×10^{-4} m s^{-1}; saturated conductivity of near-surface fractured bedrock has a comparable mean (4.7×10^{-5} m s^{-1}) but a greater range of values (7.4×10^{-4} to 9.4×10^{-9} m s^{-1}) [*Montgomery et al.*, 2002].

3. Instrumentation and Monitoring

[8] We installed three weirs between 1989 and 1991 (Figure 3). The upper weir was installed in 1989 at the channel head at the base of CB1. The lower weir was installed in October 1991 across an ephemeral, colluvial channel [*Montgomery and Buffington*, 1997] approximately 15 m downslope of the upper weir. A PVC culvert installed in the channel way routed discharge from both the upper weir and the hollow west of CB1 past the lower weir, which consequently recorded only discharge entering the channel between the upper and lower weirs. The third weir was installed along the bedrock channel approximately 15 m downslope of the CB2 channel head; no other instrumentation was installed in CB2. Each weir consisted of a v-notch flume equipped with a stilling well, stage recorder accurate to ±1 mm, and a battery-operated data logger. Coupled discharge and stage measurements obtained during natural rainfall and sprinkler experiments allowed us to generate rating curves for each weir. The CB2 weir was destroyed by a debris flow in February 1992 and the upper and lower CB1 weirs were destroyed by a debris flow in November 1996.

[9] Automated discharge records of natural storms were supplemented by more extensive measurements during some storms. Three automated, tipping-bucket rain gauges recorded rainfall every 10 minutes to the nearest 0.25 mm. Data loggers housed in a shed constructed on the eastern nose of CB1 recorded the response from automated piezometer nests and the upper weir every 10 minutes, and discharge through the CB2 weir every 20 minutes. Data loggers were downloaded every 6 to 8 weeks, at which time the weirs, stage recorders and rain gauges were examined and cleaned if necessary.

[10] Two rounds of sprinkler experiments in May 1990 and in May/June 1992 used an array of rotating sprinkler heads mounted approximately 2 m above the ground surface [*Anderson et al.*, 1997; *Montgomery et al.*, 1997; *Torres et al.*, 1998]. The first experiment applied irrigation at a rate of 1.5 ± 0.7 mm/hr for 6 days; the second experiment sustained irrigation at a rate of 3.0 ± 0.9 mm/hr for 4 days. The third experiment (in 1992) essentially repeated the first experiment. These two rainfall applications, respectively are equivalent to <1 yr and 1 to 2 yr 24-hr events based on the North Bend rain gauge record, the station closest to the study sites. In addition, the first experiment was equivalent to a 1 to 2 yr 6-day rainfall, whereas the more intense second experiment was equivalent to a 15 yr 4-day rainfall [*Montgomery et al.*, 1997].

4. Results

[11] No evidence of overland flow was observed upslope of the channel head at either site during natural or applied rainfall. Continuous data from automated piezometers record that saturation of the soil never reached the ground surface in CB1. Discharge through the upper weir rose within hours of natural rainfall during large storms (Figure 4), whereas small storms (less than about 15 mm day^{-1}) failed to generate increased discharge. Depending upon the size of the storm, discharge declined to pre-event levels within one to several days in the upper weir and more slowly in the CB2 weir. Some storm peaks were truncated in the CB2 weir due to accumulation of debris in the weir. Generation of storm runoff from CB1 requires 24-hr rainfall in excess of 20 mm during a storm with total rainfall exceeding 40 mm [*Montgomery et al.*, 1997].

[12] The extensive experimental monitoring of the site revealed the runoff generation mechanisms at CB1, which are important to understanding the site response time. Because of the high conductivities of the soil and fractured bedrock, full saturation of the soil does not occur here, and consequently all precipitation must first travel through an unsaturated zone. This flow is essentially vertical [*Torres et al.*, 1998], the rate of which is approximately that of the rainfall rate corrected for porosity [*Anderson et al.*, 1997]. Nearly all the soil water drains into the underlying fractured bedrock where subsurface flow develops, and rises back into the soil in places where the bedrock conductivity is low or fractures deliver water under exfiltrating gradients across the bedrock-colluvium contact [*Montgomery et al.*, 1997]. Partial saturation of the soil column occurs at these emergent areas. In one of these areas just above the channel head, flow emerges across less fractured bedrock and drains

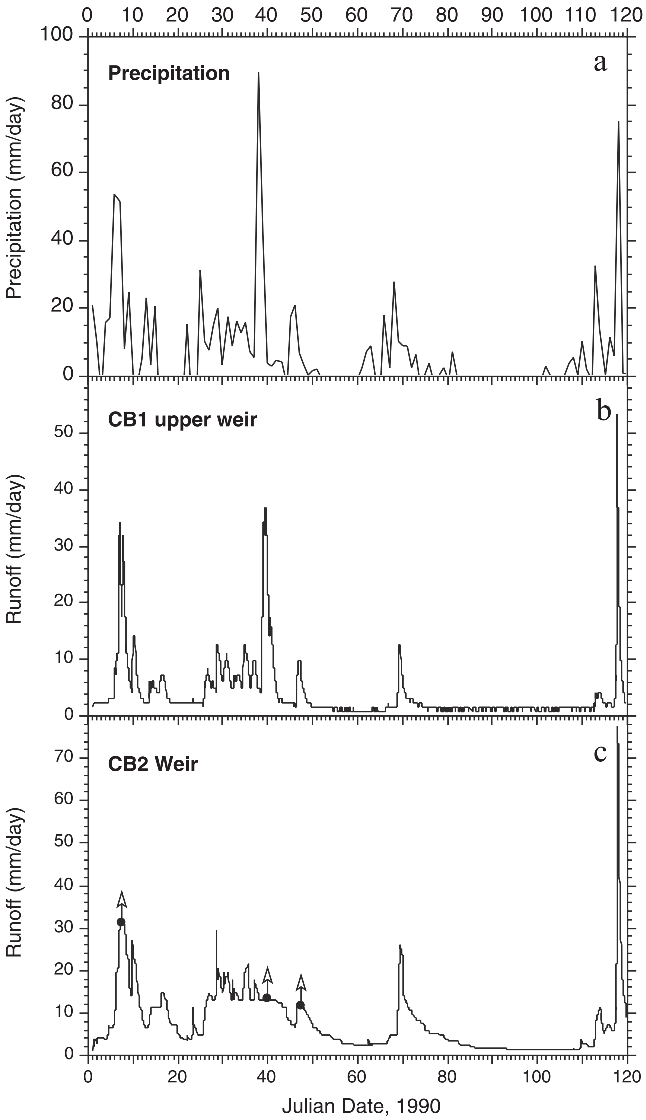

Figure 4. Automated record of (a) rainfall and runoff through the (b) upper and (c) CB2 weirs from January to April 1990. Arrows indicate discharge peaks truncated due to debris accumulation in the CB2 weir.

into the CB1 upper weir. Here vertical unsaturated flow adds to this subsurface flow in the soil, bypassing the bedrock flow path. The depth and lateral extent of this zone of shallow subsurface flow directly above the channel head can be described as a subsurface variable source area contribution to runoff generation [*Anderson et al.*, 1997; *Montgomery et al.*, 1997]. After storm events, the fractured bedrock flow diminishes and reaches near constant levels sooner than the more slowly drained vadose zone, which contributes to an extended hydrograph recession limb.

5. Hydrograph Analysis

[13] Hydrographs integrate the effects of rainfall patterns, runoff generation processes, and hydrologic and topographic properties of the catchment. The discharge recession constant (K_r) and lag-to-peak (L_p) provide simple measures of the timescale of runoff response. Comparison of these hydrograph characteristics for CB1 and CB2 with values reported previously allows comparison of our sites with others where SSSF also dominates runoff generation.

[14] Recession constants, K_r, empirically describe discharge decay by

$$Q_t = Q_p K_r^t = Q_p e^{-\alpha t} \tag{1}$$

where Q_t is the discharge at time t, Q_p is the peak discharge at the start of uninterrupted periods of declining discharge and $\alpha = -\ln K_r$ [*Tallaksen*, 1995]. Equation (1) may be expressed as

$$\ln Q_t = \ln Q_p - \alpha t \tag{2}$$

and α can be calculated from the slope of semi-logarithmic plots of discharge recession. Least squares linear regression of the natural log-transformed discharge versus time was used to calculate α for periods of declining discharge for all storms for which data were available from January 1990 to December 1992, and for our sprinkler experiments.

[15] Recession constants exhibited relatively narrow distributions for each weir (Figure 5). Measuring time in hours, the recession constant for the upper weir ranged from 0.95 to 0.99, with a mean value of 0.98. The CB2 weir had a similar range of K_r values: 0.96 to 0.99 with an identical

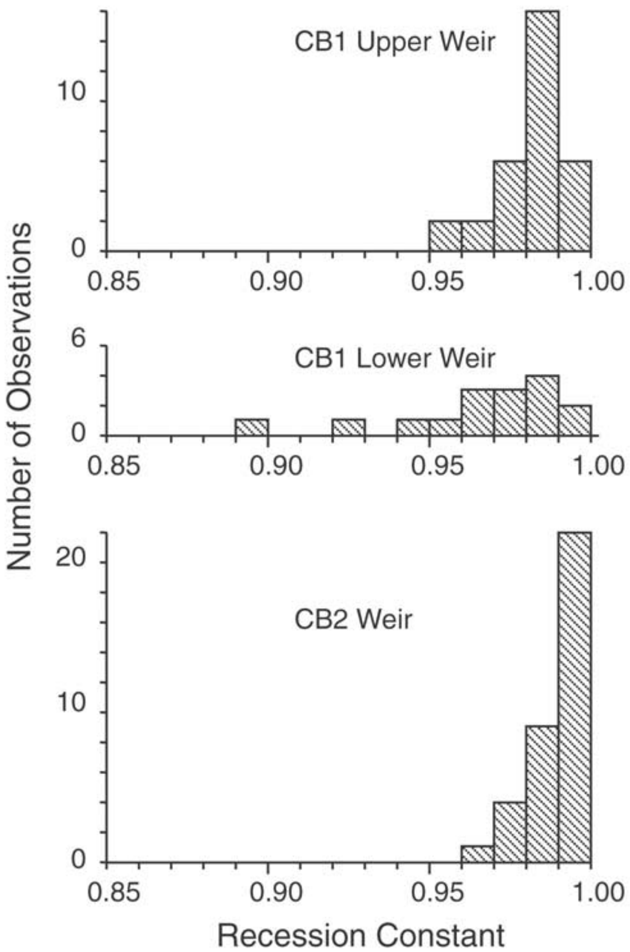

Figure 5. Frequency of discharge recession constants calculated for the upper weir, lower weir, and CB2 weir (see Figure 1 for weir locations).

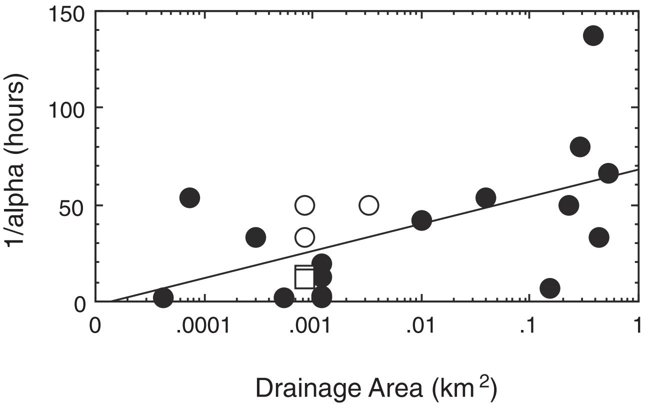

Figure 6. Characteristic recession times ($1/\alpha$) versus drainage area for data from the upper weir, lower weir, and CB2 weir and data for runoff by SSSF compiled from previous studies [*Dunne*, 1978; *Mulholland*, 1993]. Data from storms are represented by open circles, and data from sprinkler experiments are represented by open squares. Least squares linear regression yields ($1/\alpha$) = 68.2 + 14.1 log(DA) [R^2 = 0.32], where ($1/\alpha$) is in hours and drainage area DA is in km^2. Note that ($1/\alpha$) is derived from analysis of time series of discharge Q_t at time t after peak discharge Q_p, following $Q_t = Q_p\ e^{-\alpha t}$.

mean of 0.98. Although the lower weir exhibited a wider range of 0.89 to 0.99, it had a comparable mean value of 0.97. Recession constants calculated from hand-collected discharge measurements during the recession limbs of the sprinkling experiments are slightly lower than those determined for natural rainfall. The K_r value for the upper weir at the end of the first experiment was 0.93, slightly less than the minimum value of 0.95 for response to natural rainfall. Analyses of the recession limb for the upper and lower weirs from the second experiment yielded K_r values of 0.92.

[16] *Dunne* [1978] showed that K_r values cluster in different ranges for different runoff generation mechanisms. Low K_r values for Horton overland flow (0.02–0.34), indicate rapid discharge recession, whereas SSSF yields higher K_r values (0.27–0.99) that imply sustained drainage and prolonged discharge decay. Saturation overland flow hydrographs exhibit a wider range of K_r values (<0.01–0.94) due to the influence of SSSF on sustaining return flow during discharge recession. Comparison of the characteristic recession time given by $1/\alpha$ with those derived from recession constant values compiled by *Dunne* [1978] and reported by *Mulholland* [1993] reveals a weak correlation with basin size (Figure 6). In spite of the very steep slopes at CB1 and CB2, $1/\alpha$ values for all three weirs lie at the high end of the reported range for runoff by SSSF in low-gradient catchments, and show that for their size CB1 and CB2 exhibit relatively slow discharge decay.

[17] The lag to peak, as characterized by the time elapsed from when half the storm rainfall has fallen to the peak discharge, reflects the time required to route storm runoff through a catchment. We determined L_p values for the upper weir and CB2 weir from 10-minute rainfall and discharge records for storms prior to the May 1990 sprinkler experiments. The L_p estimates for the upper weir ranged from 2.3 to 10.7 hr, with a mean value of 6.2 hr. The L_p for the CB2 weir ranged from 6.2 to 22.1 hr, with a mean value of 13.4 hr. Among basins of comparable size, these L_p values for CB1 and CB2 are at the high end of the range for runoff by SSSF compiled by *Dunne* [1978], reported by *Weyman* [1970], *Cheng* [1988], *Allan and Roulet* [1994], and *Peters et al.* [1995], and derived from data and maps presented by *Anderson and Burt* [1978], *Pilgrim et al.* [1978], *Mosely* [1979, 1982], *Anderson and Kneale* [1982], *Swistock et al.* [1989], *McDonnell* [1990], *McDonnell et al.* [1991], *Mulholland et al.* [1990], *Hinton et al.* [1994], and *Wilcox et al.* [1997]. Data for both the upper weir and CB2 weir fall near the upper end of the range of values reported for SSSF at sites of comparable size.

[18] The composite data sets reveal a weak positive correlation between L_p and contributing area, A, and no relation between L_p and basin slope (Figure 7). Least squares linear regression of the natural log transformed data yields

$$L_p = 15 A^{0.18} \qquad \left(R^2 = 0.40\right) \tag{3}$$

where L_p is in hours and A is in square kilometers. Comparable regression of slope and L_p data yields no relationship (R^2 = 0.03). The compiled data reveal that contributing area

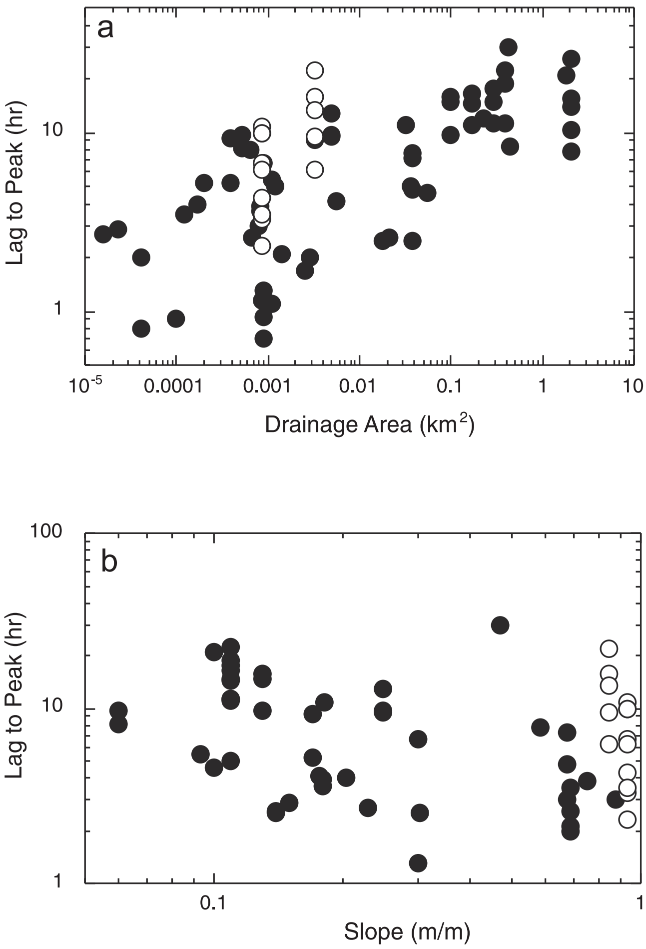

Figure 7. Mean lag to peak versus (a) drainage area DA [logLp = 1.18 + 0.18 logDA; R^2 = 0.40] and (b) average basin slope S [logLp = 0.68 − 0.23 log$S$; R^2 = 0.04] for data from the upper weir, lower weir, and CB2 weir (open circles) and data for other sites for runoff dominated by SSSF compiled by *Dunne* [1978] and from studies referred to in the text (solid circles).

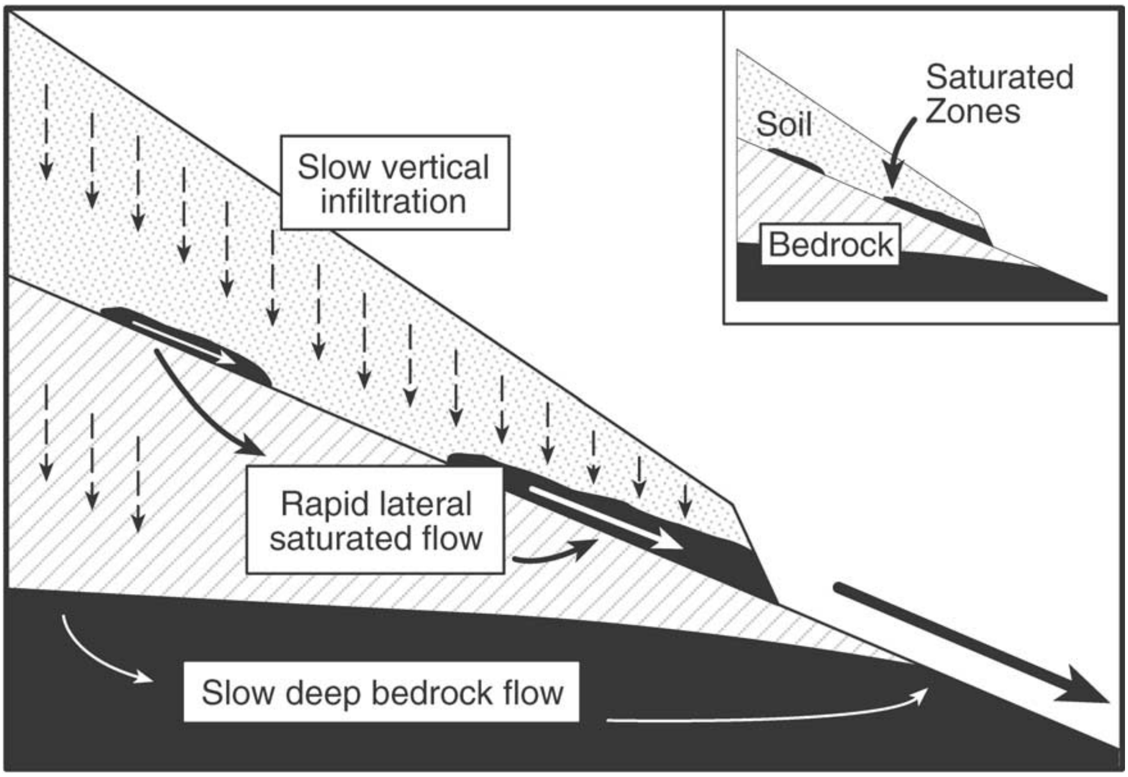

Figure 8. Runoff generation at CB1 is controlled by slow vertical infiltration that feeds rapid lateral flow from subsurface saturated source areas that govern runoff generation. Rapid flow in near-surface fractured bedrock contributes to storm flow, and unsaturated drainage and deeper bedrock flow contribute to maintaining base flows.

influences the lag to peak for runoff by SSSF, whereas hillslope gradient exerts no discernable influence.

6. Discussion

[19] In spite of extreme slopes and high saturated conductivities, discharge from CB1 and CB2 responds over timescales similar to SSSF from low-gradient sites of comparable size. We see no evidence that slope is a primary control on rates of runoff generation by SSSF, a finding that directly contradicts the prediction of simple Darcy flow routing in which one would expect that the lag to peak should increase and recession should be slower with increasing drainage area and decreasing topographic gradient. Instead we find the lack of slope dependence assumed by the hydrologic similarity approach of TOPMODEL.

[20] While the lack of a discernable influence of slope may suggest tremendous variability in hydraulic conductivity and other site-specific properties in the compiled data, we note that relations between basin size and both $1/\alpha$ and L_p hold in spite of such variability.

[21] We find it counter-intuitive that discharge recession in steep headwater areas appears no faster than in lower-gradient catchments of similar size. Simple dynamic models of subsurface runoff (such as that by *Iida* [1984]) yield area and slope dependencies to L_p if vadose zone flow is ignored, saturated conductivities are spatially constant, and the head gradient is assumed to be equivalent to the surface topography. The lack of slope dependence in the available field data may arise from local variability in soil or underlying bedrock conductivity, which could overwhelm the expected relationships for a small data set. However, we suspect that the apparent slope independence of SSSF runoff timing reflects both the effects of systematic spatial variability in antecedent soil moisture and vadose zone dynamics.

[22] Our prior observations at CB1 [*Anderson et al.*, 1997; *Montgomery et al.*, 1997; *Torres et al.*, 1998] led us to interpret our results as supporting the concept of a subsurface partial support area controlled by antecedent soil moisture and infiltration: a subsurface flow analog to the concept of partial source areas for runoff generation by saturation overland flow (Figure 8). Intuitively, a slope dependency for this mechanism is implied in that areas with higher topographic index values (i.e., larger drainage area and lower slope) correspond to areas with greater antecedent soil moisture, and therefore greater propensity to generate runoff due to interaction with infiltrating flow. The resolution to this conundrum lies in that the moisture-dependent conductivity properties, which control evolution of negative pressure fields and therefore vadose zone flow, are not slope dependent because unsaturated flow is essentially vertical. Given that the soil never was fully saturated at CB1, we know that all runoff from these sites must pass through an unsaturated zone which has time lags set not by slope but by the soil moisture retention curve [*Torres et al.*, 1998].

7. Conclusions

[23] The relatively slow runoff generation revealed by our studies at the steep Coos Bay catchments demonstrates the

inadequacy of simple models of lateral saturated flow for describing runoff generation by SSSF. Our findings highlight that the storm flow response typical of steep terrain is fed by SSSF that is no faster than on lower gradient slopes. The non-slope-dependent vadose zone response, through which all precipitation at the study sites must pass on the way to becoming runoff, strongly damps any slope dependence to the response time. This suggests that hydrologic models concerned with timing of response must include some relevant physics of the unsaturated zone, including effects when the zone is well below saturation. Advective models that include a strong lateral travel time for runoff generation will tend to incorrectly predict a strong slope dependency in runoff generation by SSSF. This issue is particularly relevant to ongoing attempts to develop coupled hydrologic and slope stability models such that the timing and location of landslides can be anticipated.

[24] **Acknowledgments.** This project was supported by the Weyerhaeuser Company, and National Science Foundation Grants NSF EAR89-17467 and CMS96-10269. John T. Heffner built much of the project infrastructure and helped periodically download data loggers. Ray Torres, Suzanne Anderson, and Keith Loague collaborated on the Mettman Ridge sprinkler experiments. Kate Sullivan was instrumental in developing support for this project and Jim Clarke provided logistical support crucial to its success. Tamara Massong, Darryl Granger, Matt Coglon, Chris Brummer and Juan Somoano helped survey the study areas; Jianfei Yu calculated recession constants and total runoff coefficients for natural storms; and Harvey Greenberg provided GIS support. We are also grateful to Kyle Comanor for sharing insights into TOPMODEL.

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