SCIENTIFIC BRIEFING

HYDROLOGICAL PROCESSES
Hydrol. Process. **17**, 1457–1467 (2003)
Published online in Wiley InterScience (www.interscience.wiley.com). DOI: 10.1002/hyp.5121

Modelling effects of forest canopies on slope stability

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Abstract

We investigated the potential effects of rainfall intensity smoothing by forest canopies on slope stability by modelling soil responses to measured rainfall and throughfall during high-intensity rain. Field measurements showed that maximum intensities of precipitation were generally reduced under forest canopies at two sites in the Pacific Northwest, USA. Modelling soil water pore-pressure responses of a hypothetical hillslope to the field data resulted in estimates of slope stability that were generally greater under forest canopy than for the same hillslope without forest canopy. Results indicate that smoothing of precipitation intensities may translate into overall greater stability of hillslopes under forest canopies. Copyright © 2003 John Wiley & Sons, Ltd.

Key Words landslides; slope stability; canopy interception; infiltration; hillslope hydrology

Introduction

Stability of slopes in steep, forested landscapes is an important issue for land management and ecosystem science. Forest disturbances, especially logging, have long been recognized as contributing to landsliding rates, especially shallow, rapid mass movements originating on steep slopes at the head of stream channels (Sidle *et al*., 1985). These slides occur when infiltration exceeds hillslope drainage during periods of extended precipitation (Van Asch *et al*., 1999), and resulting high pore pressures cause soil strength to become insufficient to resist downslope movement. This type of slope failure often initiates a debris flow (Iverson *et al*., 1997) that can be quite destructive. For example, in the Pacific Northwest, USA, debris flows dominate sediment budgets (Dietrich and Dunne, 1978; Swanson *et al*., 1982) and are hypothesized to be strong controls on aquatic habitat in many steep landscapes (Reeves *et al*., 1995). Understanding the processes that control slope stability is, therefore, an important goal, especially as it pertains to land management.

Researchers have commonly concluded that the most important effect of forest disturbances on slope stability is loss of cohesive soil strength when reinforcing roots die and decay (Burroughs, 1984; Sidle, 1992; Dietrich *et al*., 1995). This paradigm originated from investigations that revealed root density and strength declines following logging (Burroughs and Thomas, 1977; Ziemer, 1981). Confirmatory evidence includes field observations that total root biomass and strength are less in failed slopes than in unfailed slopes (Wu *et al*., 1979; Buchanan and Savigny, 1990; Schmidt *et al*., 2001), and controlled experiments that have found

Received 19 November 2002
Accepted 19 February 2003

strength of soil in direct shear increases with rooting density (Waldron and Dakessian, 1981, 1982; Wu *et al*., 1988).

There are, however, at least three reasons to doubt that root strength is the exclusive way that forests increase slope stability. First, natural spatial variability of rooting is high and may covary with other factors affecting stability, such as soil morphology or hydrology, so that failed slopes sometimes contain more roots than unfailed slopes (Schmidt *et al*., 2001). Second, the infinite slope model most commonly employed for analyses (Wu *et al*., 1979; Sidle and Swanston, 1982; Sidle *et al*., 1985; Iverson *et al*., 1997) is a simple force-balance model that is strictly inapplicable to the inhomogeneous forest soils on concave slopes typical of landslide initiation sites, and is, in any case, difficult to apply to natural slopes because several critical parameters can only be measured with low precision. Analyses based on the infinite slope model, therefore, result in stability estimates of insufficient precision to deduce either the causative mechanisms of individual slides or factors limiting the stability of unfailed slopes (Sidle *et al*., 1985). Third, previous estimates of the contribution of roots to soil strength may be too high. Direct measurements of root rupture strength and tests of rooted soils at large deformations measure strengths that likely do not come into play during failure of natural forest soils (Skaugset, 1997), because only a portion of total root strength may be mobilized before failure in the field: small roots fail at lower deformations than do large roots (Waldron and Dakessian, 1981), and loose forest soils fail at smaller deformations than do most roots (Waldron *et al*., 1983).

These observations suggest that the reasons for the correlation between disturbance and increased incidence of landslides are more complicated than simple differences in soil strength afforded by root reinforcement. Although roots most likely strongly influence the stability of some slopes, the extent of this condition is unclear. Indeed, for some slopes, other factors must control stability. For example, why do some slopes fail during a given storm while some similar slopes with lower measured root strength remain unfailed in the same storm (Schmidt *et al*., 2001)? Ascribing all effects of vegetation on slope stability to effects on soil strength ignores other potentially relevant interactions among forests, hydrology, and soil strength.

One important effect of forests on hydrology of vadose zones is on the amount and timing of precipitation reaching the soil. Quantities of throughfall are generally less than precipitation because water evaporates from forest canopies (Zinke, 1967), and there is a general damping and smoothing of intensities in throughfall compared with precipitation (e.g. Rutter *et al*., 1971; Massman, 1983; Schellekens *et al*., 1999; Xiao *et al*., 2000). Evaporation loses prominence during large (landslide-producing) storms because rainfall far exceeds the 5–10 mm sufficient to wet canopy surfaces. Beyond this point, evaporation rates are small compared with rainfall rates (Rutter and Morton, 1977; Klaassen *et al*., 1998), and smoothing of intensity becomes the dominant effect of the canopy on precipitation.

Most precipitation transfers some or all of its momentum to the canopy by striking vegetative surfaces; this loss of momentum temporarily detains precipitation, so that throughfall is composed largely of water dripping from canopy storage. The rate that stored water is released as drip is then controlled by momentum of precipitation and stored water and by interfacial tension between stored water and canopy surfaces (see discussions by Jackson (1975) and Massman (1983)). The net effect of all these interactions between canopies and precipitation is to reduce the delivery rates of intense precipitation and release stored water over time. Direct measurement of canopy storage during storms has revealed temporarily increased storage during elevated rainfall intensity (Calder and Wright, 1986; Klaassen *et al*., 1998), which is consistent with the momentum transfer mechanism of intensity smoothing. This phenomenon refutes the concept that canopy storage has an absolute maximum capacity beyond which all precipitation is transferred to the ground.

The smoothing effect that forests have on precipitation rates may be relevant to slope stability on steep, topographically convergent slopes where soils are shallow. Infiltrating precipitation dominates water budgets on these sites, so shallow, rapid landslides are triggered by groundwater fluctuations that occur over short time scales. For example, Iverson (2000) estimated that the relevant time scale of response is ~20 min at the well-studied Mettman Ridge hollow in the Oregon Coast Range, USA (Anderson *et al*., 1997; Montgomery *et al*., 1997; Torres *et al*., 1998; Montgomery and Dietrich, 2002). Smoothing of precipitation intensities by forest canopies may be sufficient to affect groundwater fluctuations at such short time scales, so that slope stability is maintained by the presence of the canopy.

The purpose of this paper is to present field data on the smoothing effects of forest canopies on rainfall intensities at the ground surface, and to present analyses that suggest these effects are sufficiently large to warrant investigation as a potential control on landsliding. Our approach to studying this phenomenon is to focus on the effects of canopies on precipitation and use a model to infer likely responses of soil water. A comparison of measured responses of hillslopes to precipitation and throughfall would be most desirable, but variability in soil physical properties makes such comparisons difficult. The overall objective of this research is to assess whether canopy effects on precipitation may be important to slope stability in a general sense.

Methods

Field data

We collected rainfall intensity data in and adjacent to two forest stands in the Pacific Northwest, USA. Each stand was equipped with a tipping-bucket rain gauge in a large nearby (<200 m away) opening, and multiple throughfall gauges randomly located under the canopy. The tipping-bucket gauges were calibrated to tip once per 0·254 mm of rainfall. The throughfall gauges consisted of similar tipping-bucket gauges augmented with two troughs to increase the spatial extent of sampling. The troughs were plastic pipes with a slot cut in each that was 19 mm wide and 2 m long. Each pipe was set at a 22·5° angle to the horizontal (slightly less than the funnel in standard rain gauges) so that travel time to the tipping bucket was minimized and correct intensity data were recorded. The area of the two troughs projected to the horizontal exactly doubled the catch area of the gauge, so that each tip recorded 0·127 mm of rain.

One stand was in the Cedar Flats Research Natural Area of the Gifford Pinchot National Forest, Cascade Mountains, southwestern Washington, USA. This stand originated about 1400 AD (600 years ago); overstory trees are Douglas-fir (*Pseudotsuga menziesii*), western redcedar (*Thuja plicata*), and western hemlock (*Tsuga heterophylla*) up to 3·1 m diameter and 84 m tall. The canopy is structurally complex because of its age. Senescent large trees have created gaps, and the spatial relationships between canopies of very large trees and younger individuals of tolerant species result in high intracanopy heterogeneity of canopy cover. We placed six throughfall gauges in this stand to sample across the spatial heterogeneity of throughfall.

The other stand was in the Dunn Research Forest of the Oregon State University College of Forestry, on the eastern margin of the Oregon Coast Range. This stand originated after a clearcut in 1940 (60 years ago), and was thinned lightly 30 years later. The overstory trees are Douglas-fir, grand fir (*Abies grandis*), and bigleaf maple (*Acer macrophyllum*) up to 0·8 m diameter and 43 m tall. The canopy is spatially quite homogeneous. We placed three throughfall gauges in this stand.

Unpublished semivariograms from spatially intense throughfall sampling revealed that small-scale (<10 m) spatial variance was completely captured by the throughfall gauges at both sites. Differences among gauges at the old-growth site suggests there was important variance at larger spatial scales attributable to canopy gaps (Spies and Franklin, 1989). There was no systematic variation among throughfall gauges at the young-growth site because there were no substantial gaps in the canopy.

We did not measure stemflow in either stand because we did not expect it to contribute substantially to throughfall during periods of high rainfall intensity. Our unpublished spatially intense throughfall sampling revealed a zone of concentrated throughfall near stems in the young stand, but not in the old stand. Observations during storms revealed no obvious stemflow, consistent with Rothacher's (1963) findings that stemflow is negligible on the rough bark of old trees in the region. Swank (1972) found substantial stemflow in a stand of small Douglas-fir, but Knutsen (1965) and Iroumé and Huber (2002) found stemflow of 7% of total precipitation for Douglas fir stands that were 27–35 years old. From these published values, we estimate stemflow to be <5% of gross precipitation in both stands.

Soil water and slope stability models

Iverson (2000) modelled pore pressures in hillslopes to predict landsliding triggered by infiltration of non-steady rainfall. The model produces one-dimensional solutions for pore pressure ψ after non-dimensionalizing a three-dimensional Richards equation by spatial and temporal scales appropriate to large, slow-moving landslides (ignored here) and shallow, fast-moving landslides (applicable here). The result is that $\psi(x, y, z, t)$ becomes $\psi\ (Z, t)$ at (x, y) of a potential failure, where Z is depth normal to the slope. The normalization is embodied by a response function R characterizing the response of $\psi(Z, t)$ according to soil properties. After solving the Richards equation for ψ using appropriate initial and boundary conditions, the response function R is a function of time, slope, and soil depth and diffusivity. A final and important assumption of the model is that the soil is initially in a state of near-zero ψ, so that hydraulic conductivity and diffusivity can be represented by a single value. This assumption is justified by the observation that landsliding occurs when the soil is extremely wet. The relationship between pore pressures and unsteady rain rates then becomes

$$\frac{\psi}{Z}(Z, t \leq T) = \beta\left(1 - \frac{d}{Z}\right) + \frac{I_z}{K_z}[R(t^*)]$$

and

$$\frac{\psi}{Z}(Z, t > T) = \beta\left(1 - \frac{d}{Z}\right) + \frac{I_z}{K_z}[R(t^*) - R(t^* - T^*)] \quad (1)$$

where

$$t^* = \frac{t}{Z^2/\hat{D}} \text{ and } T^* = \frac{T}{Z^2/\hat{D}} \quad (2)$$

are normalized times;

$$R(t^*) = \sqrt{t^*/\pi}\exp(-1/t^*) - \operatorname{erfc}(1/\sqrt{t^*}) \quad (3)$$

is the response function of the soil; θ is the volumetric water content; T is the duration of a pulse of constant-intensity rainfall; β is a constant relating to groundwater flow (we assumed slope-parallel flow here, so that $\beta = \cos^2\alpha$); α is the hillslope angle; d is the depth to the steady-state water table; I_z is the infiltration at the soil surface; K_z is the hydraulic conductivity; D_0 is the maximum characteristic diffusivity, equal to K_{sat}/min $(d\theta/d\psi)$; $\hat{D}$ is the effective hydraulic diffusivity, equal to $4D_0\cos^2\alpha$; and erfc is the complimentary error function.

The total response of soils to unsteady rainfall is obtained by linear superposition of the responses to units of rainfall broken down into periods where intensity I_z is constant. In application, we assumed I_z to be constant between bucket tips in the rain gauges.

We calculated slope stability resulting from modelled time-varying pore pressures using the infinite slope equation in the form presented by Iverson (2000):

$$FS = \frac{\tan\varphi}{\tan\alpha} + \frac{-\psi\gamma_w\tan\varphi + c}{\gamma_s Z\sin\alpha\cos\alpha} \quad (4)$$

where FS is the factor of safety, φ is the soil friction angle, γ_w is the unit weight of water, γ_s is the unit weight of soil, and c is the soil cohesion.

Infinite slope analysis assumes slope failure occurs by Mohr–Coulomb criteria (Das, 1998) when frictional stresses resisting slope movement are balanced by downslope gravitational stresses, and $FS = 1$. To model slope stability through storms, we held all terms in Equation (4) constant except the pore pressure ψ, which we calculated by Equation (1).

Analyses

No storms during the study period were large enough to cause regionally widespread landsliding, from which we infer that soils were never near saturation as required to satisfy the assumptions of the soil-water model. Therefore, instead of using observed storms in their entirety for modelling pore pressures and slope stability, we assumed that the periods of highest intensity rainfall within storms we measured in the field could reasonably be expected to occur during a hypothetical large, landslide-producing storm. The idea was to treat the measured rainfall/throughfall data as a case study of a portion of a hypothetical storm large enough to meet the requirement of the soil model.

We selected portions of two storms from each site for comparison of rainfall intensities outside the stand and under the canopies. The time periods selected were all after precipitation was sufficient to wet canopies well in excess of storage capacities, so the effects on momentum of rainfall should be most purely expressed. However, we did not measure canopy storage or evaporation, so cannot be sure that all canopy surfaces were wetted. The behaviour of throughfall during these high-intensity bursts was generally consistent with most storms during the study, but these were the highest instantaneous intensities (minimum time between tips) measured for the study. We binned data from the tipping buckets into 1 min periods, and used these data to drive the soil pore-pressure model, evaluating at time steps of 1 min.

To compare modelled pore pressure and slope stability under rainfall and throughfall during the periods of high-intensity rainfall, we analysed the modelled responses of a hypothetical soil for which rates of drainage approximated the rates of precipitation we measured. We began with soil characteristics similar to those reported by Montgomery *et al.* (1997) and Torres *et al.* (1998) for the Mettman Ridge site in the Oregon Coast Range (Table I), then adjusted conductivity and diffusivity until the model predicted pore pressures to respond to storms rather than drain rapidly or accumulate excessive water. Because we did not measure rainfall during storms large enough to produce landslides, rainfall measured in the field was never sufficient to balance drainage with raw soil parameters, so we used hydraulic conductivity one order of magnitude lower and diffusivity one order of magnitude higher than field estimates. These parameters are arbitrary, but are within natural variation and also resulted in modelled responses that we judged to be realistic. Most trial parameter sets resulted in modelled soil water responses to precipitation and throughfall that compared similarly, although the overall magnitudes of responses varied.

Results

In general, peak intensities of throughfall were damped in intensity and lagged in time relative to peak intensities of rainfall (Figures 1a and b, and 2a and b). Delays in peaks in the young stand were 0–2 min and delays in the old-growth stand were 0–4 min. Peak instantaneous intensities were damped by 21–52% in the young stand and by −31–83% in the old-growth stand.

Responses were spatially more consistent in the young stand than in the old stand, because all throughfall collectors in the young stand were under a homogeneous closed canopy. In the old stand, some throughfall collectors were in gaps (Spies and Franklin, 1989) so that throughfall closely approximated rainfall; conversely, throughfall under the largest trees differed most strongly from rainfall. Notable deviations from the trend of damped throughfall were at two of the throughfall gauges in the old forest, which are in gaps (Figure 2b). These two gauges experienced peak throughfall intensities that exceeded peak rainfall intensity in one storm, albeit lagged by up to 2 min. Throughfall at these gauges may have been augmented by some other phenomenon, such as a gust of wind releasing stored water from the canopies of adjacent trees. The rainfall–throughfall relationships during this storm were not typical of the data, but we present them here to illustrate the greater spatial and temporal variation in throughfall that occurs in the old stand.

Total throughfall was less than rainfall in all high-intensity rainfall spikes we analysed, but rainfall and throughfall were more nearly equal in the 30 min periods following each spike. Stand-average throughfall was 67–84% of rainfall during the high-intensity spikes and 75–242% of rainfall in the 30 min following each spike. Throughfall was therefore 72–108%

Table I. Values of soil parameters used to model hillslope pore pressures during rainfall. Values for Mettman Ridge (a well-studied hollow in the Oregon Coast Range) are provided for comparison

Parameter	Symbol	This analysis	Mettman Ridge (source)
Hillslope (°)	α	45	43 (Montgomery *et al.*, 1997)
Hydraulic conductivity (m s^{-1})	K_z	1×10^{-5}	10^{-4}–10^{-3} (Montgomery *et al.*, 1997; Pierson, 1980)
Diffusivity ($m^2 s^{-1}$)	D_0	1×10^{-2}	$\sim 10^{-3}$ (Montgomery *et al.*, 1997; Torres *et al.*, 1998)
Depth to free water surface (m)	d	1	varies (Montgomery *et al.*, 1997)
Soil depth (m)		1	0.5–2 (Montgomery *et al.*, 1997)
Soil friction angle (°)	φ	38	33–41 (Schroeder and Alto, 1983; Dietrich *et al.*, 1995)
Unit weight of water (N m^{-3})	γ_w	9800	
Unit weight of soil (N m^{-3})	γ_s	15 700	15 700 (Montgomery *et al.*, 2000)
Soil cohesion (Pa)	c	10 000	10 000 (Montgomery *et al.*, 2000)

of rainfall for the time period containing the spikes and following 30 min. We infer that water storage increased during the high-intensity spikes, then drained after rainfall intensity decreased. There was no evidence that interception loss was more important for canopy smoothing of intensity spikes in summer storms: throughfall was the lowest percentage of rainfall during and after the rainfall spike in the January storm in the young stand, even though total storm interception loss was greatest from the August storm in the same stand.

Damping and lagging of rainfall intensity by canopies at both sites generally increased modelled slope stability relative to the opening (Figures 1c and d, and 2c and d) by reducing the transitory build-up of pore pressure. The exceptions were the two gauges that experienced higher throughfall intensities than rainfall in this opening, and a third gauge during that storm, where there was more total throughfall than rainfall during the burst. For the soil parameters we used, slope failure would have occurred during one storm in the old stand at the no-canopy site and at three of the six under-canopy sites (Figure 2c). The lowest factor of safety obtained in any of the simulations was at the no-canopy site for this storm.

Discussion

Observed intensity smoothing by both forest canopies was sufficient to be a potentially important process in hydrology and stability of natural slopes. However, this process must be viewed in a complicated hydrological and ecological context.

The results of this research follow previous findings that canopy storage increases with rainfall intensity (Calder and Wright, 1986; Klaassen *et al.*, 1998). Impact and accumulation of precipitation on canopy surfaces dictates a loss of precipitation momentum

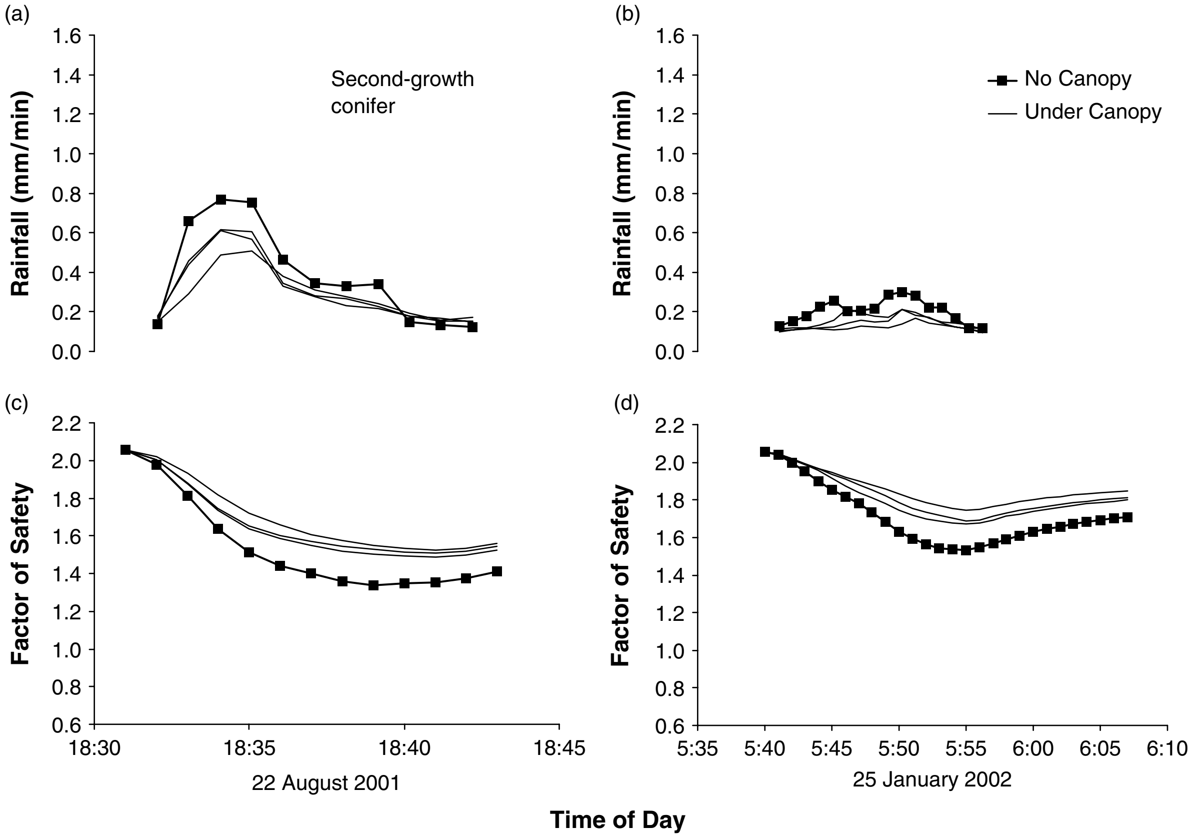

Figure 1. Observed rainfall and throughfall (a and b) and corresponding modelled stability of a hypothetical 45° hillslope with soil 1 m deep (c and d). Slope failure occurs theoretically at FS = 1. Data are from a 60-year-old stand of Douglas-fir in western Oregon

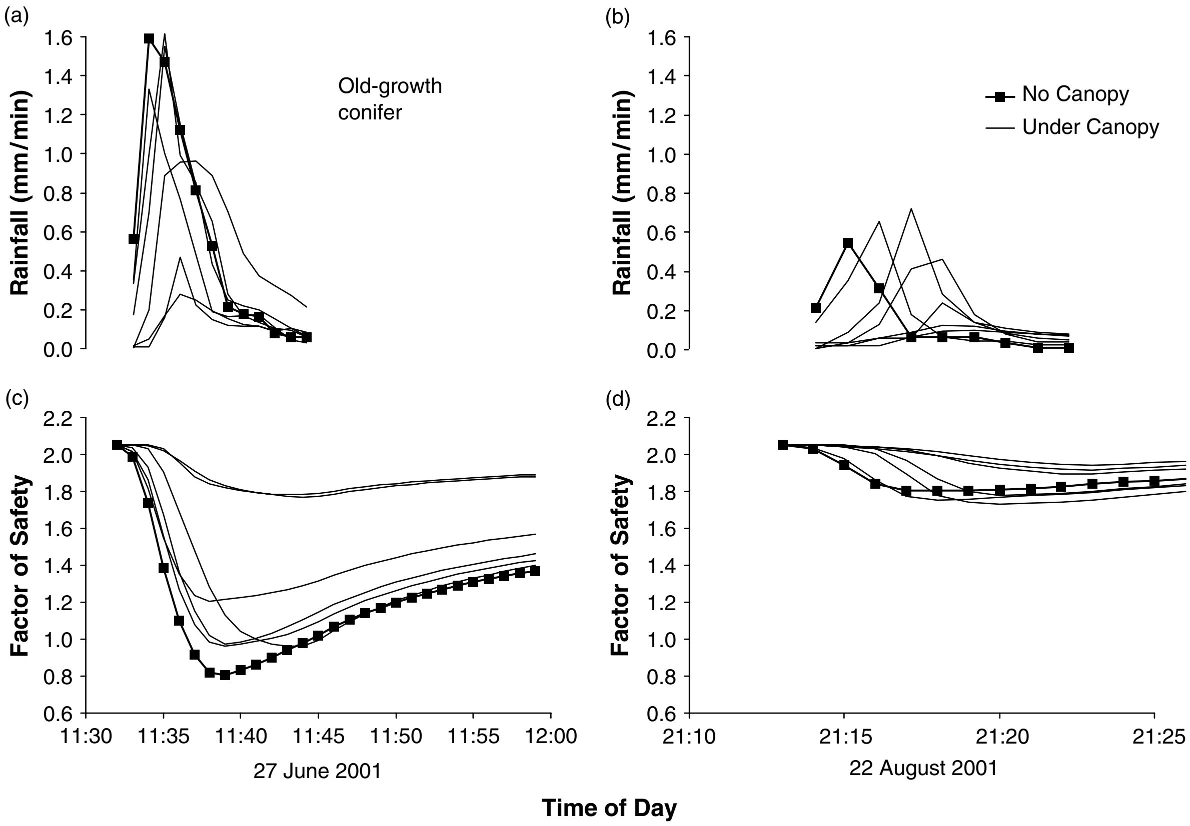

Figure 2. Observed rainfall and throughfall (a and b) and corresponding modelled stability of a hypothetical 45° hillslope with soil 1 m deep (c and d). Slope failure occurs theoretically at FS = 1. Data are from a 600-year-old stand of conifers in southwestern Washington. The three highest maximum throughfall intensities of each storm occurred in canopy gaps

(Grubin, 1963), resulting in general smoothing of intensity by canopies and attenuation of high-intensity rainfall bursts. Bucket models of canopy storage (e.g. Liu, 1997) cannot explain this phenomenon. We suggest a more appropriate conceptual model of dynamic canopy storage, in which the traditional 'canopy storage capacity' is termed 'static canopy storage capacity'. We see static canopy storage capacity as analogous to soil field capacity, defined as the storage held primarily by static interfacial forces between canopy surfaces and stored water. Storage and release of additional water exceeding static capacity is, by contrast, controlled more by momentum of precipitation. Limits on this 'dynamic storage' are indefinite, and depend on the total balance of momentum within the canopy. External inputs of momentum into the canopy, such as wind, may preferentially effect drainage from dynamic storage, but may also overwhelm interfacial tensions and release some water held in static storage. High-intensity rainfall may also wet new canopy surfaces to increase static storage, as water is more likely to splash or flow along canopy surfaces to areas not directly exposed to precipitation (Herwitz, 1987). We infer that the storms we sampled were sufficient to satisfy static storage capacity, and the intensity smoothing observed during high-intensity bursts of rainfall was mainly an effect of dynamic storage.

By using Iverson's (2000) model of landslide initiation, we have implicitly assumed that failures are caused by precipitation infiltrating through the unsaturated, homogeneous soil matrix to reach a water table below. This simplification of complex hillslope hydrology may be inappropriate when and where other hydrologic controls of pore pressure and slope stability are important. For example, Reid (1997) found that inhomogeneities in hydraulic conductivity reduce slope stability by causing locally increased pore pressures. A potentially important complication is preferential flow in macropores, which can cause local concentrations of pore pressure (Pierson, 1983; McDonnell, 1990) and may usurp any effects of infiltration through the soil matrix. If the soil matrix is the main conduit for infiltrating water, then fluctuating intensity may cause higher pore pressures low in the soil profiles than if infiltration is steady, because of enhanced drainage from the matrix caused by increased momentum of soil water (Germann and DiPietro, 1999; Torres and Alexander, 2002). More work is required to elucidate how infiltrating precipitation causes hydrologic response in various soils, and to determine the limits for models of pore pressure propagation.

In addition to phenomena in the vadose zone, Montgomery *et al.* (1997) and Montgomery and Dietrich (2002) showed that water flowing in saprolite and bedrock may exfiltrate into the soil on steep slopes, creating an important mechanism of saturation and positive pore pressure development. In such situations, flow vectors include a component perpendicular to the slope, reducing the stresses resisting downslope movement and increasing the chance of failure (Iverson and Major, 1986).

Although observed rainfall intensities in this study were high, the bursts were short enough that relatively little total water (<10 mm) was involved. The dramatic modelled decreases in factor of safety occur as a result of the assumed initial conditions and soil moisture release curve. When soil is near saturation and conductivity is high, both theory and field observations suggest that small additions of water at the surface result in rapid transmission of less-negative pore water pressure through soil (Smith, 1983; Charbeneau, 1984; Torres *et al.*, 1998). When these pressure disturbances reach a free water surface lower in the soil profile, the result may be a rapid rise of the water table and a quick 'flash' of positive pressure (Reid *et al.*, 1997). Although this phenomenon is characteristic of soil matrix transport, we suggest that rapid transmission of water via bypass flow may have similar effects on pore pressure at the base of the soil profile.

We assumed identical initial soil conditions under canopies and openings, which may not be valid because evaporative loss from sites with canopies is consistently greater and the soils there should generally be drier. However, landslide initiation sites are often in the axis of topographically convergent hillslope hollows (Dietrich *et al.*, 1986) and may, therefore, receive storm-total lateral inflows far in excess of soil porosity during large storms. If so, soil moisture at the site of potential failure may be controlled mainly by rate of outflow once total precipitation has reached some threshold and the landslide initiation site has become part of the saturated variable source area for streamflow generation. This suggests soil moisture status at landslide initiation sites may become independent of vegetation in large storms, and justifies our assumption of identical initial conditions. Conversely, hollows where canopies do not contribute to evaporative loss may develop more extensive and extended saturation during storms. Increased overall chance of slope failure would result, because more area would be near failure for longer periods of time.

More detailed field investigations may elucidate spatial relationships between rainfall and throughfall that affect pore pressures and slope stability. Interactions between geomorphology, hillslope hydrology, and vegetation may be strong enough to require a model more complicated than the one-dimensional approach we employed here. In some cases, preferential flow may allow lateral influx of water to be rapid enough for infiltration on adjacent hillslopes to control short-term variations of pore pressures in the axis of the hollow. It is possible, therefore, that either the vegetation on the site of potential failure or the vegetation in the adjacent contributing area is most relevant. This is important, because vegetation communities can vary systematically within first-order watersheds. For example, frequent disturbance in hollows of the Oregon Coast Range favours early-successional-stage vegetation dominated by deciduous trees and shrubs, whereas coniferous trees usually dominate adjacent hillslopes (Roering *et al.*, 2003).

The forest floor is an important mediator of infiltration. The analyses in this paper have treated the forest floor as part of the soil, not as a separate entity. This assumption may be reasonable in our region, where soils are generally highly porous and high in organic matter, but the forest floor is often more distinct in many parts of the world. In such cases, models may be improved by explicit representation of the forest floor. The forest floor can store amounts of water comparable to tree canopies (Pitman, 1989; Putuhena and Cordery, 1996), and should have specific effects on intensity smoothing beyond those of the canopy. Disruption of the forest floor during disturbance can expose mineral soil to raindrop impact, which can reduce infiltration capacity by surface particle movement, crusting, and sealing (Hillel, 1980). The combination of canopy removal and forest floor disturbance may, therefore, result in infiltration-excess overland flow during high-intensity rainfall and affect a range of hydrological processes, including infiltration and erosion.

We have presented data that describe the effect of the overstory vegetation on throughfall rates, but vegetation in forests is often distributed among several strata. We explicitly eliminated the shrub and herbaceous layers from the data we collected, but the leaf area in such layers can often be substantial, particularly in gaps where shrubs can persist and establish dominance (Spies *et al.*, 1990; Tappeiner *et al.*, 1991). Similarly, shrubs can dominate early successional stages after disturbance, potentially completely occupying the site in only a few years (Oliver, 1981). The effect of these canopies on throughfall intensities deserves consideration.

We ignored the contribution of stemflow to intensity smoothing because we assumed it contributed little mass in our stands, but stemflow is important in other stands. Because it must travel a slower pathway to the soil, stemflow results in more smoothing of rainfall intensity than does throughfall. In stands where stemflow is substantial, soil preferential flowpaths may develop at the location of stemflow inputs (Buol *et al.*, 1989). Removal of canopies on these slopes has the potential to shift infiltration away from concentrated inputs into developed preferential flowpaths toward diffuse infiltration through the soil matrix. The consequences of such changes for slope stability are difficult to predict.

The most striking difference between the two stands in this study was the greater temporal and spatial variability of rainfall–throughfall relationships in the old stand. Gauges in the old stand showed both the greatest and least damping and lagging of throughfall in relation to rainfall. We can easily ascribe the variations to canopy density at each gauge. We infer that throughfall smoothing at a given site will follow stand development, and spatial heterogeneities in canopy coverage that develop with age (Spies and Franklin, 1989) should result in more varied throughfall intensities within the stand.

The storms we investigated varied between sites, so interstand comparisons of the effect on slope stability are weak. We are building stochastic models of the effects of canopy interception on throughfall to investigate the intensity–duration–frequency characteristics of infiltration beneath various stand types and make predictions of the effects these stands might have across a range of storms. Intensity–duration thresholds associated with landsliding (Caine, 1980; Keefer *et al.*, 1987) are one example of a potentially useful tool in making such comparisons.

The importance of canopies in attenuating rainfall intensities in the context of such complicated hydrological conditions is unclear. As a general statement, however, decreasing the magnitude of fluctuations of inputs to the soil should lead to steadier flows and decrease development of high pore pressures.

Conclusions

Although the potential role of vegetation in affecting hydrology and stability of natural slopes has been recognized for some time (e.g. Croft and Adams, 1950; Bishop and Stevens, 1964), most research on the effect of vegetation has remained focused on root reinforcement of soils. The field data we have presented here clearly indicate that a full accounting of vegetative effects on slope stability must include hydrological effects. Forest canopies do modify intensity of precipitation, such that their presence may prevent sliding in some instances. Our results are, thus far, restricted to measured effects of overstory trees only in two example stands; investigation of a broader range of vegetation is needed.

Acknowledgements

This research was supported by grant 00-34158-8978 from the US Department of Agriculture, Cooperative State Research Education and Extension Service, Centers for Wood Utilization Research. We thank Nalini Nadkarni and Bob Van Pelt for sharing their Cedar Flats research plots, and Steve Rentmeester for development work on the throughfall gauges. Two anonymous referees provided comments on an earlier manuscript that substantially improved this work.

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