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Debris-flow initiation from large, slow-moving landslides

M.E. Reid & D.L. Brien
U.S. Geological Survey, Menlo Park, CA, USA

R.G. LaHusen
U.S. Geological Survey, Vancouver, WA, USA

J.J. Roering
University of Oregon, Eugene, OR, USA

J. de la Fuente
U.S. Forest Service, Yreka, CA, USA

S.D. Ellen
Redwood City, CA, USA

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ABSTRACT: In some mountainous terrain, debris flows preferentially initiate from the toes and margins of larger, deeper, slower-moving landslides. During the wet winter of 1997, we began real-time monitoring of the large, active Cleveland Corral landslide complex in California, USA. When the main slide is actively moving, small, shallow, first-time slides on the toe and margins mobilize into debris flows and travel down adjacent gullies. We monitored the acceleration of one such failure; changes in velocity provided precursory indications of rapid failure. Three factors appear to aid the initiation of debris flows at this site: 1) locally steepened ground created by dynamic landslide movement, 2) elevated pore-water pressures and abundant soil moisture, and 3) locally cracked and dilated materials. This association between debris flows and large landslides can be widespread in some terrain. Detailed photographic mapping in two watersheds of northwestern California illustrates that the areal density of debris-flow source landsliding is about 3 to 7 times greater in steep geomorphically fresher landslide deposits than in steep ground outside landslide deposits.

1 INTRODUCTION

Debris flows initiate in a variety of geomorphic settings. They can mobilize from shallow landslides originating on steep slopes (Campbell 1975, Iverson et al. 1997), from landslides in topographic swales or hollows (Reneau & Dietrich 1987), from mobilized stream channel materials (Hungr et al. 1984, Takahashi 1991), or from other processes (Johnson 1984). Some researchers have noted that debris flows can initiate from the toe or margins of larger, deeper landslides (Morton & Campbell 1974, Swanson & Swanston 1977), but these events are often viewed as localized occurrences.

In this paper, we focus on debris-flow initiation from larger landslides. First, we present three seasons of hydrologic and displacement data, collected in real-time, that illustrate the movement behavior of a large landslide complex in California, USA. Debris flows are spawned from the margins of this complex when the main slide is actively moving downslope at a slow rate; we identify the conditions conducive to debris-flow initiation in this setting. Then, we examine whether debris-flow sources can be related regionally to the presence of larger landslide deposits. We selected two areas of northwestern California that have been sculpted by both numerous large landslides and debris flows. Using multiple sets of aerial photographs taken over approximately 60 years, we mapped debris-flow sources and large landslide deposits. Based on this extensive mapping, we then used a geographic information system (GIS) to analyze the spatial relations between debris-flow sources and large landslide deposits.

2 MONITORED DEBRIS-FLOW INITIATION FROM AN ACTIVE LANDSLIDE

The sinuous Cleveland Corral landslide complex, located in the Sierra Nevada Mountains, California, USA, extends about 450 m downslope and terminates in a toe perched about 50 m above U.S. Highway 50 (Fig. 1a). This landslide has moved many times in the past, including the winter seasons of 1996, 1997, 1998, 1999, and 2000. Two nearby large landslides, within 3 km of the Cleveland Corral landslide, have failed catastrophically, one in April 1983 (Kuehn & Bedrosian 1987) and the other in January 1997 (Sydnor 1997); both slides temporarily dammed the South Fork of the American River and blocked Highway 50. In addition, the 1997 landslide partially transformed into a debris flow, enhancing its travel distance. The Cleveland Corral landslide provides an opportunity to capture the transition from slow movement to rapid catastrophic failure and mobilization into a debris flow. Since February 1997, we have monitored the movement of the Cleveland Corral landslide using real-time data acquisition systems and repeated high-precision ground surveys. Although catastrophic failure of the entire slide has not occurred, we have observed both slow movement of large masses and rapid failure of small pieces within the complex. We were able to capture the progression of one small mass at the toe of the complex from slow movement to rapid failure to the generation of a debris flow. In this section, we first discuss the landslide and our monitoring efforts, then describe the slow slide movement that creates conditions conducive to small debris-flow initiation from the margins of the complex. We then describe the factors leading to this debris-flow initiation and illustrate the behavior with real-time data collected during transformation.

The Cleveland Corral landslide complex is located in an area where over 600 landslides have been mapped along a 24 km corridor of Highway 50 (Spittler & Wagner 1998). Sliding in the Cleveland Corral landslide complex involves only colluvium and older landslide deposits, but it is located near an underlying contact between schist and diorite bedrock. Varying in width from 25 to 70 m, the complex includes several internal scarps and toes as well as subsidiary flank slides (primarily on the NW margin) and levee deposits adjacent to the downslope section (Fig. 1a). Ground surface slopes in the complex typically vary between about 10° and 30°, but locally steepen to over 40° (Fig. 1b). To the west, slope deposits mantle a steeper slope leading to a small intermittent stream (Fig. 1b); these deposits contain colluvium and older slide debris that has spilled over from movement of the main slide, and small shallow failures separate from the main complex. The main slide material typically contains 10-15% cobble-sized clasts of schist and a matrix composed of clay (14-33%), silt (18-47%), and sand (27-61%). Material from the primary slip surface, exposed in lateral flanks near the downslope part of the slide, is highly sheared and clay rich, with a liquid limit of 36 and a plastic limit of 22 (weight % water content). Based on two shallow seismic surveys, landslide materials vary in thickness from 5 to 10 m, with the main slip surface located just above the underlying bedrock.

We have recorded hydrologic conditions and landslide movement at the Cleveland Corral landslide using U.S. Geological Survey (USGS) real-time data acquisition and telemetry systems originally developed for remote monitoring of active volcanoes. Information from a variety of sensors installed in or on the central region of the landslide are transmitted via radio telemetry to USGS computers. Data from this site are available in near real-time to the public over the internet at http://landslides.usgs.gov/hwy50 (Reid & LaHusen 1998). Extensometers, anchored to the ground surface outside of the landslide and oriented in the direction of slide movement, record the amount of downslope surface displacement. Geophones, with a wider dynamic response than standard earthquake seismometers (LaHusen 1996), are buried near the landslide to monitor ground vibrations associated with slide movement. A rain gage monitors precipitation at the site, while buried pressure transducers monitor groundwater pore pressures at different depths within the slide. Sensors are sampled every second and data are routinely transmitted every 15 minutes. Some of this monitoring was conducted in collaboration with the California Department of Transportation (Reid & LaHusen 1998). We also measured the spatial variability in movement patterns over the entire complex using repeated annual high-precision ground surveys (Fig. 1a).

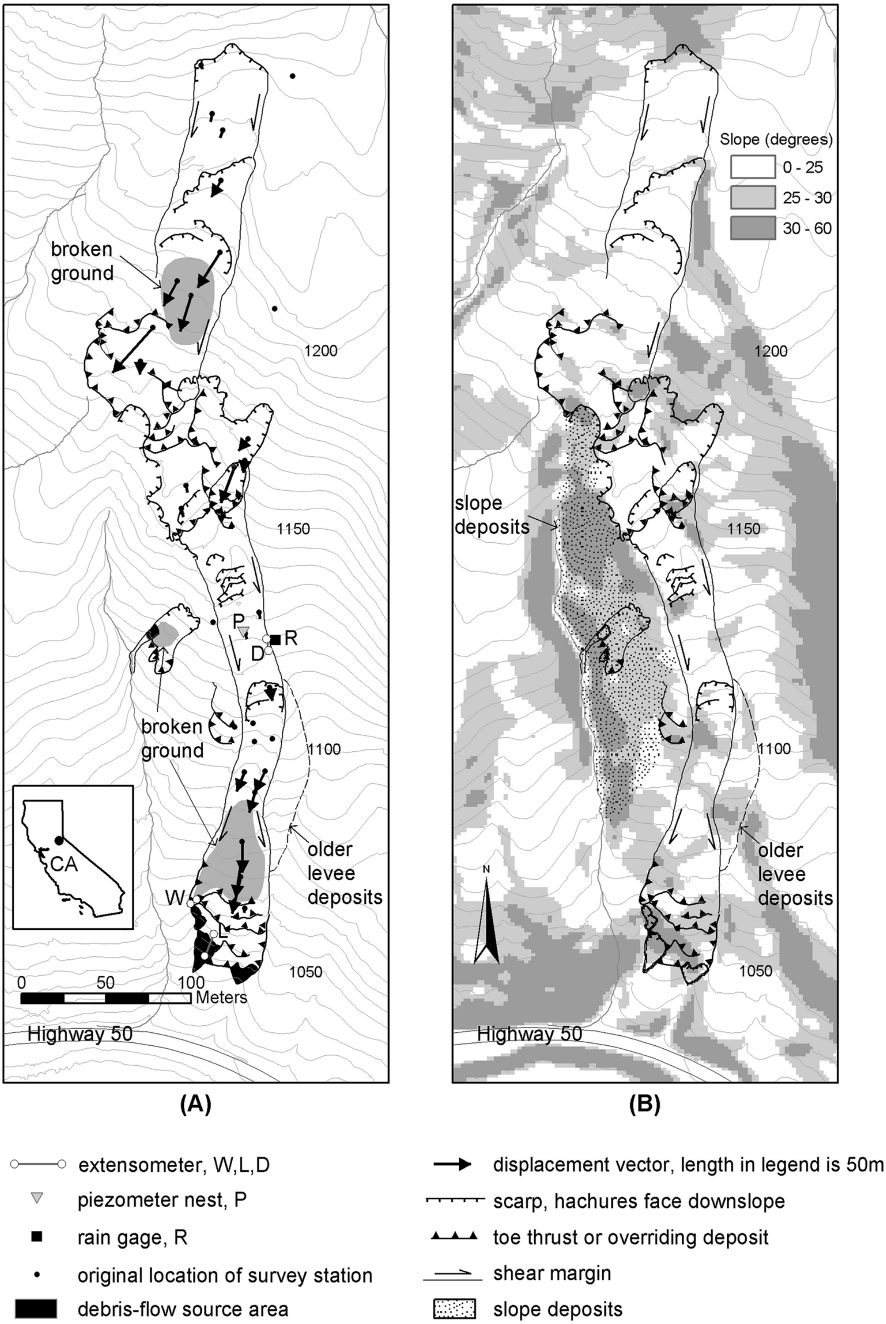

Figure 1. Maps of the Cleveland Corral landslide complex. Topographic base map is from aerial photography April 1999. Small inset shows site location in California, USA. a) Major surface structural features of the landslide complex, areas of highly broken ground (gray areas) containing numerous sub-meter blocks, locations of real-time monitoring instruments, and horizontal displacement vectors measured at monitoring stations for the time interval Aug. 1996 to June 2000. Numerous smaller structural features have been mapped but are not shown here. Mapping symbols follow Fleming & Johnson (1989). b) Slope map of the landslide area, derived from a 2.5 m DEM, and locations of western slope deposits.

2.1 *Movement behavior in the active landslide complex*

We observed that debris-flow activity affected only small parts of the Cleveland Corral complex between 1996 and 2000, and that under specific conditions slow movement of the main slide appears to enhance conditions conducive for secondary debris-flow initiation. The entire landslide complex remains stationary during the dry summer months and begins moving only after large amounts of winter precipitation elevate subsurface pore-water pressures. The rainy season in central California typically extends from late October to early April, and nearly all of the year's precipitation occurs during this time. Mean annual rainfall at the National Weather Service station at Pacific House, located about 9 km from the landslide, is 1350 mm, based on 51 years of record. Most precipitation on the landslide occurs as rain; colder winter storms may bring snow but this commonly melts within a few days. During the initial wetting period in the autumn, pore-water pressures increase until landslide movement begins. Slow movement continues as long as pore pressures remain elevated. Rapid movement of shallow debris-flow sources may occur during particularly wet times in the late winter and early spring. Precipitation eventually tapers off in the late spring, and pore pressures decline and movement ceases in early summer. Much of the slide is unsaturated in the late summer until autumn rains start the cycle again.

Figure 2 illustrates response patterns over three wet seasons. During the wet El Niño-influenced winter of 1997-1998, surface movement at the downslope toe began in late January and ended in mid-June (Fig. 2a); the central area started and ended movement a few weeks later than the toe. During 1998-1999, movement began in late February and ended in late April (Fig. 2b). During 1999-2000 only about 1200 mm of rain fell, and only a few cm of movement occurred in the central region (Fig. 2c). Because the wet season lasted longer in 1997-1998 than in the other two seasons, cumulative movement was greater during this season. Our observations indicate that about 800 mm of antecedent rainfall is usually required to initiate movement in the downslope part of the landslide. When the slide is actively moving, a pressure head greater than about 120 cm is present in the deeper piezometer, located 2.4 m beneath the ground surface in the central region of the complex.

Cumulative displacement at surveying monuments on the landslide, obtained by repeated annual surveys between 1996 and 2000, is shown in Figure 1a. Some of our surveying monuments were destroyed by construction in 1997, therefore some vectors (particularly those in the upslope section) represent a minimum displacement. During this period, the largest displacements occurred in both an upslope toe region and a downslope toe region (see broken ground areas on Fig. 1a) and in smaller failures on the NE margin. These areas of large displacement, but without flow behavior, commonly resulted in areas of pervasively broken ground, containing numerous sub-meter size blocks separated by open extension fractures. Localized slumping, in areas sloping ~25° to 35°, slightly steeper than the surrounding ground, appeared to create the broken ground.

We observed that both the timing and distribution of ground movement varied within the complex. For example, the upslope region of the landslide moved up to 12 meters downslope during 1996-97, while the central and downslope toe regions moved only about 1 meter. In contrast, during 1997-98 the upslope and central regions moved only about 1 meter, while the toe moved over 10 m. The timing and rate of movement also varied within a wet season. During 1997-98, the downslope region began moving in late January with velocities averaging about 4 cm/day, whereas the central region began moving in mid-February, with an overall velocity of about 0.5 cm/day. Once moving, however, the central region of the slide accelerated rapidly to 1-2 cm/day within a few days following precipitation; even rainfall events as small as 50 mm caused acceleration (see for example late May 1998 in Fig. 2a). This suggests that the slide is in finely balanced equilibrium during these times, and that movement rate is influenced directly by even small variations in subsurface pore-water pressures induced by rainfall.

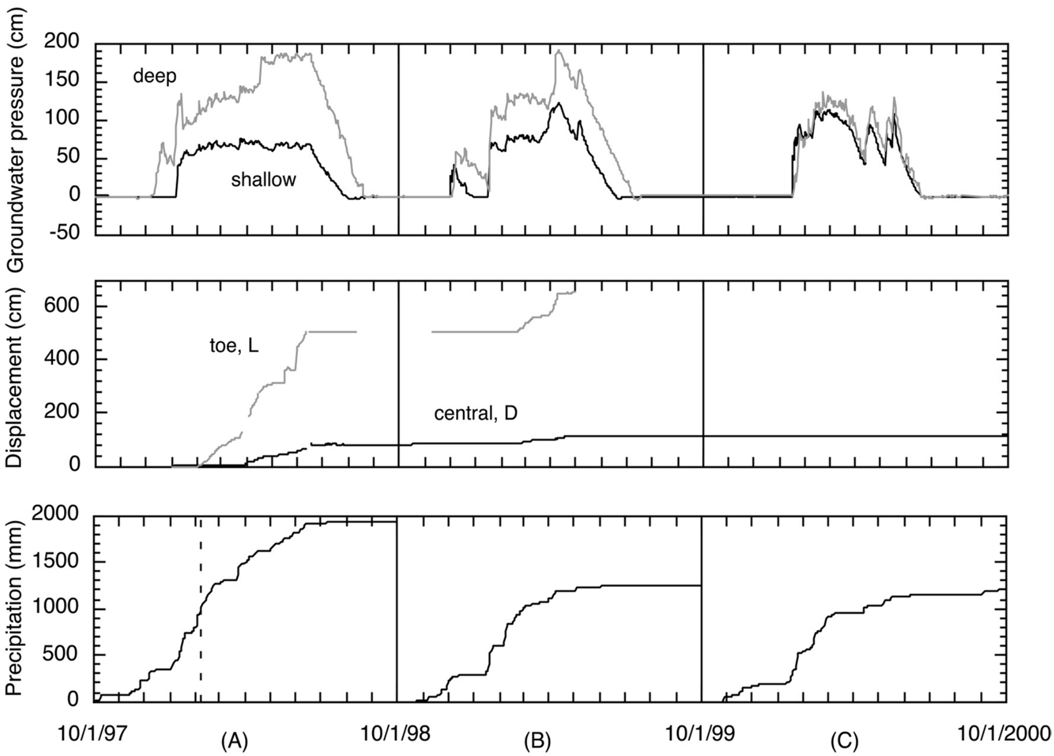

Figure 2. Hydrologic and displacement responses for three years (a, b, c.) at the Cleveland Corral landslide complex. Each set of graphs shows cumulative precipitation for a water year (starting Oct. 1), cumulative surface displacements on the toe (extensometer L) and in the central region (extensometer D), and groundwater pore-pressure head responses in shallow (1.8 m depth) and deep (2.4 m depth) piezometers (at piezometer nest P in the central region). Extensometer at toe (L) was discontinued May 1999. Vertical dashed line denotes time of small rapid failure shown in Figure 3. Instrument locations are shown in Figure 1a.

2.2 *Conditions conducive to secondary debris-flow initiation*

During wet seasons, we observed that small debris flows initiated from shallow failures in two settings: (1) on the downslope toe of the main slide complex, and (2) in slope deposits west of the main slide (Fig. 1a). During the wet 1998 El Niño-influenced winter, three shallow (about 1 m thick) failures on the toe partially mobilized into debris flows. Another failure occurred during the spring of 2000 in the western slope deposits; part of this mass transformed into a debris flow that deposited material for about 100 m along a small channel. All of these rapid failures occurred during prolonged rainy periods while the main slide was actively moving. Three additional conditions were present at these sites: 1) locally steepened (~35° to 42°) ground created either by dynamic landslide movement oversteepening the downslope toe or by earlier emplacement of disrupted materials on the steep western slope (Fig. 1b), 2) elevated pore-water pressures and abundant soil moisture, and 3) localized, effectively dilated materials, as evidenced by pervasively cracked ground, particularly at the downslope toe. These conditions can enhance debris-flow generation (Iverson et al. 1997, Iverson et al. 2000). In addition, the shallow source failures deposited material directly into small gullies that likely had transient surface-water flow during rainy periods; this may have further enhanced debris-flow generation. However, because no streams exist immediately at the base of the initial failures, they were not triggered directly by stream undercutting. Shallow failures also occurred elsewhere on steep margins of the slide complex (such as along the NE margin), but they did not mobilize into debris flows. These sites did not appear to have locally dilated materials, nor did they lead directly to channels.

Beginning late on February 5, 1998, and extending into February 6 during a prolonged rainy period, extensometer W (Fig. 1a) captured the behavior of a 64-m^3 shallow, first-time failure as it slumped from the downslope toe of the main slide and partially mobilized into a debris flow. Over the same period the toe also moved, but at a much slower rate (Fig. 2a). As the 64-m^3 mass began to accelerate, the displacement record shows an abrupt jump of about 2 cm, perhaps due to partial toppling of the extensometer (Fig. 3a). Following this jump, the mass continued to displace with instantaneous velocities eventually exceeding 100 cm/day and accelerations exceeding 50 cm/day^2 with brief spikes over 4000 cm/day^2 (Fig. 3b). These values exceed the critical thresholds of 5 cm/day or 0.5 cm/day^2 indicating impending rapid failure in schist, as suggested by Salt (1988). Following this relatively short-lived rapid displacement, extensometer data became erratic and halted. Subsequent field visits discovered the damaged extensometer on a remnant piece of the failure mass. About half of the mass had entered a small gully at the base of the failure and thin debris-flow deposits extended from this base about 50 m down the margins of the gully. Displacement data from extensometer W can be used to attempt to predict, after the fact, the time to failure using reciprocal of velocity (1/velocity) relations originally developed by Saito (1965) and later expanded upon by others (Fukuzono 1990, Varnes 1983, Voight 1989). Figure 3a shows 1/velocity during the period leading to rapid failure. Although there are oscillations in these data, for most of the six hours prior to rapid failure the change in 1/velocity is approximately linear. Extrapolating this downward overall trend results in a predicted failure time about 2 hours sooner than actually occurred. In this case, monitoring 1/velocity could have provided some precursory indication of the time of failure hours prior to the event.

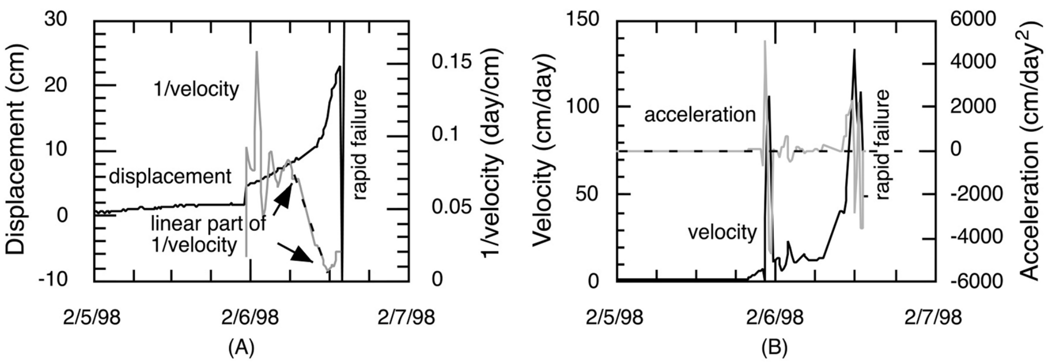

Figure 3. Response leading to rapid failure of a small, 64 m^3 block at toe of the Cleveland Corral landslide complex. Part of this mass transformed into a debris flow. a) Displacement measured by extensometer W. 1/velocity shows downward trend for about 6 hours prior to failure. b) Velocity and acceleration computed from displacement record of extensometer W.

3 WIDESPREAD DEBRIS-FLOW INITIATION ASSOCIATED WITH LARGE LANDSLIDE DEPOSITS

Our studies of the Cleveland Corral landslide illustrate that debris-flow initiation can be directly related to the activity of a larger, slow-moving landslide. We selected two study areas in the Coast Range, about 350 km northwest of the Cleveland Corral landslide, to investigate whether debris-flow initiation from larger landslides might be a widespread, regional phenomenon and not just restricted to isolated cases. Both rapid debris flows and slow-moving landslides are abundant in the Coast Ranges and Klamath Mountains of northwestern California (de la Fuente et al. 1998, Kelsey 1980). In this heavily forested region, weak rocks, large earthquakes, vigorous rainstorms, and ongoing tectonic uplift combine to destabilize most hillsides over time. Our two study areas are located in watersheds contributing to the lower section of the Eel River; they are within about 8 km of each other and are located at approximately the same latitude. Although these areas have likely been subjected to similar destabilizing forces, they differ in topography, underlying rock, and landslide style.

One study area, encompassing the watersheds of Bear and Jordan Creeks (an area of about 34 km^2), is underlain largely by Coastal terrane rocks of the Coastal belt of the Franciscan Complex, consisting of variably disrupted, interbedded sandstone and argillite (McLaughlin et al. 2000). This area contains angular topography with extensive ground steeper than 33°, the result of incision by deep canyons (some with steep inner gorges) and an array of steep sidehill drainages. The area has a history of debris-flow activity triggered by large rainstorms (Pacific Watershed Associates 1998, 1999). The second study area, covering a 43 km^2 piece of the Larabee Creek watershed, is underlain by Yager terrane of the Coastal belt of the Franciscan Complex, which also consists of bedded sandstone and argillite, parts of which are disrupted (McLaughlin et al. 2000). Here, however, apparently mechanically weaker rocks underlie a less regular topography consisting primarily of moderate and gentle slopes. Large areas of irregular, poorly incised topography with slopes <20° appear to be underlain by thick masses of predominantly argillite. Lesser amounts of steep ground are underlain by packets of thick-bedded sandstone. Previous mapping has identified numerous debris flows and landslide deposits in both areas (Spittler 1983) as well as evidence for uncommonly large landslide deposits (McLaughlin et al. 2000). Although the two study areas differ in underlying rock types, within each study area lithologies are similar.

3.1 *Landslide activity in the northwestern California study areas*

We used multiple sets of vertical aerial photography to delineate both large landslide deposits and debris-flow source locations in the two study areas. Sets of aerial photographs taken in 1963, 1966, 1983, 1984, and 1997 were used to map the Bear/Jordan study area; sets taken in 1941-42, 1968, 1980, 1984, 1991, and 2000 were used to map the Larabee-Yager study area. Although photo identification of debris slides or large landslide deposits can be limited by photo scale and vegetation cover (Pyles & Skaugset 1998), examining a series of photos from different times provides a historical record of minimum debris-slide activity triggered by a number of rainstorms. Moreover, the different years of photography revealed different parts of the study area deforested by timber harvest at different times and helped us identify landslide deposits and rate their recency or geomorphic freshness. Both historical activity and geomorphic freshness can be difficult to determine exclusively from limited modern field visits in heavily forested terrain.

For each set of aerial photographs, we mapped debris-flow initiation areas on a 1:24,000 topographic map base to obtain a record over time. These initiation areas are typically rapid, shallow translational landslides, herein called debris slides (Cruden & Varnes 1996). Not all debris slides leave photographic evidence of extended flow, and many located near streams deliver sediment directly to the channel without creating a flow track. Care was taken to distinguish source slides from flow paths; nearly all source areas were large enough to be mapped as polygons (minimum size 100 to 300 m^2). Furthermore, we attempted to correlate debris slides with the presence of any nearby logging roads, tractor yarding skid trails, or landings that existed at the time of initial debris sliding. These features can instigate debris sliding through cuts or fills, or by concentration of water runoff. If definite or probable spatial relations between slide and road existed, we considered the debris slide to be potentially road related. In this paper, our analyses use only non-road related debris slides, to remove possible road influences on debris-slide initiation.

We also mapped landslide deposits formed by deeper and slower movement, such as translational slides, rotational slides, and earth flows, as well as head and lateral scarps where large enough to be mappable. These deposits, when active, tend to move progressively and intermittently, often in seasonal increments, over periods of years, decades, centuries, or possibly millennia (Swanson & Swanston 1977) and may not be currently active. Recency of movement can be difficult to measure quantitatively without on-site monitoring, but it can be inferred approximately from the geomorphic freshness of landslide features (Keaton & DeGraff 1996). For photo interpretation of recency of movement, we rated the geomorphic freshness of each landslide deposit or element of a landslide complex using qualitative measures of the apparent freshness of headscarps, lateral scarps, toes, and internal topography, ranging from fresh and well defined to subtle or absent. Overall freshness was then determined from the preponderance or most convincing freshness of these specific observations. Debris-slide activity on the larger landslide deposits was not used as an indicator of freshness. The ratings were then grouped into two categories, fresher and more subdued landslide deposits.

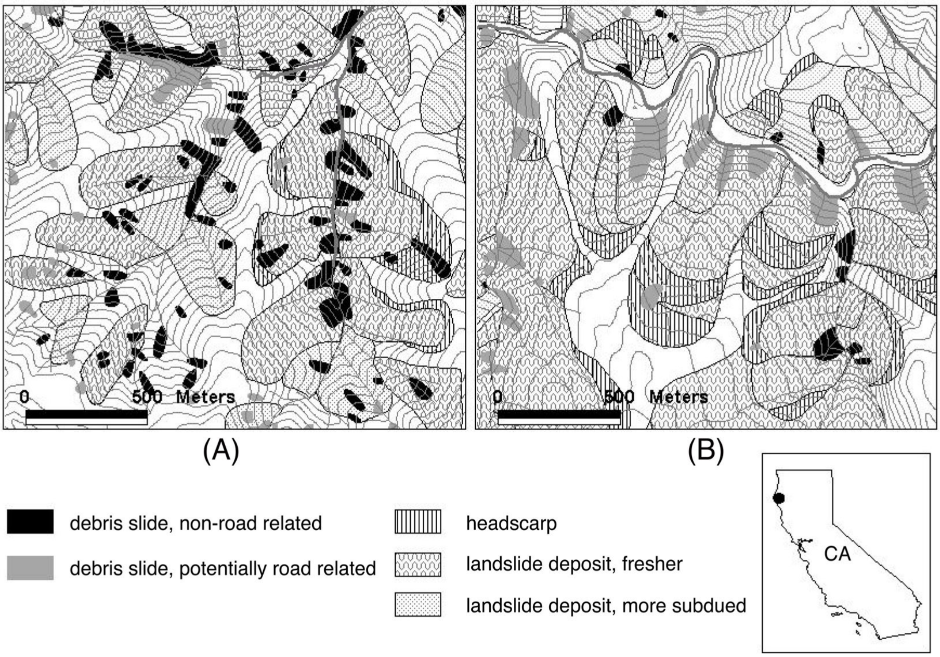

Figure 4. Mapped landslides in two study areas in northwestern California, showing non-road-related debris slides, potentially road-related debris slides, geomorphically fresher large landslide deposits, more subdued landslide deposits, and scarps where mappable. Inset shows location of study areas within California. a) Part of Bear and Jordan Creek study area. b) Part of Larabee Creek study area. Note that many of the mapped debris slides in both areas occur on the toes of larger landslide deposits, adjacent to streams.

Both study areas have had widespread debris-flow activity over the time period evaluated and also contain numerous large landslide deposits. Maps in Figures 4a and b show small pieces of each study area that illustrate several consistent patterns. Movement has produced extensive landslide deposits in each study area, exceeding that mapped by earlier investigators. In the Bear/Jordan study area, over 300 different large landslide deposits underlie about 65 % of the ground; in the Larabee-Yager at least 200 large deposits underlie about 71 % of the ground. Although the two study areas differ in rock type and topography, these numbers indicate that both landscapes have been modified by large landslides. Most of the mapped debris slides occur at the toes, and to a lesser extent at the lateral margins and interiors, of the large landslide deposits, which generally are adjacent to steam channels. Many debris slides near channels are larger than previously mapped inner gorges (Spittler 1983).

The style of large landslide deposits varies between the two study areas. In the Bear/Jordan study area, these deposits are usually discrete masses and typically occupy the steep slopes that line the major canyons (Fig. 4a). The average ground slope in these deposits is about 27° (obtained from a 10-m digital elevation model). In the Larabee-Yager study area, most of the ground consists of irregular, moderate to gentle slopes composed of blanket-like complexes of landslide deposits that can extend for 1-3 km from ridge to stream. Other areas underlain by Yager consist of steep ground having few landslide deposits or moderate to steep ground that includes discrete landslides and landslide complexes (shown in Fig. 4b). In this study area, average ground slope in the landslide deposits is about 20°.

3.2 *Relation between debris-slide activity and fresher landslide deposits*

Visual evaluation of the study area mapping suggests that non-road-related debris slides are located preferentially within large landslide deposits. To quantify possible relations, we computed the areal density of debris slides in different land categories. We present results from analyses restricted to ground with slopes >33° (using a 10-m DEM); debris slides are more likely on such ground than on gentler slopes. Areal density is here defined as the ratio of debris-slide area to the total area within a given land category. Land categories chosen for analysis were: 1) geomorphically fresher landslide deposits, 2) more subdued landslide deposits, and 3) hillslopes outside landslide deposits.

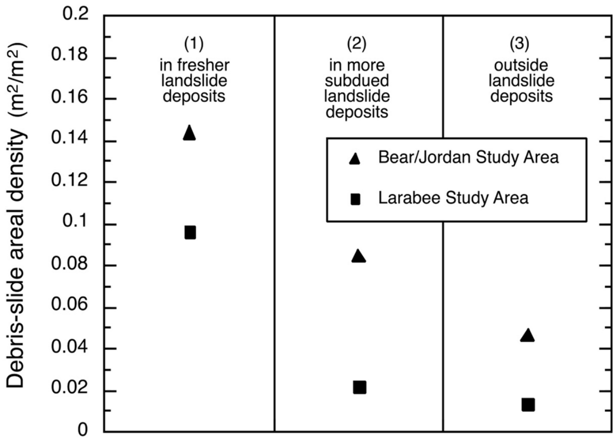

Figure 5. Areal density of non-road-related debris slides for the two study areas in northwestern California, over the time periods analyzed. Densities shown are within steep (>33°) parts of (1) geomorphically fresher landslide deposits, (2) more subdued landslide deposits, and (3) outside landslide deposits. Areal density is the ratio of the area affected by debris sliding to the total area within a given land category.

Several patterns are evident in Figure 5 for both study areas. Within each study area, the highest areal density of debris sliding occurs in steep, geomorphically fresher landslide deposits (category 1), with lower densities in more subdued landslide deposits (category 2), and even lower densities in steep ground outside landslide deposits (category 3). Areal densities in fresher deposits are about 3 (in Bear/Jordan) to 7 (in Larabee-Yager) times greater than in steep ground outside landslide deposits. Despite the shorter time period analyzed for Bear/Jordan study area, the areal densities of debris sliding for all three categories in this study area are greater than in the Larabee-Yager area.

Although the differences in areal density between categories appear significant in Figure 5, we further examined the differences in probabilities using a one-sided chi-square, χ^2, statistical test (Conover 1980). A one-sided test is useful here because rejection of the null hypothesis indicates that the areal density in one category is greater than in the other category. For each study area, the statistical tests indicate that debris-slide areal density is higher in fresher landslide deposits than in more subdued landslide deposits, and that the density is higher in more subdued deposits than in other ground outside landslide deposits; in all cases the tests reject the null hypothesis at a 95 % confidence level. Thus, these tests suggest that the patterns evident in Figure 5 are significant.

Large, slow-moving landslides might predispose slopes to localized debris-flow initiation by reducing the shear strength or dilating materials through deformation, by modifying hydrologic patterns and increasing subsurface pore-water pressures, by locally steepening slopes, or by changing near-surface stress distributions within the larger slide. However, areal densities alone do not directly indicate causality—either that large landslide deposits cause debris slides or vice versa. These densities do, nevertheless, indicate differences in the likelihood of debris sliding. Some nearby areas, such as the Freshwater Creek watershed, do not appear to show strong associations between debris-slide locations and large landslides deposits (Falls 1999). However, elevated areal densities in the two northwestern California study areas, combined with our study of activity at the Cleveland Corral landslide, suggests that larger landslides can focus the location of debris sliding and resulting debris flows in a variety of terrain.

4 CONCLUSIONS

Our observations of the active Cleveland Corral landslide complex and of mapped debris-flow sources and larger landslide deposits in northwestern California lead to the following conclusions:

(1) Debris flows can initiate from the toes and margins of larger landslides. At the Cleveland Corral landslide complex, secondary debris flows occurred during rainy periods when the main slide complex was active and moving slowly. Three additional conditions appear conducive to debris-flow mobilization from the shallow failures at this site: 1) locally steep (~35° to 42°) ground, particularly on the toe oversteepened by dynamic landslide movement, 2) elevated pore-water pressures and abundant soil moisture, and 3) locally cracked and dilated materials. Failure into nearby small gullies may also have aided the transformation into flow.

(2) Using real-time data acquisition at the Cleveland Corral complex, we monitored the rapid acceleration of a small first-time failure on the slowly moving toe of the larger slide. This small failure reached instantaneous velocities >100 cm/day before it mobilized into a debris flow and flowed down a gully. Monitoring rapid increases in velocity and acceleration could provide precursory indication of impending rapid failure.

(3) The link between debris-flow sources and large landslides is not restricted to an isolated landslide. Aerial photographic mapping of debris-flow sources (debris slides) and larger landslide deposits in two study areas of northwestern California reveals that most debris flows have originated from larger landslide deposits. The bedrock materials and style of large landsliding differ between the two areas, but both landscapes have abundant large landslide deposits.

(4) In both northwestern California study areas, the areal density of non-road related debris sliding from steep (>33°) hillslopes, over the time periods analyzed, is greater within geomorphically fresher large landslide deposits than within more subdued deposits or ground outside landslide deposits. Fresher deposits have better defined head scarps, lateral scarps, toes, and internal topographic features than subdued deposits, suggesting more recent activity.

(5) In three diverse settings, large landslide deposits focus the location of shallow landslides that transform into debris flows. Identification of both temporal and spatial relations between large landslides and debris flows is important to understanding debris-flow initiation mechanisms and to hazard assessments.

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