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ASSESSING SUSCEPTIBILITY AND TIMING OF SHALLOW LANDSLIDE AND DEBRIS FLOW INITIATION IN THE OREGON COAST RANGE, USA

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ABSTRACT

Effective management of debris-flow hazard relies on accurate assessments of debris-flow susceptibility. Mathematical models of rainfall infiltration and slope stability can be applied to predict the temporal and spatial variation of debris-flow susceptibility. These models require high-resolution (<10 m) topographic data, as well as (ideally also high-resolution) data on initial groundwater conditions, physical properties of near-surface earth materials, depth to bedrock, and rainfall. A case study from the Oregon Coast Range, USA illustrates the use of generalized data from a soil survey, limited field measurements, and simple models to parameterize a combined infiltration and slope stability model for predicting debris-flow timing and source-area locations. Although the model over-predicts the extent of debris-flow source areas, results are consistent with mapping which shows channels to be the preferred source areas. Simulation of a November 1996 storm that produced debris flows in the study area indicates that instability probably developed near the end of the period of most intense rainfall; however precise timing of debris flows in the study area is unknown. Model results also indicate that differences in rainfall interception between forested and logged areas may account at least in part for the observed differences in debris-flow susceptibility.

Key words: debris-flow susceptibility, rainfall infiltration, rainfall interception, Oregon Coast Range

INTRODUCTION

Information on where and when debris flows are likely to occur is greatly needed to reduce resultant losses and deaths. Shallow, rainfall-induced landslides that transform into debris flows commonly occur under conditions of transient infiltration into initially unsaturated soils. Application of mathematical models of transient, unsaturated infiltration and slope stability over broad regions to assess timing and potential locations of debris flow requires understanding of unsaturated zone hydrology and soil mechanics as well as information on rainfall, topography and the distribution and properties of hillside soils. In this paper, we survey techniques for generating the spatial and temporal input data for such models and present example calculations for a debris-flow producing storm using our spatially distributed mathematical model for Transient Rainfall Infiltration and Grid-based Slope Stability (TRIGRS) (Baum *et alii*, 2008; 2010). The model combines an analytical solution to assess the pore-pressure response to rainfall infiltration into unsaturated soil with an infinite-slope stability calculation to estimate the timing and locations of slope failures. Pore-pressures and factors of safety are computed on a cell-by-cell basis and can be displayed or manipulated in a grid-based geographic information system (GIS). Input data are high-resolution topographic data and simple descriptions of initial pore-pressure distribution and boundary conditions for a study area in western Oregon, USA. Rainfall information was taken from a nearby recording rain gage. Material strength and hydraulic properties were gleaned from the literature and measured both in the field and laboratory. Results are tested by comparison with an inventory of shallow landslides that mobilized to debris flows.

SETTING AND CLIMATE

Debris flows typically initiate from shallow landslides in the Oregon Coast Range, where hillsides are susceptible to shallow landslides and debris flows due to steep topography, heavy rainfall, and land-use activities associated with timber harvesting (e.g. Brown & Krygier, 1971; Pierson, 1977). Recent work in the region on shallow landslide and debris-flow initiation has focused on the influence of vegetation roots on slope stability (Schmidt *et alii*, 2001; Roering *et alii*, 2003) and the role of pore-water response from rainfall in soil and shallow bedrock (Montgomery *et alii*, 2009; Ebel *et alii*, 2010). The topography of the area is deeply dissected and steep (> 30º). Shallow landslides tend to occur in bedrock hollows (Deitrich & Dunne, 1978), where colluvium is several meters thick (Benda, 1990). Average annual precipitation of about 2000 mm falls mainly as rain during the winter wet season of November to March (Daly *et alii*, 1994). The focus of this study is the North Charlotte Creek basin located about 20 km ESE of Reedsport, OR (Fig. 1) in the Elliott State Forest. The forest is administered by the Oregon Department of Forestry (ODF). The basin is underlain by the middle Eocene Tyee Formation; a tan, medium- to coarse-grained thick-bedded sandstone with interbedded layers of siltstone and shale (NIEM and NIEM, 1990). The bedrock weathers to highly permeable colluvium consisting of abundant sandstone clasts in a matrix of sandy silt or silty sand, covered with a layer of decomposed plant material 0 - 10 cm deep (NCRS, 2010). Depth to bedrock is commonly 25-110 cm deep (NCRS, 2010) except in hollows, where it can be greater than 300 cm deep (Schmidt *et alii*, 2001). Vegetation consists of Douglas Fir, Red Alder, and associated species of coniferous, hardwood, and understory plants (Schmidt *et alii*, 2001).

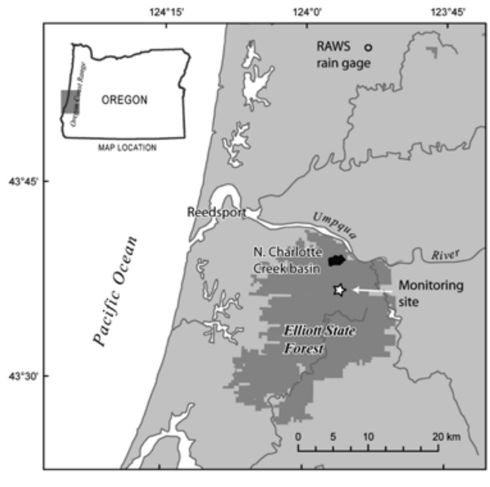

Fig. 1 - Location of study area in Oregon, USA

MODELING APPROACH

We apply the TRIGRS model to predicting the shallow landslide source areas of debris flows in the study area. In this section we briefly describe the theoretical basis of the TRIGRS model (Baum *et alii* 2008; 2010) and review required input data and strategies for defining these inputs. Specific strategies applied to the study area are described in more detail in a later section.

TRANSIENT VERTICAL GROUNDWATER FLOW

The characteristic time scales for lateral and slope normal pore-pressure transmission in initially wet, homogeneous, isotropic hillslopes are A/D_0, and H^2/D_0, respectively, where A is the upslope contributing area above a given location, D_0 is the saturated hydraulic diffusivity, and H is the depth of the failure surface in the slope-normal direction. For areas of the landscape where the ratio $\varepsilon = H/\sqrt{A} << 1$, long- and short-term pressure-head response to rainfall is dominated by vertical flow implying that pore-pressure variation in initially wet materials can be adequately described by simplified, one dimensional forms of the governing equation for infiltration (IVERSON, 2000). The Richards equation describes infiltration and vertical flow through the unsaturated zone (Freeze & Cherry, 1979). When transformed to account for the effects of a sloping surface the equation is

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial Z}\left[K(\psi)\left(\frac{1}{\cos^2 \delta}\frac{\partial \psi}{\partial Z} - 1 \right) \right] \qquad (1)$$

where θ is the volumetric water content, ψ is the pressure head, $K(\psi)$ is the pressure head dependent hydraulic conductivity, and δ is the slope angle. Following Srivastava & Yeh (1991), we linearize equation 1 using Gardner's (1958) exponential hydraulic parameter models, $K(\psi) = K_s \exp(\alpha\psi)$ and $\theta = \theta + (\theta_S - \theta_r)\exp(\alpha\psi)$, where α is the air-entry head or inverse of the height of the capillary rise, K_S is the saturated hydraulic conductivity, θ is the moisture content, and θ_S and θ_r are the saturated and residual moisture contents, respectively. See Baum *et alii* (2008; 2010) for details of the solution scheme and a discussion of limitations of this approach.

STABILITY OF INFINITE SLOPES

With some exceptions (e.g. Reid *et alii*, 2000), models designed for assessing slope stability over large areas using digital topography typically rely on a statically determinate infinite-slope stability analysis (e.g. Montgomery & Dietrich, 1994; Crosta & Frattini, 2003), which assumes slopes are infinitely long and that the failure plane is parallel to the ground surface (Taylor, 1948). Stability of an infinite slope is characterized by the factor of safety ratio, F_S, of resisting basal Coulomb friction to shear stress and is calculated at an arbitrary depth Z by

$$F_S(Z,t) = \frac{\tan\phi'}{\tan\delta} + \frac{c' - \psi(Z,t)\gamma_W \tan\phi'}{\gamma_S Z \sin\delta\cos\delta} \quad (2)$$

where c' and ϕ are soil cohesion and friction angle for effective stress, respectively, γ_W is the unit weight of water, γ_S, the unit weight of soil, and ψ is the groundwater pressure head. For stability above the water table where pressure heads are negative we use an approximation of Bishop's (1959) effective stress and calculate suction stress (Lu & Likos, 2004) as the product of $(\theta - \theta_r)/(\theta_S - \theta_r)$ and $\psi(Z,t)\gamma_W$ (Baum *et alii*., 2008; 2010). Slope failure is predicted where F_S<1, stability holds where $F_S \geq 1$, and F_S=1 indicates a state of limiting equilibrium.

INPUT AND TEST DATA

Grid-based digital topography, soil thickness, materials strength and hydrologic properties, initial groundwater conditions, and time-varying rainfall intensity and duration are required input data for application of the model over broad geographic areas (Baum *et alii*, 2008; Godt *et alii*, 2008a). Landslide maps are needed to test modeling results and should include information on sources, tracks, and deposits, and where and when the landslides occurred.

Digital Elevation Models (DEMs) are regularly spaced arrays of elevation values that are used to calculate local topographic slope, δ. Of the emerging remote-sensing technologies used to generate DEMs, Airborne Laser Swath Mapping (ALSM also commonly referred to as LiDAR - Light Detection and Ranging) has arguably had the most impact on landslide hazard mapping and modeling (e.g. Schulz, 2007). These data are typically of very high spatial resolution (<5 m) with small absolute elevation errors, and can be processed to reveal the "bare earth" surface beneath vegetation (Slatton *et alii*, 2007).

Maps of the soil depth on steep hillsides are required for deterministic shallow landslide models that include the effects of infiltration or soil cohesion (Godt *et alii*, 2008a). Field observations of soil depth in landslide-prone areas of landscapes with well-developed drainage networks indicate that colluvium tends to collect in areas of topographic convergence (hollows) (e.g. Dietrich & Dunne, 1978; Reneau & Dietrich, 1987). On steep slopes (> 20°) colluvium depth tends to decrease exponentially with slope angle (De Rose *et alii*, 1991; De Rose, 1996). Collecting sufficient measurements of soil thickness to it map over broad areas is a practical impossibility, and empirical models have been used to create soil depth maps for landslide susceptibility studies (e.g. De Rose *et alii*, 1996; Salciarini *et alii*, 2006; Godt *et alii*, 2008b). Such models can be constrained using information on soil or colluvial depth from soil surveys, field investigations, and geophysical mapping.

Linearized solutions for infiltration are generally very sensitive to initial water-table depths and moisture conditions of the unsaturated zone and if applied to simulate actual landslide events require some knowledge of initial conditions (Godt *et alii*, 2008a). Where detailed information on groundwater conditions is unavailable, parametric studies assuming a range of initial conditions may indicate the conditions that have caused shallow landslides in the past (Crosta & Frattini, 2003; Salciarini *et alii*, 2006).

Rainfall data are needed to determine the flux at the ground surface for transient modeling of infiltration. Attempts to recreate conditions for past landslide events may benefit from rainfall data with high spatial (kilometer scale) and temporal (hourly) resolution such as can be obtained from weather radars (Crosta & Frattini, 2003). Empirical rainfall intensity-duration thresholds can be used in applications designed to estimate the likelihood of landslide occurrence for a given set of initial and rainfall conditions (Godt *et alii*, 2008b).

The strength and hydrologic properties of hillside materials and an estimate of their spatial distribution can be obtained using field and laboratory tests and soil surveys or geologic mapping (Godt *et alii*, 2008a; 2008b).

Soil strength parameters (angle of internal friction, ϕ' and cohesion c') and saturated hydraulic conductivity, K_S, are typically obtained from standard geotechnical tests (e.g. SAVAGE & BAUM, 2005), however, tests to determine the hydraulic conductivity function $K(\psi)$, or the relation between pressure head and moisture content, $\theta(\psi)$ are needed to estimate the air-entry head, α. Prediction of these relations and parameters from other more easily obtained properties such as particle-size distributions and bulk density is possible (e.g. LEIJ *et alii*, 2002). Plant roots are thought to impart significant strength to hillside soils (e.g. SCHMIDT *et alii*, 2001); however, because the resisting forces imparted by plant roots do not typically act normal to the failure plane, incorporating their effects into one-dimensional infinite-slope stability models (i.e. equation 2) is problematic.

MODEL APPLICTION

We applied the TRIGRS model (BAUM *et alii*, 2008: 2010) to a 2.4 km² study area in the Oregon Coast Range to study model parameterization and performance in an area with a strongly developed dendritic drainage pattern. Heavy rainfall during November 1996 produced numerous debris flows in the central Oregon Coast Range (WILEY, 2000), including those mapped in the study area (Fig. 2). Specific timing of the debris flows during the rainfall is unknown.

The model data were managed in a geographic information system (GIS). Topography was represented by a 3-m resolution DEM, obtained by resampling a 1-m resolution "bare-earth" DEM derived from LIDAR elevation data. We used a cubic convolution resampling algorithm to smooth facets and other artifacts in the 1-m data caused by poor ground-point densities in areas of thick vegetation. Slope is computed using the standard D-8 algorithm (Fig. 2a).

Depth to bedrock was estimated using an exponential function of slope angle (DE ROSE, 1996), with a modification to account for the effect of curvature. Field measurements at a monitoring site several kilometers to the southeast of the study area (Fig. 1) revealed a strong secondary dependence of colluvial thickness on topographic curvature. At locations of equal slope, colluvium was thicker in concave areas along axes of zero order drainage basins and thinner on convex interfluves between the basins. Depth of bedrock ranged from 0.47 m to 3.1 m and slopes ranged from 26° to 41°. We computed depth of bedrock, d_{LZ} (Fig. 2b), in the study area using $(5.0\text{-}1.5\ \mathrm{sgn}(\kappa))\exp(-0.04\delta)$, in which δ is the slope angle and κ is the curvature. Using this model, colluvium is 2.3 m in concave areas and 1.2 m in convex areas on 26° slopes and 1.3 m (concave) or 0.68 m (convex) on 41° slopes. Slopes steeper than 45° were treated

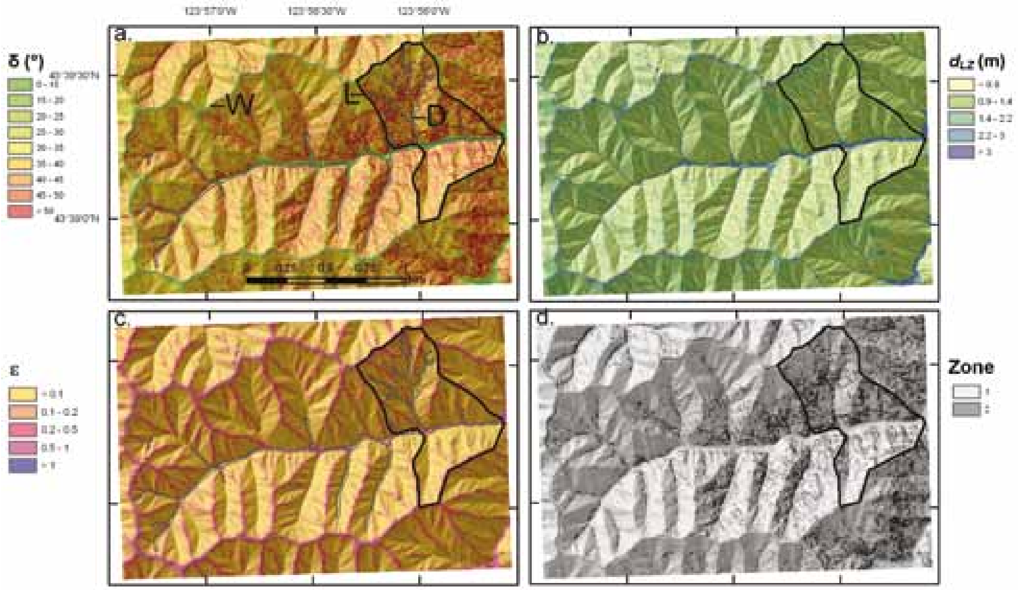

Fig. 2 - Maps showing (a) topographic slope angle, δ, W is watershed boundary, L is logged area, D is debris flow inventory from COE *et alii 2010, (b) depth to bedrock, d_{LZ}, (c) distribution of ratio ε, (d) property zones, 1 is colluvium and 2 is bedrock*

as bedrock outcrops even though thin discontinuous patches of colluvium locally exist there. Assignment of steep slope areas to bedrock outcrop results a higher proportion of rock outcrop in the eastern one third of the basin, consistent with NRCS (2010) soil mapping in the study area (Fig. 2a).

The fall season initial water table depth for the study area was approximated by assuming lateral flow in the colluvium and weathered bedrock. We solved Darcy's law (Freeze & Cherry, 1979) for the height of flow above unweathered bedrock, H_f,

$$H_f = (A / w)(I_{ZLT} / K_S)/(\tan\delta + 0.001), \qquad (3)$$

in which A is the upslope contributing area, w is the cell width (3 m), I_{ZLT} is the average winter infiltration rate, K_S is the saturated hydraulic conductivity and, slope of the ground surface, $\tan\delta$, approximates the hydraulic gradient. The constant, 0.001, prevents division by zero in flat areas. For this computation, both A and $\tan\delta$ were estimated using Tarboton's (1997) D-∞ algorithm. We estimated initial water-table depth using

$$d = H_w + d_{LZ} - H_f\,;\ H_f < H_w + d_{LZ} - d_{\min} \qquad (4a)$$

and

$$d = d_{\min}\,;\ H_f \geq H_w + d_{LZ} - d_{\min} \qquad (4b)$$

in which H_w is the vertical thickness of weathered, permeable bedrock and $d_{\min}$ is the minimum initial water table depth. In this example, H_w = 2.25 and $d_{\min}$ = 0.75 m to ensure that $F_S > 1$ at time t = 0 During fall and winter months water commonly flows in many of the channels; however, we lack data on the channel slope angles where open channel flow begins.

Rainfall during the November 1996 storm was estimated from the nearest available hourly record at similar elevation and distance inland (RAWS rain gauge, Fig. 1). The hourly data were approximated by three time periods of constant rainfall intensity (Fig. 3). The forest canopy intercepts rainfall and modifies the amount and intensity reaching the ground surface (e.g Moore & Wondzell, 2005), and these effects may reduce the potential for landslide generation (Keim & Skaugset, 2003). Mature stands of Douglas fir may reduce the amount of rainfall reaching the ground surface by as much as 40 percent (e.g. Link *et alii*, 2004). Keim *et alii*, (2004) present a stochastic model that describes the reduction in rainfall intensity for a given storm duration and return period. In general, rainfall from high-intensity storms is affected less than that from low-intensity storms. For the case presented here, rainfall intensities were applied at

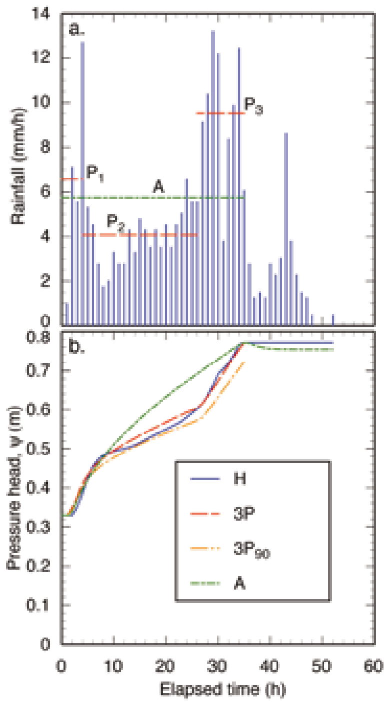

Fig. 3 - Rainfall and computed pressure head response (a) Rainfall at the Goodwin Peak Remote Automated Weather Station (RAWS) during the November 1996 storm, A is average rainfall, P_1 P_2, and P_3, are rainfall during first, second, and third, time periods, respectively. (b) Computed pressure head for a typical cell in the debris-flow source areas H is hourly rainfall, A is Average rainfall, 3P is hourly rainfall averaged over three time periods, $3P_{90}$ is 90% of 3P, used in forested areas. (α=40°, d=0.75 m, d_{LZ}=1.3 m)

100% in logged areas (Fig. 2) and 90% in forested areas to account for interception consistent with observed storm intensity (Keim *et alii*, 2004).

We identified areas in the digital landscape (Fig 2c) where vertical transmission of pore-pressure as implemented in the TRIGRS model is dominant over lateral transmission (Iverson, 2000) by comparing the slope normal depth, H_D, and the square root of the upslope contributing area, A. Where the ratio $\varepsilon = H_D / \sqrt{A} << 1$ long and short-term pressure-head responses to rainfall can be adequately described by 1-D linear and quasi-linear approximations to the Richards equation (Iverson, 2000). We computed the upslope contributing area for each grid cell using Tarboton's (1997) D-∞ algorithm and calculated slope-normal depths using the vertical thickness (Fig. 2b) and topographic slope (Fig. 2a). Fig. 2c compares the distribution of mapped shallow land-

Map Unit (Zone)	I_{ZLT} (m/s)	K_S (m/s)	D_0 (m²/s)	θ_r	θ_s	α (m^{-1})	c' (kPa)	ϕ'
Colluvium (1)	1.0×10^{-9}	1.4×10^{-5}	1.4×10^{-4}	0.082	0.47	3.0	6.0	40°
Weathered bedrock (2)	1.0×10^{-9}	1.0×10^{-5}	6.0×10^{-5}	0.05	0.25	0.5	45.0	45°

Tab. 1 - Material Strength and Hydraulic Properties for the Two Map Units in Fig. 2d

slides (Coe *et alii*, 2010) and ε in the study area. Values of ε are typically much less than unity along the steep slopes and in the channel network where the mapped shallow landslides (debris-flow source areas) lie. At most points within the mapped landslides, $\varepsilon = \leq$ as suggested by Iverson (2000) (Fig. 2c).

We used two property zones (table 1, Fig. 2d), bedrock outcrop (as defined previously), and colluvium. Strength parameters of the colluvium were based on published values for previous work in nearby areas of similar geology (Ebel *et alii*, 2010). Although the contribution of root strength and pullout resistance could be a further basis for distinguishing differences between the logged and forested areas, implementation using (2) is problematic, as noted previously. Hydraulic properties of the colluvium were estimated from a published soil survey (NRCS, 2010) and unpublished monitoring data. Continuous unsaturated conditions observed throughout the fall and winter rainy season are consistent with $\alpha \approx 3.0$ m^{-1}, rather than $\alpha \approx 0.5$ m^{-1} as indicated by analysis of the soil survey data. Hydraulic diffusivity, D_0, was estimated as described by Baum *et alii* (2010) so that $2K_S \leq D_0 \leq 10K_S$.

Properties of weathered bedrock were estimated by adjusting colluvial properties toward lower porosity, permeability and diffusivity and higher strength (Beaulieu & Hughes, 1975; Freeze & Cherry, 1979).

RESULTS

Simulated rainfall infiltration resulted in gradual pore pressure rise and reduction in factor-of-safety. Simulating infiltration using three periods of constant rainfall rather than hourly data (35 periods) preserves the major features and timing of pore pressure rise (Fig. 3b) while reducing computational effort by about 90%. Infiltrating water percolated to the shallow water table, causing it to rise gradually. The water table rose above the top of weathered bedrock only in channel areas and colluvium above the water table remained unsaturated (Fig. 4). Expansion of saturated areas up the channels into hollows is consistent with observations in similar topographic settings elsewhere (Wilson & Dietrich, 1987). Factor-of-safety gradually decreased in the steep channels, where debris flows occurred in 1996, as well as in other similar channels. Due to higher infiltration rates, pore pressures rose more rapidly in the logged areas than in the forested areas (Fig 3b, 4). Consequently the proportion of steep channel grid cells where F_S<1 by the end of the storm was greater in the logged area than in the forested areas.

DISCUSSION

Several points related to data preparation, model results, and interpretation of results merit further explanation. In the case study presented, the TRIGRS model produced results that are consistent with the observed distribution of debris-flow source areas despite the use of generalized and in some cases estimated data. Although we are currently in the process of acquiring more detailed data on physical properties and hydrologic response of debris-flow source areas in the Coast Range, these results indicate that the TRIGRS model can be applied successfully even where limited data are available. The model over-predicts potential landslides as a result of data uncertainty, particularly fine-scale subsurface variations in the depth of bedrock, strength and hydrologic properties, and initial groundwater and soil moisture conditions. Consequently results must be interpreted in a probabilistic sense.

GIS methods used in preparing some of the input grids, notably the initial water table depth resulted in artifacts that skew some of the model results. Both the D-8 and the D-∞ methods of computing upslope contributing area result in many cells adjacent to the main channels having very small contributing areas. Consequently (3), (4a) and (4b) computed very low water tables at many grid cells near the valley bottom (Fig. 4). A finite-difference or finite-element model for steady saturated-unsaturated groundwater flow would compute a more realistic initial water table, probably at considerable computational expense.

Several factors may account for the much higher density of debris flows in the logged area compared to forested areas (Fig. 5). These include (1) greater difficulty detecting debris flows beneath forest cover (ODF, 1999), (2) the greater proportion of steep

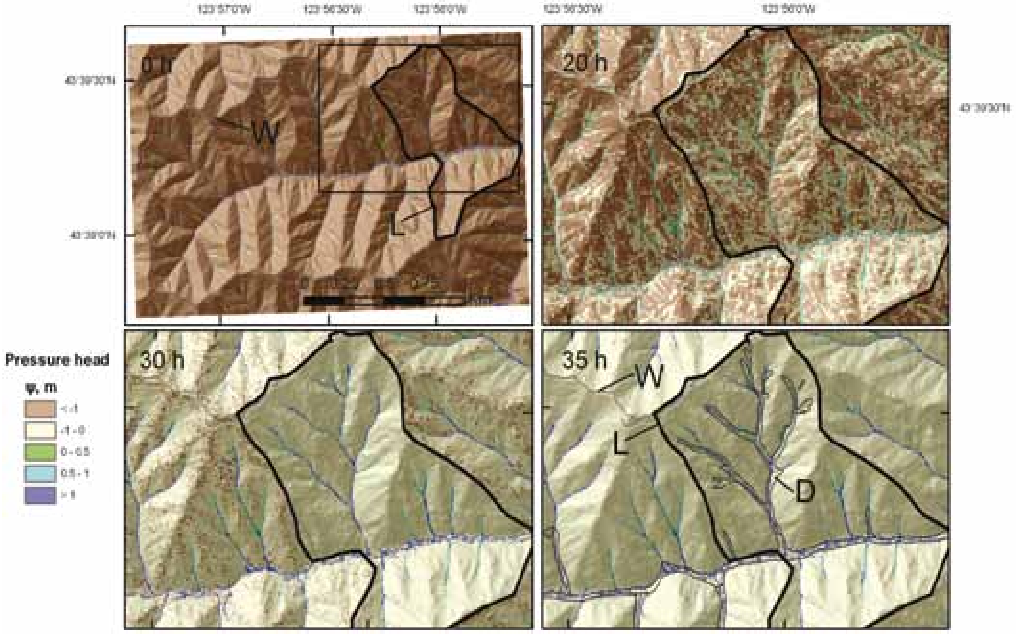

Fig. 4 - Maps showing evolution of pressure head, ψ at base of colluvium at different stages of the three rainfall periods in Fig. 3 and boundary and initial conditions depicted in Fig. 2. Rainfall after 20 h=85 mm, after 30 h =161 mm and after 35 h = 201 mm. Rectangle in upper left panel (0 h) indicates approximate area shown in other panels, L is logged area, D is boundaries of debris flows

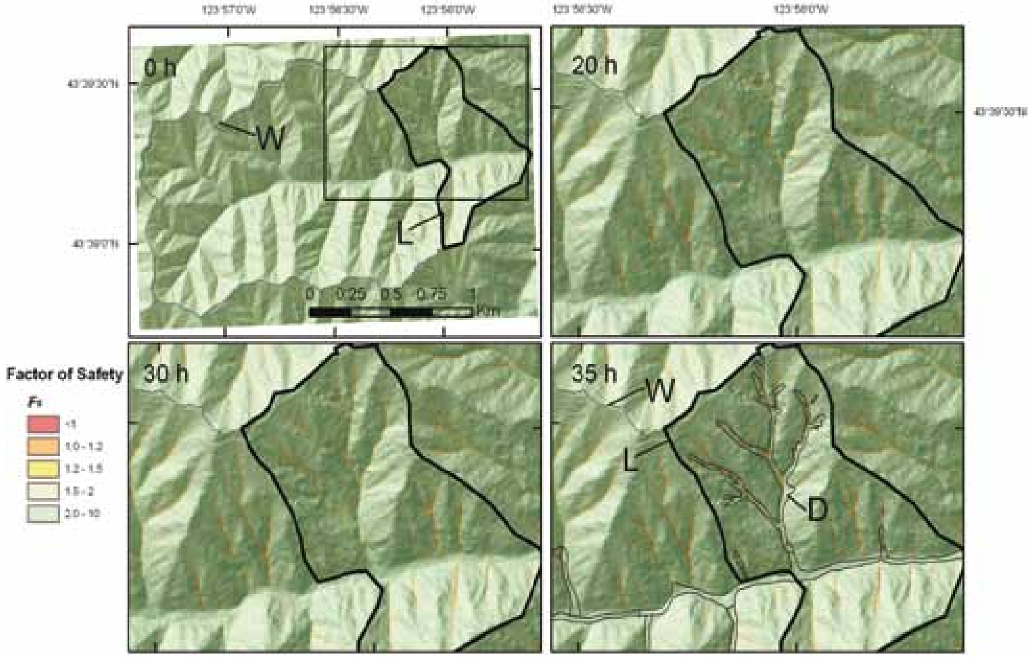

Fig. 5 - Maps showing evolution of factor of safety. Rectangle in upper left panel (0 h) indicates approximate area shown in other panels, L is logged area, D is boundaries of debris flows

slopes in the logged area (Fig. 2), (3) differences in evapotranspiration, which might result in drier initial conditions in the wooded areas, (4) loss of root strength and pullout resistance in logged areas, and (5) interception of rainfall by the forest as considered in our model. We are confident that all debris flows in the study area that resulted from the November 1996 rainfall are shown on the map (Fig. 4). Given sufficient data the TRIGRS model could account for the other four factors. Our model accounts for two of these (distribution of steep slopes, and interception) and our results indicate that even the modest effect of interception explains at least some of the difference in susceptibility between logged and wooded areas.

CONCLUSIONS

The TRIGRS model was used to simulate rainfall infiltration and occurrence of shallow landslides from a November 1996 storm that resulted in widespread landslides and debris flows in the Oregon Coast Range.

- Modeling results are consistent with observations that steep channels in this setting are more susceptible to shallow landslides and debris flows than either hillsides or basin rims as observed in other settings.
- Parameter estimation for the TRIGRS model can be based on generalized data and simple models of soil depth distribution and initial water table, where site-specific data are lacking. Availability of more specific data would reduce model uncertainty.
- Storm rainfall interception can account for at least part of the observed landslide susceptibility difference between logged and forested areas.

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REFERENCES

Baum R.L., Savage W.Z. & Godt, J.W. (2008) - TRIGRS - *A Fortran program for transient rainfall infiltration and grid-based regional slope-stability analysis, version 2.0.* U.S. Geological Survey Open-File Report 2008-1159.

Baum R.L., Godt J.W. & Savage, W.Z. (2010) - *Estimating the timing and location of shallow rainfall-induced landslides using a model for transient, unsaturated infiltration*. Journal of Geophysical Research, doi:10.1029/2009JF001321.

Beaulieu J.D. & Hughes P.W. (1975) - *Environmental geology of western Coos and Douglas Counties, Oregon*. Oregon Department of Geology and Mineral Industries Bulletin 87.

Benda L. (1990) - *The influence of debris flows on channels and valley floors in the Oregon Coast Range, U.S.A*. Earth Surface Processes and Landforms, **15**: 457-466.

Bishop A.W. (1959) - *The principle of effective stress*. Teknisk Ukeblad, **39**: 859-863.

Brown G.W. & Krygier J.T. (1971) - *Clear-cut logging and sediment production in the Oregon Coast Range.* Water Resources Research, **7**(5): 1189-1198.

Coe J.A., Michael J.A. & Burgos M.M. (2010) - *Map showing debris flows caused by precipitation in 1996 in parts of the Reedsport and Deer Head Point quadrangles, southern Coast Range, Oregon.* U.S. Geological Survey Open File Report.

Crosta G.B. & Frattini R. (2003) - *Distributed modelling of shallow landslides triggered by intense rainfall*. Natural Hazards and Earth System Sciences, **3**: 81-93.

Daly C., Neilson R.P. & Phillips D.L. (1994) - *A statistical-topographic model for mapping climatological precipitation over mountainous terrain.* Journal of Applied Meteorology, **33**: 140-158.

De Rose R.C., Trustrum N.A., & Blaschke P.M. (1991) - *Geomorphic change implied by regolith-slope relationships on steepland hillslopes, Taranaki, New Zealand*. Catena **18** : 489-514.

De Rose R.C. (1996) - *Relationships between slope morphology, regolith depth, and the incidence of shallow landslides in eastern Taranaki hill country*. Zeitschrift für Geomorphologie Supplementband **105** : 49-60.

Dietrich W.E. & Dunne T. (1978) - *Sediment budget for a small catchment in mountainous terrain.* Zeitschrift für Geomorphologie Supplementband, **29**: 191-206.

Ebel B.A., Loague K., & Borja R.I. (2010) - *The impacts of hysteresis on variably saturated hydrologic response and slope failure*. Environmental Earth Sciences, doi: 10.1007/s12665-101-0448-z.

Freeze R.A. & Cherry J.A. (1979) - *Groundwater.* Prentice-Hall Inc., Englewood Cliffs, New Jersey.

Gardner W.R. (1958) - *Some steady-state solutions of the unsaturated moisture flow equation with application to evaporation from a water table*. Soil Science, **85**: 228-232.

Godt J.W., Baum R.L., Savage W.Z., Salciarini D., Schulz W.H. & Harp,E.L. (2008a) - T*ransient deterministic shallow landslide modeling: Requirements for susceptibility and hazard assessment in a GIS framework.* Engineering Geology, **102**: 214-226.

Godt J.W., Schulz W.H., Baum R.L. & Savage W.Z. (2008b) -*Modeling rainfall conditions for shallow landsliding in Seattle, Washington*. In: Baum R.L, Godt J.W. & Highland L.M. (eds.) *Landslides and Engineering Geology of the Seattle, Washington, area*. Geological Society of America Reviews in Engineering Geology, **20**: 137-152.

Iverson R.M. (2000) - *Landslide triggering by rain infiltration.* Water Resources Research, **36**(7) : 1897-1910.

Keim R.F. & Skaugset A.E. (2003) - *Modelling effects of forest canopies on slope stability*. Hydrological Processes, **17**: 1457-1467.

Keim R.F., Skaugset A.E., Link T.E. & Iroume A. (2004) - *A stochastic model of throughfall for extreme events*. Hydrology and Earth System Sciences, **8**(1): 23-34.

Leij P.J., Schaap M.G. & Arya L.M. (2002) - *Indirect methods.* In: Dane, J.H. & Topp, G.C. (eds.) *Methods of Soil Analysis* - Part 4 Physical Methods: 1009-1045, Soil Science Society of America, Madison, WI.

Link T.E., Unsworth M. & Marks D. (2004) - *The dynamics of rainfall interception by a seasonal temperate rainforest.* Agricultural and Forest Meteorology, **124**: 171-191.

Lu N. & Likos W.J. (2004) - *Unsaturated soil mechanics*. Wiley, New York.

Montgomery D.R., Schmidt K.M., Dietrich W.E. & McKean J. (2009) - *Instrumental record of debris flow initiation during natural rainfall: Implications for modeling slope stability*, Journal of Geophysical Research, **114**, F01031.

Montgomery D.R., & Dietrich W.E. (1994) - *A physically-based model for the topographic control on shallow landsliding.* Water Resources Research, **30**: 1153-1171.

Moore R.D. & Wondzell S.M. (2005) - *Physical hydrology and the effects of forest harvesting in the Pacific Northwest: A review.* Journal of the American Water Resources Association, **41**(4): 763-784.

NRCS (2010) - *Soil Survey Geographic (SSURGO) database for Douglas County Area, Oregon. U.S.* Department of Agriculture, Natural Resources Conservation Service, on-line at http://soildatamart.nrcs.usda.gov/Metadata.aspx?Survey=OR649&UseState=OR (accessed 30 Aug. 2010).

ODF (1999) - *Storm Impacts and Landslides of 1996: Final Report*. Oregon Department of Forestry, Forest Practices Technical Report 4.

Niem A.R. & Niem W.A. (1990) - *Geology and oil, gas, and coal resources, southern Tyee Basin, southern Coast Range, Oregon.* Department of Geology and Mineral Industries, Open-File Report O-89-3.

Pierson T.C. (1977) - *Factors controlling debris-flow initiation on forested hillslopes in the Oregon Coast Ranges.* Ph.D. thesis, University of Washington, Seattle, Washington.

Reid M.E., Christian S.B. & Brie D.L. (2000) - *Gravitational stability of three dimensional stratovolanco edifices.* Journal of Geophysical Research, **205**: 6043-6056.

Reneau S.L. & Dietrich W.E. (1987) - *The importance of hollows in debris flow studies: Examples from Marin County, California*. In: Costa J.E. & Wieczorek G.F. (eds.) - *Debris flows/avalanches: Process, recognition, and mitigation.* Geological Society of America Reviews in Engineering Geology, **7**: 165-658.

Roering, J.J., Schmidt, K.M., Stock., J.D., Dietrich, W.E., & Montgomery, D.R. (2003) - *Shallow landsliding, root reinforcement, and the spatial distribution of trees in the Oregon Coast Range,* Canadian Geotechnical Journal, **40**: 237-253.

Salciarini D., Godt J.W., Savage W.Z., Conversini R., Baum R.L. & Michael J.A. (2006) - *Modeling regional initiation of rainfall-induced shallow landslides in the eastern Umbria Region of central Italy*. Landslides, **3**: 181-194.

Savage W.Z. & Baum R.L. (2005) - *Instability of steep slopes*. In: Jakob M. & Hungr O. (eds.) *Debris-Flow Hazards and Related Phenomena*: 53-77, Springer, Berlin.

Schmidt K.M., Roering J.J., Stock J.D., Dietrich W.E., Montgomery D.R., & Schaub T. (2001) - *The variability of root cohesion as an influence on shallow landslide susceptibility in the Oregon Coast Range.* Canadian Geotechnical Journal, **38**: 995-1024.

Schulz W.H. (2007) - *Landslide susceptibility revealed by LiDAR imagery and historical records, Seattle, Washington.* Engineering Geology, **89**: 67-87.

Slatton K.C., Carter W.E., Shrestha R.L. & Dietrich W. (2007) - *Airborne Laser Swath Mappping: Achieving the resolution and accuracy required for geosurficial research*. Geophysical Research Letters, **34**: L23S10.

Srivastava R. & Yeh T.-C.J. (1991) - *Analytical solutions for one-dimensional, transient infiltration toward the water table in homogeneous and layered soils.* Water Resources Research, **27**: 753-762.

Taylor D.W. (1948) - *Fundamentals of soil mechanics.* Wiley, New York.

Tarboton D.G. (1997) - *A new method for the determination of flow directions and contributing areas in grid digital elevation models.* Water Resources Research, **33**(2): 309-319, doi: 10.1029/96WR03137.

Wiley T.J. (2000) - *Relationship between rainfall and debris flows in western Oregon.* Oregon Geology **62**(2): 27-43.

Wilson C.J. & Dietrich W.E. (1987) - *The contribution of bedrock groundwater flow to storm runoff and high pore pressure development in hollows.* Erosion and Sedimentation in the Pacific Rim, International Association of Hydrological Sciences Publication, **165**: 49-59.