Size and location of colluvial landslides in a steep forested landscape

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ABSTRACT An inventory of 61 landslide scars in part of the central California Coast Ranges is used to document sites of instability and to infer conditions necessary for failure. Scar dimensions cluster around widths of 7-10 m, lengths of 10-20 m, and depths of 0.7-1.1 m. Simple theoretical analyses indicate that root strength along the margins of a potential landslide imposes constraints on landslide size; typical scar size may reflect the size of a deposit required for instability at sites with typical vegetation, slope gradients, soil texture, and hydrology. Hollows are the main source of landslides, consistent with the convergence of shallow groundwater flow and the long-term accumulation of colluvium in hollows, but conditions of sufficiently thick soil and high pore pressures for failure are also attained on side slopes.

INTRODUCTION

A central problem in landslide studies is to determine where and when conditions for failure can be satisfied. In steep, well-vegetated, soil-mantled landscapes, roots provide strength that can greatly increase the stability of a site (e.g., Burroughs & Thomas, 1977; Gray & Megahan, 1981; Ziemer, 1981; Riestenberg & Sovonick-Dunford, 1983; Tsukamoto & Kusakabe, 1984; Wu, 1984). The probability of failure at a site should increase as soils thicken and root penetration into bedrock decreases (Dietrich & Dunne, 1978), and therefore failure should be most likely at sites where soil thickness progressively increases and where recurrent high pore pressures are produced. These conditions are met in hollows, where topographically induced convergence of colluvial debris and shallow groundwater flow occur, and is consistent with the common occurrence of landslides in hollows in many regions (see review in Reneau & Dietrich, 1987).

Steps have been made towards quantifying the conditions necessary for failure by relating site stability to increasing soil thickness over time (e.g., Okunishi & Iida, 1981; Shimokawa, 1984). Although instability in shallow soils is generally treated as a one-dimensional problem, approached by an infinite slope analysis, recent workers have recognized that a pure infinite slope model for failure is insufficient because of strength provided by roots along the margins of a potential failure (Burroughs & Thomas, 1977; Wu, 1984). Riestenberg & Sovonick-Dunford (1983), Burroughs (1984), and Tsukamoto & Kusakabe (1984) have included such edge effects in slope stability analyses. Implicit in their stability equations are limitations on the size of a colluvial deposit that is stable under given conditions. Observations of landslide scars suggest that scar dimensions tend to have typical values in any area (e.g., Lehre, 1982; Tochiki, 1985), and this may record prevailing conditions of root strength, soil texture, hydrology, and slope gradient.

In this study, the distribution and characteristics of 61 recent landslides in a portion of the central California Coast Ranges are used to document sites of instability and to infer conditions necessary for failure. The dimensions of these landslide scars cluster around characteristic values, and simplified theoretical analyses will be used to address the controls on scar size, and implications for the location and timing of failure.
A 3.8 km$^2$ portion of San Pedro Ridge in the Coast Ranges of Marin County, California, was selected for detailed study. The area has a Mediterranean climate, with mean annual precipitation of about 700 mm. Elevations in the study area range from near sea level to 322 m. The vegetation is predominantly a mixed hardwood forest composed of California laurel (Umbellularia californica), coast live oak (Quercus agrifolia), and Pacific madrone (Arbutus menziesii); the vegetation is a native community, and slopes are mainly undisturbed by human activity. Bedrock consists of interbedded arkosic sandstone and shale of the Franciscan assemblage (Rice et al., 1976). The sandstone beds vary from 0.1 to at least 15 m thick, and are generally steeply dipping, isoclinally folded, and faulted.

Shallow landslides in colluvial soils were abundant in the study area during a major storm on 3-5 January 1982. These landslides mobilized as debris flows, causing extensive erosion in downslope canyons and destroying two houses along one third-order channel (Smith & Hart, 1982; Reneau & Dietrich, 1987). The storm was a discrete event lasting 24 to 36 hours, with 24-hour rainfall totals of from 230 to 380 mm in southern Marin County. A gauge at the Marin County Civic Center, 2 km east of the study area, recorded 258 mm in 24 hours. Antecedent rainfall was high before the storm, with 209 mm falling in the previous 17 days, and 78 mm in the previous 6 days at the Marin County Civic Center.

Rainfall intensity-duration-frequency relationships for the January 1982 storm were calculated for the Marin County Flood Control District by J. Goodridge (unpublished data). For two stations with the longest periods of record (26 and 27 years at Mill Valley and San Anselmo), the 12- and 24-hour intensities exceeded the estimated 100-year storm. Shorter duration intensities at these stations were less unusual, and 30-minute to 3-hour intensities did not exceed the 10-year storm. Locally, however, short-term intensities were substantially greater. At Nicasio Dam, 24 km east of the study area, a 30-minute intensity of 37 mm was recorded. This is thrice the 13 mm intensities recorded at other stations, exceeding the estimated 100-year 30-minute intensities for the Mill Valley and San Anselmo stations (26 and 29 mm, respectively). The 1-hour intensity at Nicasio Dam also exceeded the estimated 100-year event.

The landslides in the study area all occurred within colluvial soils, and, except in rare cases, the failures completely mobilized as debris flows. The soils are typically gravelly loams, with 5-15% clay and 10-30% gravel. Failure planes were either within the colluvium or near the bedrock-colluvium interface, as shown in Figure 1. Although roots are abundant in the landslide scarps, failures occurred below the depth of major rooting and few roots were observed crossing failure surfaces. The landslide scars are typically 10-20 m long, 7-10 m wide, 0.7-1.1 m deep, and less than 200 m$^3$ in volume. Data on scar size and topographic setting are shown in Figure 2 and Table 1. Additional landslide data from Marin County, collected by Lehre (1982), are also shown in Table 1, and will be referred to later. Scar dimensions tend to be log-normally distributed (e.g., Lehre, 1982), and this biases average values towards the larger scars. The mean of log-transformed data more accurately represents typical scar size, and both these and arithmetic means are shown in Table 1.

Colluvium exposed within eight landslide scars on San Pedro Ridge has been dated using radiocarbon analysis of charcoal, providing ages of 9000-25,000 years BP for the lower portions of these deposits (Reneau et al., 1986, and unpublished dates). Two dated cross sections are shown in Figure 1. The dates place upper limits on the time since major evacuation last occurred at each site, although unconformities may be present and landsliding therefore more frequent (Reneau et al., 1986). Relative soil profile development suggests that significantly older colluvium, with redder hues and higher clay content, is exposed in only three 1982 landslide scars.
Colluvial landslides

FIG. 1 Scar cross-sections, showing location of radiocarbon dates. (a) Lindenwood Court landslide, and (b) Mosquito landslide (San Rafael 2 and 5 sites of Reneau et al., 1986).

FIG. 2 Histograms of (a) scar volume, (b) scar length, (c) average scar depth, (d) average scar width, and (e) average slope gradient, showing topographic setting of each scar. Data for 61 landslides from January 1982 storm. Convergence angle, $\beta$, is illustrated, and study area location is shown.
TABLE 1 Landslide dimensions from Marin County, California (data in m; scars in brush and grass from Lone Tree Creek basin, Lehre, 1982)

<table>
<thead>
<tr>
<th>Vegetation</th>
<th>n</th>
<th>Arithmetic Mean</th>
<th>Mean of Log-Transformed Data</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>length</td>
<td>width</td>
</tr>
<tr>
<td>mixed hardwood forest</td>
<td>61</td>
<td>20.2</td>
<td>8.8</td>
</tr>
<tr>
<td>northern coastal scrub</td>
<td>46</td>
<td>19.1</td>
<td>6.8</td>
</tr>
<tr>
<td>grass</td>
<td>22</td>
<td>11.3</td>
<td>5.6</td>
</tr>
</tbody>
</table>

LANDSLIDE LOCATION

In order to provide a quantitative comparison of site topography, an angle of topographic convergence is used here, defined as the angle between the orientation of the hollow axis and the orientation of the adjacent side slopes (Fig. 2). In this paper, hollows are arbitrarily grouped as either distinct, with convergence angles of roughly 30-50°, or subtle, with angles of 10-30°. Convergence angles into hollows rarely exceed 50°, and below 10° concavities are very difficult to recognize in the field; as such, slopes with little or no convergence are classed as side slopes.

Within the study area, hollows were the most important source of landslides, accounting for 62% of the scars by number and 77% by volume. Failures occurred in both distinct and subtle hollows, and at all positions longitudinally within hollows. While 31% of failures in hollows extended to the base of the swale, consistent with the higher pore pressures expected here (e.g., Pierson, 1980), failure in the remainder occurred upslope, producing debris flows that passed over the lower hollow. Twenty one percent were restricted to the uppermost portion of hollows where topographic convergence is weak, with headscars occurring as close as 10 m below ridgecrests. Many older, degraded scars are also present this close to ridges, roughly where the slope steepens to 30° or greater below ridgetop convexities. This suggests that the minimum distance from a ridge where failure can occur may be defined primarily by the local ridgecrest convexity, and not by hydrologic constraints.

The largest scars occurred in distinct hollows, but if the very largest are excluded, no major variation in size with topographic setting is evident (Fig. 2). The four largest scars, accounting for 40% of the total scar volume, occur within 0.4 km of each other, suggesting that they were triggered by an exceptionally intense, localized rainfall cell. The rain gauge at Nicasio Dam indicates that local intensities could be much higher than experienced at most other stations in the region. Close temporal association of major debris flows on Inverness Ridge, Marin County, during the January 1982 storm, may record another such rainfall cell (Ellen, in press).

The presence of many failures on side slopes indicates that conditions of sufficiently thick soils and elevated pore pressures can be met in areas of little or no topographic convergence. Pockets of thicker colluvium within bedrock irregularities or local depressions appeared to exist at several of the side slope scars, but only two were clearly at sites of exceptionally thick colluvium. Headscars are located as close as 10 m below ridgecrests, as observed in subtle hollows, suggesting that sufficient water for failure may, in some cases, be provided by rainfall directly onto the site, with little contribution of subsurface runoff from upslope. This is consistent with storm rainfall totals and average site conditions. Soil porosity is approximately 0.5, and a field capacity of about 0.25 is reasonable for loamy soils (e.g., Dunne & Leopold, 1978). For a 0.9 m thick soil column, at field capacity, only 225 mm of rain is then needed to reach complete saturation if drainage is insignificant, similar to the 24-hour rainfall total of 258 mm recorded nearby.
Typical hydraulic conductivities are on the order of $1 \times 10^{-3}$ cm s$^{-1}$ for these forest soils. If flow is restricted to the soil matrix, water would drain less than 1 m downslope over a 24-hour period with this hydraulic conductivity, and less than 10 m for a 10-fold increase in conductivity.

Within hollows, roughly half of the failure planes are located within the colluvium, and half at or near the bedrock-colluvium interface (e.g., Fig. 1). The development of failure planes within the colluvium may be due to subtle increases in the factor of safety with depth. Bulk density of colluvium in the study area progressively increases with depth, and this change in density causes an increasing factor of safety with depth, as illustrated in Figure 3. Because of this, failure planes should develop as high in a deposit as possible, and the actual height is probably controlled by the depth of significant root penetration. The increasing bulk density with depth may also result in increases in the friction angle, as reported by Gray and Megahan (1981), similarly increasing the factor of safety with depth. Although direct shear tests of the colluvium at two scars in Marin County revealed no major strength variations with depth, variations in the friction angle could be within measurement errors.

**EFFECT OF DEPOSIT SIZE ON STABILITY**

Strength is added to soils by roots that extend both vertically and laterally across the boundaries of a potential landslide. While the vertical decrease in roots may control the depth of the failure plane, strength from lateral roots may help determine the size of a deposit required for failure. As a first approximation, the effect of increasing deposit size on stability can be shown quantitatively by using an infinite slope analysis modified by adding root strength along the margins. In the following calculations, several simplifying assumptions are made: 1) the colluvium is completely saturated, with hydrostatic pore pressures; 2) the basal shear surface is immediately below the rooting zone, and strength here is entirely due to soil friction (no root strength or soil cohesion); and 3) roots along the entire perimeter of a deposit contribute strength equally. The effects of lateral pressure, passive pressure at the downslope end, and arching, included by Burroughs (1984), are not included here to allow a simple analytical solution; as long as failure depth is not

\[
F = \frac{(\rho_s - \rho_w) \tan \theta}{\rho_s \tan \phi}
\]

where

- $\rho_s$ = average density of saturated soil column
- $\rho_w$ = density of water
- $\phi$ = friction angle
- $\theta$ = effective angle of friction
- $\rho_s = \rho_B - \rho_H$

**FIG. 3** Bulk density data from Lindenwood Court landslide scar (Fig. 1a), and changes in the ratio $\rho_s - \rho_w / \rho_s$ with depth. The factor of safety, $F$, is directly proportional to the ratio $\rho_s - \rho_w / \rho_s$ under conditions of complete saturation, no cohesion, and no root strength, causing $F$ to increase with depth. Failure plane at depth of 1.0 m in this scar.
significantly deeper than the rooting depth, the equation of Burroughs (1984) provides similar results to the following much simpler equation. By modifying equation (15) of Riesterberg & Sovonick-Dunford (1983), we can express the factor of safety, \( F \), as

\[
F = \frac{A (\gamma_s - \gamma_w) \cos \Theta \tan \Phi + \text{Sr} \ P}{A \ \gamma_s \ sin \Theta}
\]

(1)

where \( A \) is deposit area, \( \gamma_w \) and \( \gamma_s \) are the unit weights of water and saturated soil, \( \Theta \) is slope gradient, \( \Phi \) is the soil friction angle, \( \text{Sr} \) is the strength of roots along the margins, and \( P \) is perimeter length. Defining \( A \) as length \( (l) \times \) width \( (w) \), and \( P \) as \( 2(l + w) \), where deposit length and width refer to colluvium thicker than the main rooting zone, allows equation (1) to be rewritten as

\[
l = \frac{2 \ \text{Sr} \ kw}{w - 2 \ \text{Sr} \ k} \quad \text{or} \quad w = \frac{2 \ \text{Sr} \ kl}{l - 2 \ \text{Sr} \ k}
\]

(2)

where

\[
k = \left[ \gamma_s \ cos \Theta \ (F \ tan \Theta - \left( \frac{\gamma_s - \gamma_w}{\gamma_s} \ tan \Phi \right)) \right]^{-1}
\]

(3)

Using equation (2), any length-width ratio is theoretically possible for a landslide, and Figure 4a shows two solutions of this equation for \( F = 1.0 \). A family of curves exists for varying conditions of soil friction, slope gradient, and root strength. A decline in root strength, such as after major fires or logging, would shift a \( F = 1.0 \) curve down to the left, along the \( l/w = 1.0 \) line, allowing failure of deposits with lower lengths and widths. Low gradients or high soil friction provide curves representing higher lengths and widths.

![FIG. 4](image)

(a) Length vs. width of potentially unstable deposits. Each curve represents possible length-width combinations at \( F = 1.0 \) for varying \( \Phi, \Theta, \text{Sr} \), and pore pressures. (b) Length vs. width of 38 landslide scars; scar depth = 0.75-1.2 m, and \( \Theta = 31-40^\circ \). Solid curve passes through mean of log-transformed data \( (w = 8.7 \ m; l = 14.0 \ m) \); \( \text{Sr} \) of 13.1 kN m\(^{-2} \) calculated using \( \Theta = 36^\circ, \Phi = 41^\circ, \) and \( \rho_{\text{dry}} = 1250 \) kg m\(^{-3} \). Dashed lines represent +/- 20% range in \( \text{Sr} \). (c) Factor of safety vs. deposit width for a typical site. Stippled area represents range in \( F \) for +/- 20% range in \( \text{Sr} \).
Pore pressures in excess of hydrostatic can be produced by flow through bedrock (e.g., Wilson et al., 1986), and this would allow failure of smaller deposits than would occur under hydrostatic pore pressure conditions.

Equation (2) imposes minimum lengths and widths, of equal magnitude, on potential landslides, although no maximum values are imposed. In nature, actual lengths and widths are probably limited by topographic and hydrologic constraints (Fig. 4a). Minimum landslide area along a $F=1.0$ curve occurs at $l/w=1.0$, and might be considered the most probable size, but the depositional zone in hollows is elongate and deposits tend to have a much greater length than width. Only a portion of the total length of a hollow tends to fail in each event, and the area with sufficient thickness and width to fail may be constrained by prior landsliding upslope or downslope, or by the presence of bedrock steps. Additionally, pore pressures may only be high enough for failure along a portion of the total hollow length.

Figure 4b shows length-width data for 38 landslide scars in the study area that have typical slope gradients and depths. The solid curve is plotted through log-average values of length and width, and the two dashed curves enclose 82% of the scars. Deleted from this plot are three compound scars; scar shapes on these suggest the merging of separate scars, and apparently do not represent initial failure. Length-width ratios range from 0.6-4.2, and 76% have ratios of 1.0-2.5; arithmetic mean is 1.8 and mean of log-transformed data is 1.6. The size of each scar may record the first point on a $F=1.0$ curve that is reached as each deposit lengthens and widens over time, and the relatively small spread in $l/w$ may be due to similar topographic and hydrologic constraints.

Precise prediction of landslide size is difficult due to uncertainties in root strength, friction angle, and pore pressure conditions at each site. For example, root size and density, hence root strength, is undoubtedly variable beneath a forest, and a range in average $S_r$ should be expected. The two dashed curves in Figure 4b represent a +/- 20% change in $S_r$ from the calculated average value of 13.1 kN m$^{-2}$, and variations in root strength of this magnitude may account for much of the variation in scar size. The same curves would result if $S_r$ was constant and $\Phi$ varied from 34-45°, although the actual range in $\Phi$ may be much less. Bedrock control on scar location, and the possibility of artesian pore pressures, is suggested at many sites, and may also contribute to the variation in scar size. Two of the scars in the lower left portion of Figure 4b, plotting off the general trend, showed evidence of local bedrock control, while lower vegetation density is suggested at the other two. Differences in slope gradient should also be very important, but no trend in size with varying gradient is apparent within the data in Figure 4b. In addition, use of the slope stability equation of Burroughs (1984) indicates that landslide size should be a function of failure depth. A thicker rooting zone, forcing failure planes to be deeper, should result in longer and wider landslides, although the data in Figure 4b show no systematic relation between size and depth.

Figure 4c shows the relationship of $F$ and deposit size at a typical site, assuming a length-width ratio of 1.6, illustrating the decrease in factor of safety as a deposit enlarges over time. The curves in Figure 4c represent minimum values of $F$ at a given width under conditions of complete saturation and non-artesian pore pressures. The width at which a deposit can fail is sensitive to variations in root strength, slope gradient, soil friction angle, and deposit shape. For example, the width at failure could range from 7.0-10.5 m for a typical site if $S_r$ varied by +/- 20% around the average value from Figure 4b (Fig. 4c).

Although the values of root strength in Figure 4 should be considered approximate due to the simplifying assumptions that were made, published values of root strength through a vertical section of soil show these estimates to be reasonable. Riestenberg & Sovonick-Dunford (1983) estimate values of 6.2-7.0 kN m$^{-2}$ along landslide scarps in an Ohio sugar maple forest, and Burroughs & Thomas (1977) estimate potential strengths of 8.3 and 16.7 kN m$^{-2}$ for Douglas fir forests in central Idaho and western Oregon, respectively. In addition, Ziemer (1981) reports that roots of northern California hardwoods
are stronger than Douglas fir, suggesting that strength values in this hardwood forest could be relatively high.

DISCUSSION

The similarity in size of most landslides in the San Pedro Ridge study area suggests that instability occurs after a critical combination of deposit length and width is exceeded, with scar sizes reflecting typical soils, vegetation, slope gradients, and pore pressure conditions. The hypothesis that the size of the deposit required for failure, and therefore scar size, is related to local vegetation, is supported by landslide data elsewhere in Marin County where bedrock is similar. Lehre (1982) reports that average scar widths and lengths are smaller in grasslands than in brush, and these are exceeded by scars in the hardwood forest (Table 1). In New Zealand, Selby (1976) also reports that failures in forests are larger than in grasslands. Variations in scar size in any area may partially reflect local variations in vegetation, with the smallest scars occurring in areas of low root density.

The actual timing of failure is dependent on site conditions and stochastic variations in rainfall, with certain combinations of antecedent moisture and storm intensity required for failure. Once deposits have reached a size where failure is possible, the frequency of a hydrologic event sufficient to trigger failure should vary with topographic setting, local hydrology, and deposit size. The frequency of such a storm should also decrease at a site over time due to increasing deposit size; larger deposits require more water for any level of saturation, hence a less frequent event. If a site has sufficiently low gradient, or if high permeability layers exist that can keep a deposit drained, a storm of such low frequency may be required for failure that deposits can attain sizes much greater than normal. Drainage into bedrock fractures can also occur (e.g., Harr & Yee, 1975; Wilson et al., 1986), and this should help maintain the stability of some sites. Dating of colluvium from landslide scars on San Pedro Ridge suggests that the frequency of landsliding from a site has the same time scale as major climatic changes (Reneau et al., 1986), and significant changes in storm frequency may also occur over the period of accumulation.

Colluvial deposits in hollows that greatly exceed the 1.5-2.5 m thickness typical of recently failed deposits are common on many steep slopes in the San Francisco Bay area, including the Marin Headlands deposits discussed by Reneau & Dietrich (1987). The existence of these deposits suggests that hydrologic conditions have not been adequate to trigger failure at all sites, and that although there is a tendency for failure in steep hollows as a deposit thickens, local site conditions can allow relatively long-term stability. The local presence of colluvium much older than removed by landslides in 1982, occupying bedrock hollows with no topographic expression, also records long-term stability. Stability of the thick deposits is an important practical problem because these deposits can produce exceptionally large debris flows and can be major sediment sources; for example, the largest 1982 failure in the study area accounted for 21% of the total scar volume.

The failure of these thick deposits may be related to unusual, local circumstances, such as progressive channel head failure, secondary failures due to debris flows generated upslope, or exceptional rainfall intensity. The largest failure in the study area occurred in a deposit that reached 7 m in maximum thickness, including 4 m of stratified, openwork gravels. These gravels may have kept the deposit drained and inhibited failure. Failure of the thick downslope section may have been triggered by a debris flow generated from the shallower section upslope. Several debris flows in the study area apparently triggered landslides down-swale or down-channel, possibly due to rapid undrained loading during passage of the debris flows. Secondary failures were seen within many other 1982 debris flow paths in Marin County (e.g., Ellen, in press). The
next deepest failure in the study area similarly had a section of coarse openwork gravels at the base; failure here may have been due to artificially high runoff provided by road drainage. As discussed earlier, spatial clustering of the largest scars in the study area suggests exceptional local rainfall conditions; included are the two sites with openwork gravels discussed above, and the scar with the lowest gradient. Progressive channel head retreat seems a viable mechanism for evacuating other thick deposits in this area. For example, a 1974 failure in a 4.5 m thick colluvial deposit in the Lone Tree Creek basin, Marin County, was preceded by gully head retreat at the downslope end; progressive failure is suggested by an unusually small cross-sectional area for the debris flow at the lower end of the scar, despite a large scar size (Lehre, 1982; site map in Reneau et al., 1986).

CONCLUSIONS

Landslides on San Pedro Ridge have a characteristic size, and this may record the critical deposit size required for instability at typical sites. Simple theoretical analyses indicate that landslide size should be a function of root density and strength, soil friction, slope gradient, and pore pressure conditions. Any length-width ratio is theoretically possible for given site conditions, although length-width ratios are typically 1.0-2.5 and may be determined by local topographic and hydrologic constraints. At sites where storms capable of triggering failure are common, failure should occur once a deposit reaches specific dimensions, and the frequency of failure is then determined by the depositional rate of colluvium. Most failures occur in hollows, where topographic convergence leads to long-term deposition of colluvium, but adequate accumulations of colluvium and sufficiently high pore pressures can also occur on side slopes. Development of deposits much larger than required for failure may record unusual site conditions that promote stability; eventual failure can produce exceptionally large debris flows, and stability of these remains an important practical problem.

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