Forest clearing and regional landsliding

David R. Montgomery
Kevin M. Schmidt*
Harvey M. Greenberg
Department of Geological Sciences, University of Washington, Seattle, Washington 98195, USA
William E. Dietrich — Department of Geology and Geophysics, University of California, Berkeley, California 94720, USA

ABSTRACT

The influence of forest clearing on landsliding is central to long-standing concern over the effects of timber harvesting on slope stability. Here we document a strong topographic control on shallow landsliding by combining unique ground-based landslide surveys in an intensively monitored study area with digital terrain modeling using high-resolution laser altimetry and a coarser resolution regional study of 3224 landslides. As predicted by our digital terrain–based model, landslides occur disproportionately in steep, convergent topography. In terrain predicted to be at low risk of slope failure, a random model performs equally well to our mechanism-based model. Our monitoring shows that storms with 24 hr rainfall recurrence intervals of less than 4 yr triggered landslides in the decade after forest clearing and that conventional monitoring programs can substantially underestimate the effects of forest clearing. Our regional analysis further substantiates that forest clearing dramatically accelerates shallow landsliding in steep terrain typical of the Pacific Northwest.

Keywords: landsliding, debris flows, forestry, erosion.

INTRODUCTION

For more than a century associations between forest cutting and landsliding have been interpreted as evidence that forest clearing accelerates erosion in mountainous terrain (Lyell, 1853). The widespread use of aerial photography in forest management has led to many studies that concluded that forest clearing dramatically accelerates rates of landsliding over rates in undisturbed forest (Sidle et al., 1985). However, quantitative estimates of the effect of timber harvest on landslide rates have been challenged because the difficulty in detecting small landslides under a forest canopy could impart a bias to aerial photograph inventories (Pyles and Froehlich, 1987; Robison et al., 1999). In addition, the timing of storm events relative to forest practices is variable and most studies cover only a single watershed, a situation that has contributed to variations in the reported influence of timber harvest on rates of landsliding. Here we combine comprehensive, ground-based mapping of postharvest landslides in a small watershed with a regional analysis to evaluate the relation of contemporary and longer term rates of landsliding.

METTMAN RIDGE LANDSLIDING

Repeat mapping of a 0.43 km^2 area along Mettman Ridge near Coos Bay, Oregon, provides a unique record of postlogging slope instability in which all landslides were documented over a 10 yr period (Fig. 1). The study area burned in the late nineteenth century, was clear-cut logged in 1987, and replanted in 1988. Mettman Ridge is underlain by Eocene sandstone and the colluvial soil is close to cohesionless, the net soil cohesion being dominated by root strength (Schroeder and Alto, 1983; Schmidt, 1999).

In December 1989, channels, landslide scars, and colluvial deposits in topographic hollows were mapped by walking the entire study area (Montgomery and Dietrich, 1992, 1994). Most subsequent landslides were mapped in the field, although several of the 1996 landslides were mapped from low-altitude aerial photographs. We identified 16 landslide scars during initial mapping in 1989. Although we do not know the absolute ages of these scars, their fresh appearance and the lack of stumps or logging slash within the scarps indicated that they postdate logging; therefore, the landslides occurred between 1987 and 1989. Five additional landslides occurred in 1990, three in 1991, five in 1992, one in 1993, and five more during the storm of record in November 1996. All of the landslides extended to near the soil-bedrock interface, and 30 of the 35 landslides occurred in topographic hollows that define the axes of unchanneled valleys. Ten of the landslides were associated with drainage from a ridgetop road. Due to rapid regrowth of understory vegetation none of the pre-1996 scarps are discernable on low-altitude (1:2400 scale) aerial photographs flown in January 1997.

Landslides in shallow soil typically occur during intense or long-duration storms (Caine, 1980), and two intensity-duration thresholds are apparent at Mettman Ridge for slide-producing storms from 1987 to 1998 (Fig. 2). Caine's (1980) global threshold for widespread landsliding works

*Present address: U.S. Geological Survey, MS 989, 345 Middlefield Road, Menlo Park, California 94025, USA.

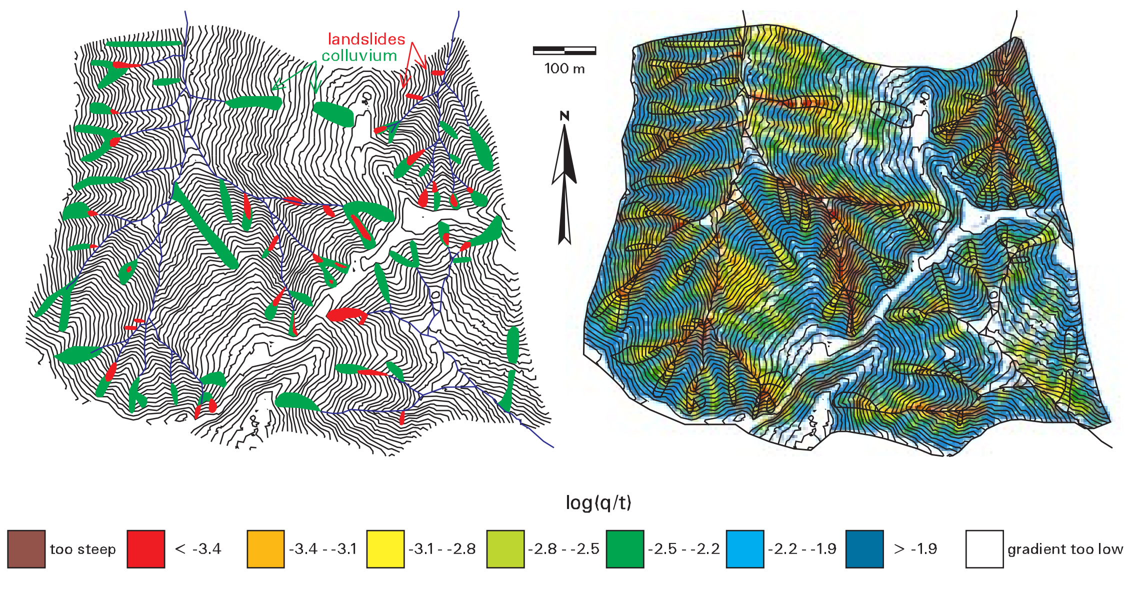

Figure 1. Mapped channel network (blue lines), colluvial fills in topographic hollows, and landslides that occurred from 1987 to 1997 on Mettman Ridge, Oregon. Base map is derived from laser altimetry survey in which average data spacing was 2.5 m, with estimated 0.3 m vertical uncertainty; contour interval is 5 m. Predicted $(q/T)_c$ values defined by equation 2 for C_b = 0, and C_l = 6 kPa (see text).

well for defining an upper bound to storms that did not produce landslides in our mapping area, and a lower rainfall intensity-duration threshold fits the lower bound of the storms that triggered landslides. Piezometric observations at an intensively instrumented hollow on Mettman Ridge indicate that pore-water pressures, and hence landslide initiation, are sensitive to hourly rainfall bursts (Montgomery et al., 1997; Torres et al., 1998) but we lack a historical rainfall record with temporal resolution finer than 24 hr.

The maximum 24 hr rainfall in years during which landslides were documented ranged from a recurrence interval of <2 yr to the largest on record (Fig. 3). The 1996 storm was the 24 hr storm of record, with 16 cm day^{-1} at the North Bend gauge (and >14 cm day^{-1} at a Mettman Ridge rain gauge that was destroyed by a 1996 debris flow), but other landslide-producing storms had 24 hr rainfall recurrence intervals of 3–4 yr and 1–2 yr based on the 65 yr record from the North Bend gauge. However, the maximum hourly rainfall at Mettman Ridge (13 mm) was comparable to the maximum hourly intensitites for the smaller landslide-producing storms of 1990 and 1992 (10 and 16 mm hr^{-1}). The relatively modest number of landslides in the storm of record may reflect hourly rainfall intensity, prior failure of the most susceptible areas, and regeneration of root strength 9 yr after cutting. In contrast, more than half of the landslides observed since direct monitoring began occurred from 3 to 5 yr after cutting. The low intensity-duration threshold for Mettman Ridge after timber harvest and the observation that nearly half of the landslides occurred during rainstorms with a <4 yr 24 hr recurrence interval show that common storms can trigger landslides in the decade following cutting in steep, landslide-prone terrain.

The average recurrence interval for landsliding from hollows is the product of the number of hollows and the period of observation divided by the number of hollows in which landslides occurred during the monitoring period. We mapped colluvial deposits in topographic hollows, most of which were unchanneled valleys upslope of channel heads. There are between 48 and 56 hollows in the mapped area (some unchanneled valleys bifurcate upslope of the channel head and some do not support a channel head). The total of 35 mapped landslides in hollows in the 10 yr since clearcutting yields a rate of 8 landslides km^{-2} yr^{-1} and an average recurrence interval for landsliding from a hollow of <20 yr. Considering only the 25 landslides that could not have been related to road drainage yields a rate of 5.8 landslides km^{-2} yr^{-1} and a recurrence interval of about 20 yr.

The erosion rate from landsliding on Mettman Ridge can be calculated from the 8700 m^2 surface area involved in the landslides shown in Figure 1. The landslides along Mettman Ridge are about 1 m thick and using a dry bulk density for the colluvial soil of 1.2 t m^{-3} to convert the volumetric soil loss to a sediment yield during the 10 yr monitoring period over the entire 0.43 km^2 area yields 2400 t km^{-2} yr^{-1}, which corresponds to a net lowering rate of 2.0 mm yr^{-1}. This contemporary lowering rate is more than an order of magnitude greater than the 0.1 mm yr^{-1} rate of soil production at Mettman Ridge calculated from cosmogenic radionuclide analysis of weathered bedrock and river sand (Heimsath, 1999) and the lowering rate of 0.07 mm yr^{-1} calculated from rates of hollow infilling (Reneau and Dietrich, 1991).

Background recurrence intervals for landsliding from hollows can be independently estimated from ^{14}C dating of charcoal from basal colluvium. Although landslide rates likely varied through the Holocene, basal colluvium dates constrain the time since the most recent landsliding. At Mettman Ridge, charcoal fragments collected from the base of a colluvial deposit in one of the hollows that failed during the monitoring period indicate ^{14}C ages of 660 ± 60 B.P. at the channel head and 4070 ± 90 B.P. in a soil pit upslope along the hollow axis. The younger date at the channel head could reflect either progressive infilling of a single original landslide scar from upslope, or a more recent failure of the channel head. Previously reported basal colluvium dates from hollows in the Oregon Coast Range span 1600 to >40 000 B.P. (Benda and Dunne, 1987; Reneau and Dietrich, 1990). Discounting the >40 000 B.P. date and adding our two dates to those reported previously yields a sample of 13 dates, with a mean of 5600 yr and a standard deviation of 2800 yr. Using the 3–8 k.y. range of recurrence intervals defined by one standard deviation from the mean ^{14}C date together with a typical range for hollow density of 80–100 hollows km^{-2} predicts a long-term hollow failure rate of 0.01–0.03 landslides km^{-2} yr^{-1} for the Oregon Coast Range, two orders of magnitude less than the observed contemporary rate at Mettman Ridge.

SLOPE STABILITY MODEL

In the Oregon Coast Range lateral root strength along the perimeter of a potential landslide provides the primary reinforcement because few roots penetrate into the underlying bedrock. The widely used infinite-slope stability model can be modified to account for lateral root strength (Riestenberg and Sovonick-Dunford, 1983; Burroughs, 1984; Reneau and Dietrich, 1987) and rearranged to solve for the critical proportion of the soil that must be saturated (m_c) to trigger slope failure:

$$m_c = \frac{C_l P z + C_b A}{A \rho_w g z \cos^2\theta \tan\phi} + \frac{\rho_s}{\rho_w}\left(1 - \frac{\tan\theta}{\tan\phi}\right), \qquad (1)$$

where θ is the local slope angle, C_l and C_b are the apparent lateral and basal cohesion due to both soil and root properties, ρ_s and ρ_w are the density of soil and water, g is gravitational acceleration, z is the soil thickness, ϕ is the friction angle of the soil, and A and P are the basal area and perimeter of the landslide. An m_c value greater than 1 indicates that pore-water pressure in excess of hydrostatic is required for landsliding. Assuming negligible basal root reinforcement for a small (10 × 5 × 1 m) landslide in the generally cohesionless soils of the Oregon Coast Range, m_c greatly exceeds 1 for most slopes for C_l > 12 kPa (Fig. 4). In the Oregon Coast Range, the apparent cohesion attributable to mature conifers is generally >10 kPa, whereas that attributable to cut stumps and understory is 2–4 kPa (Burroughs and Thomas, 1977; Schmidt, 1999). Hence, in a mature forest, landsliding should initiate only where local hydrologic conditions lead to pore pressures well above hydrostatic (Montgomery et al., 1997) or in local areas of low root strength. Notably, C_l < 12 kPa allows for slope instability at less than fully saturated conditions, which accounts for landsliding during low recurrence interval storms after timber harvest.

We extend this approach to predict locations susceptible to landsliding after timber harvest using a spatially distributed model that quantifies the topographic control on shallow landsliding by assuming infiltration of rainfall and topographically driven flow through soil over impermeable bedrock (Montgomery and Dietrich, 1994). Specifically, we assume that $m = qa/(Tb\sin\theta)$, where q is the effective precipitation (precipitation less evapotranspiration and deep drainage), a is the drainage area to a contour length b (here assumed to be the grid cell size), and T is the transmissivity (the vertical integration of the saturated hydraulic conductivity). This is a solution for steady state precipitation and runoff for shallow subsurface flow, which is assumed to follow the surface topography. The relative potential for shallow landsliding is given by the q/T required to generate instability:

$$(q/T)_c = \frac{\sin\theta}{(a/b)}\left[\frac{C_l P z + C_b A}{A \rho_w g z \cos^2\theta \tan\phi} + \frac{\rho_s}{\rho_w}\left(1 - \frac{\tan\theta}{\tan\phi}\right)\right]. \qquad (2)$$

Values for $(q/T)_c$ are calculated for locations with slopes between those that will not fail even if saturated $\{\tan\theta < [(C_l P z + C_b A)/A \rho_s g z \cos^2\theta] + [1 - (\rho_w/\rho_s)] \tan\phi\}$ and those predicted to be unstable even when dry, and therefore considered unconditionally unstable areas $\{\tan\theta \ge \tan\phi + [(C_l P z + C_b A)/A \rho_s g z \cos^2\theta]\}$. Smaller values of $(q/T)_c$ indicate less rainfall needed for instability, and hence a greater potential for instability.

We used a 2 m grid digital elevation model (DEM), ρ_s/ρ_w = 1.6, ϕ = 33°, and a landslide length of 10 m and width of 5 m in our Mettman Ridge simulations. Instead of predicting soil depth (Dietrich et al., 1995), we used a uniform soil depth of 1 m because detailed mapping revealed sub-

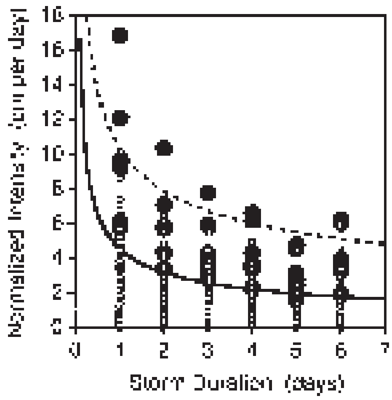

Figure 2. Rainfall intensity versus storm duration for January 1987 to March 1998 based on North Bend, Oregon, rain gauge. Large dots represent storms in periods during which debris flows were triggered at Mettman Ridge; smaller open circles represent all other storms. Upper dashed line is Caine's (1980) threshold for debris-flow initiation based on global data compilation; $I = 14.8D^{-0.39}$, where I is intensity (mm/hr) and D is duration (hr); lower solid line is relation with similar form fit to lower bound of slide-producing storms at Mettman Ridge; $I = 9.9D^{-0.52}$.

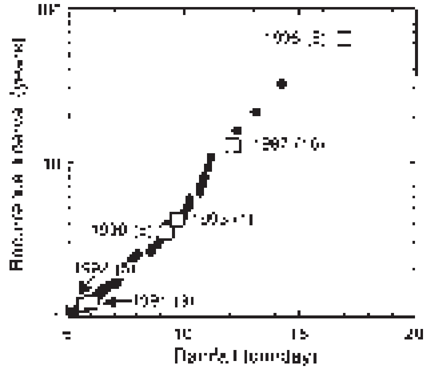

Figure 3. Recurrence intervals for 24 hr rainfall based on North Bend, Oregon, rain gauge for 1931 through 1996. Squares designate years (first number) during which landslides were mapped in study area and second number (in parentheses) is number of landslides during that year.

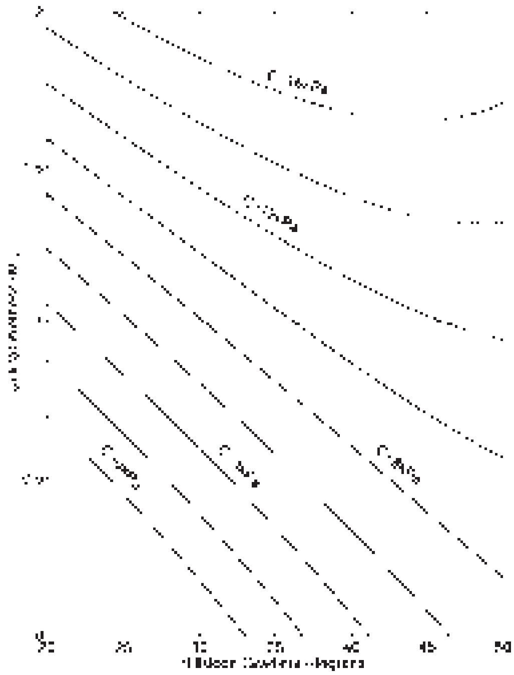

Figure 4. Critical wetness, m_c, defined by equation 1 versus ground slope for C_l of 0–16 kPa, C_b = 0, ρ_s/ρ_w = 1.6, ϕ = 33°, z = 1 (see text), and landslide length of 10 m and width of 5 m.

stantial local variability to soil thickness—likely due to tree throw and previous removal of colluvium by landsliding (Heimsath, 1999; Schmidt, 1999). The ϕ value of 33° is based on field measurements using a shear vane on in situ material at Mettman Ridge and previous published values for similar Oregon Coast Range soils (Yee and Harr, 1977; Schroeder and Alto, 1983). Although the parameters in equation 2 vary across a landscape and can be considered as having a distribution of values (Hammond et al., 1992), the spatial pattern of such variability is unpredictable. Consequently, we assign mean values and interpret variability in failure occurrence within a slope stability class as being due, in part, to the unknown spatial variance in site properties.

Assuming negligible basal root reinforcement, even high postlogging C_l of 4–6 kPa predicts large areas of Mettman Ridge to be potentially unstable, with the greatest hazard in steep, convergent topography (Fig. 1). As C_l increases, zones of potential instability become restricted to topographic hollows and disappear at C_l > 12 kPa. Mapped landslides at Mettman Ridge overlapped locations predicted to be unconditionally unstable at C_l < 4 kPa.

We can also compare (q/T) values for storms that triggered landslides at Mettman Ridge to the range of steady-state $(q/T)_c$ values predicted by equation 2. Falling head conductivity tests and soil depth profiles indicate that T = 65 m^2 day^{-1} at Mettman Ridge (Montgomery and Dietrich, 1994; Montgomery et al., 1997). Using the lower bound of 5 cm day^{-1} for the 24 hr rainfall observed to cause landslides at Mettman Ridge (Fig. 2) yields $\log(q/T) = -3.1$, a value that corresponds to the minimum $\log(q/T)_c$ values within the footprint of Mettman Ridge landslides.

REGIONAL LANDSLIDING

We evaluated contemporary rates of shallow landsliding based on a regional set of 3224 landslides mapped during watershed analyses that covered 2993 km^2 of Oregon and Washington (Montgomery et al., 1998). These watersheds provide a representative sampling of industrial forest lands in the region; they encompass a wide range of topographic and geologic settings in glaciated and unglaciated terrain, and all of the watersheds had been logged over the periods of record.

We used maps of landslides compiled from aerial photographs during watershed analyses conducted by a private timber company using coverage from the earliest available (typically from the 1940s) to the most recent (generally the 1990s). Our own examination of selected aerial photographs and landslide maps from several of the watershed analyses revealed that many small landslides were not mapped. Due to this underrepresentation and because aerial photograph analyses further undercount landslide frequencies under a forest canopy (Robison et al., 1999), this regional compilation provides a minimum rate for historic landsliding. Note, however, that road-related landslides account for about half of the mapped landslides (Montgomery et al., 1998).

In four of the watersheds, we used a 10 m grid DEM based on scanned contours from U.S. Geological Survey (USGS) 7.5′ topographic quadrangles. In the other watersheds we used USGS 30 m grid DEMs. To avoid the complication of variable landslide sizes in a regional analysis and isolate the topographic control on landslide initiation, we used ρ_s/ρ_w = 1.6, C_l = 0, C_b = 2 kPa, z = 1 m, and ϕ = 33° for all of the watersheds. Each mapped landslide was assigned to the minimum $(q/T)_c$ value predicted within its boundary. Because this approach introduces a bias toward low $(q/T)_c$ values, we conducted 10 simulations of an equivalent number of randomly placed landslides of comparable size for each watershed. These biased random simulations allow an independent test of model performance because if the model correctly identifies areas of greater landslide initiation potential, then we should observe higher rates of sliding in high hazard, low $(q/T)_c$ terrain than in the analysis of randomly placed slides.

The aggregated data set for all 14 watersheds reveals an inverse relation between $(q/T)_c$ and landslide density in potentially unstable portions of the landscape (Fig. 5), confirming that shallow landslides tend to occur in areas predicted to have low $(q/T)_c$ values. Comparison of mapped landslide frequencies with the randomly placed landslides reveals that landslides occurred two to three times more frequently than predicted by the random model for $\log(q/T)_c \le -2.8$, but that for locations with $\log(q/T)_c > -2.2$ factors other than the topographic control embodied in equation 2 must dominate landslide initiation. As terrain with $\log(q/T)_c < -2.8$ occu-

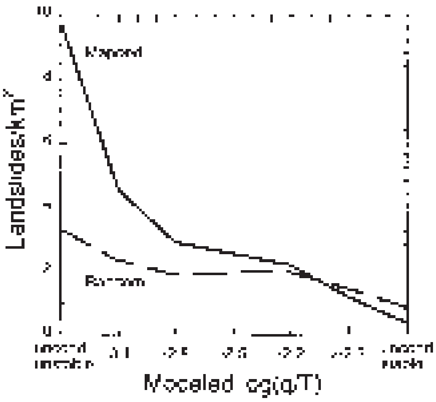

Figure 5. Mapped regional landslide rate based on 3224 landslides in Oregon and Washington and aggregated results of simulations of equal number of randomly placed landslides stratified by $(q/T)_c$ (see text).

pies only 13% of the total area of the study watersheds, the greatest landslide hazard is restricted to a small and topographically identifiable portion of the region.

A minimum regional constraint on the contemporary rate of sliding from hollows is the rate in areas where $\log(q/T)_c \le -2.8$, sites that at the 10–30 m grid scale of our analysis typically correspond to areas with well-developed topographic hollows. Dividing the aggregated 4.7 landslides km^{-2} observed for all areas where $\log(q/T)_c \le -2.8$ by the 50 yr time period spanned by the aerial photograph records yields 0.09 landslides km^{-2} yr^{-1}, or 3 to 9 times the background rate estimated here for the Oregon Coast Range (0.01–0.03 landslides km^{-2} yr^{-1}).

CONCLUSIONS

Our unique 10 yr record of landsliding from steep, slide-prone terrain and analysis of an extensive regional landslide survey document a discrepancy between contemporary and longer term rates of shallow landsliding in the Pacific Northwest. Together these complementary lines of evidence confirm that forest clearing increases regional landslide frequency. Our data from Mettman Ridge further show that small landslides triggered during common storms, and therefore overlooked by conventional monitoring, can dramatically increase sediment yield. Mettman Ridge is located in some of the steepest and least stable terrain in the Oregon Coast Range, and should have one of the highest landslide rates in the region. Therefore the dramatic acceleration of contemporary landslide rates at Mettman Ridge is not representative, but rather provides an estimate of the maximum impact in landslide-prone terrain in the region. However, the regional analysis, which provides a minimum estimate due to its reliance on aerial photograph interpretation, reveals a rate of three to nine times background rates for the region as a whole. The magnitude of the impact, however, should vary with local geology, storm history, and land-use practices. Our analysis also supports the hypothesis that at high apparent cohesion, landslides in steep terrain are restricted to locations with either excess pore pressures or thick soils (e.g., topographic hollows), but that landsliding occurs more widely across the landscape in areas with low apparent cohesion. This fundamental change in landscape dynamics is particularly relevant to long-term forest planning, especially where urban areas are extending into landslide-prone terrain.

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