POTENTIALLY UNSTABLE SLOPES AND LANDFORMS

IDAHO DEPARTMENT OF LANDS

State Forester Forum

OVERVIEW

Landslides occur naturally in forested basins and are an essential process in the delivery of wood and gravel to streams. Wood and gravel play significant roles in creating stream diversity that is necessary for fish use as habitat and spawning grounds. When the potential for instability is recognized, the likelihood that sediment and debris would travel far enough to threaten a public resource or public safety is considered. Many factors are part of that concern including initial failure volume and nature of a landslide, landslide runout distance, and landscape geometry.

Landslide Types and Effects

Shallow landslides occur in bedrock hollows, convergent headwalls, and inner gorges with slopes, on toes of deep-seated landslides with slopes, and on the outer edges of meander bends. There are generally three types of shallow landslides: debris slides, debris flows, and hyper-concentrated floods. They are distinguished from each other by the ratio of water to solids contained in them.

Debris slides consist of aggregations of coarse soil, rock, and vegetation that lack significant water and move at speeds ranging from very slow to rapid down slope by sliding or rolling forward. The results are irregular hummocky deposits that are typically poorly sorted and non-stratified. Debris slides include those types of landslides also known as shallow rapid, soil slips, and debris avalanches. If debris slides entrain enough water, they can become debris flows.

Debris flows are slurries composed of sediment, water, vegetation, and other debris. Solids on average compose >60% of the volume. Debris flows usually occur in steep channels, as landslide debris becomes charged with water (from soil water, or on entering a stream channel) and liquefies as it breaks up. These landslides can travel thousands of feet from the point of initiation, scouring the channel to bedrock in steeper channels. Debris flows commonly slow where the channel makes a sharp bend and stop where the channel slope gradient becomes gentler than about 3 °, or the valley bottom becomes wider and allows the flow to spread out.

Hyper-concentrated floods are flowing mixtures of water, sediment (dominantly sand-sized), and organic debris with solids that range between 20% and 60% by volume (Pierson and Scott, 1985). In forested mountains, they are commonly caused by the collapse of dams, such as those formed by landslide dams or debris jams. Impounded water and debris released when the dam is breached sends a flood wave down the channel that exceeds the magnitude of normal floods. Such hyper-concentrated floods can rise higher than normal rainfall or snowmelt-induced flows along relatively confined valley bottoms, driving flood waters, sediment, and wood loads to elevations high above the active channel and, if present, the active floodplain.

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Debris flows and hyper-concentrated floods can occur in any unstable or potentially unstable terrain with susceptible valley geometry. In natural systems, debris flows and hyper-concentrated floods caused by dam-breaks are responsible for moving sediment and woody debris from hill slopes and small channels down into larger streams. But debris flows can also cause damage to streams by scouring channel reaches, disturbing riparian zones, impacting habitat and dumping debris onto salmonid spawning areas. Debris flows can cause elevated turbidity, adversely affect water quality downstream, threaten public safety, and damage roads and structures in their paths (Figure 1).

Figure I. Road-initiated debris flows in inner gorges, main stem Beaver Creek, North Fork Clearwater River.

At least two of these visible torrent tracks were active within the last several years, and sediment and debris appears to have been delivered across the lower road to Beaver Creek. The debris slides originated at an existing road at the gorge break-in-slope and torrented down the steep inner gorges (swales).

Deep-Seated Landslides

A more detailed explanation of deep-seated landslides is covered later in this section because deep-seated landslides are also landforms. Despite the failure mechanism, deep-seated landslides are those in which the slide plane or zone of movement is well below the maximum rooting depth of forest trees (generally greater than 10 feet) and may extend to hundreds of feet in depth often including bedrock. Deep-seated landslides can occur almost anywhere on a hill slope and are typically associated with hydrologic responses in permeable geologic materials overlying less permeable materials. The larger deep-seated landslides can often be identified from topographic maps or aerial photos.

Certain key areas of deep-seated landslides may be sensitive to forest practices. The bodies and toes of deep-seated landslides are made up of incoherent collapsed material weakened from previous movement and therefore may be subject to debris slide and debris flow initiation in response to harvest or road building. Sediment delivery from shallow landslides on steep stream-adjacent toes of deep-seated landslides and steep side-slopes of marginal streams on the bodies of deep-seated landslides is common.

SLOPE FORM

Slope shape is an important concept when considering the mechanisms behind shallow land sliding. Understanding and recognizing the differences in slope form is key in potentially unstable landform recognition. There are three major slope forms to be observed when looking across the slope (contour direction): divergent (ridge top), planar (straight), and convergent (spoon-shaped) (Figure 2). Landslides can occur on any of these slope forms but divergent slopes tend to be more stable than convergent slopes because water and debris spread out on a divergent slope whereas water and debris concentrate on convergent slopes. Convergent slopes tend to lead into the stream network, encouraging delivery of landslide debris to the stream system. Planar slopes are generally less stable than divergent slopes but more stable than convergent slopes. In the vertical direction, ridge tops are convex areas (bulging outward) and tend to be more stable than planar (straight) mid-slopes and concave areas (sloping inward) (Figure 3).

Additionally, slope steepness can play a significant role in shallow land sliding. Steeper slopes tend to be less stable. The soil mantle, depending upon its make-up, has a natural angle at which it is relatively stable (natural angle of repose). When hill slopes evolve to be steeper than the natural angle of repose of the soil mantle, the hill slope is less stable and more prone to shallow landslides, especially with the addition of water. The combination of steep slopes and convergent topography has the highest potential for shallow land sliding.

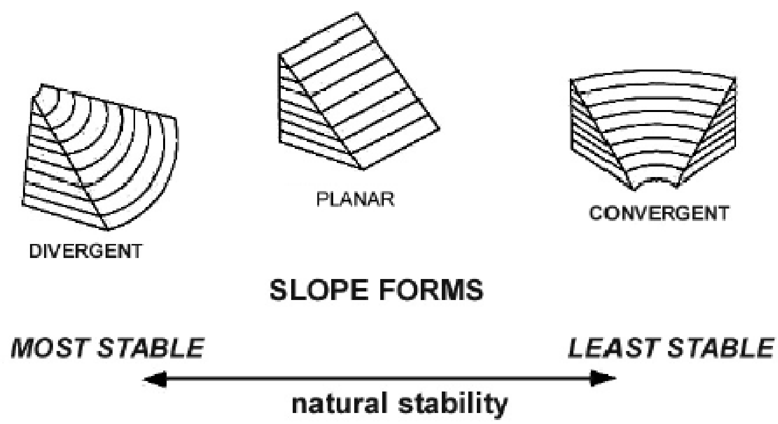

Figure 2. Slope configurations as observed in map view

This figure shows three major slope forms (divergent, planar, and convergent) and their relative stability. These slope form terms are used in reference to contour (across) directions on a slope. Convergent areas with slope greater than 60% are the most shallow landslide-prone.

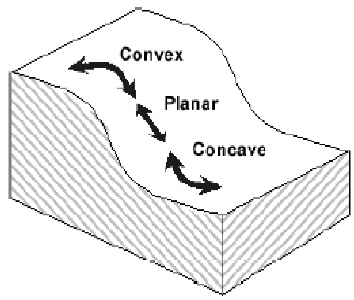

Figure 3. Slope configurations as observed in profile: convex, planar, and concave. These terms are used in reference to up and down directions on a slope (Drawing: Jack Powell, WDNR, 2004)

DESCRIPTION OF UNSTABLE AND POTENTIALLY UNSTABLE LANDFORMS AND PROCESSES

Areas of unstable landforms can usually be identified with a combination of topographic and geologic maps, aerial photographs, CWE mass failure hazard rating maps, and modeled slope stability morphology (SHALSTAB, SINMAP, LISA) output maps. However, field observation is normally required to precisely delineate landform boundaries, gradients, and other characteristics.

Bedrock Hollows, Convergent Headwalls, Inner Gorges

These three landforms are commonly associated with each other as shown in figures 4 and 5.

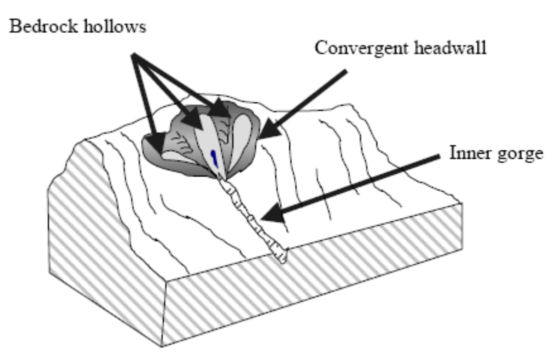

Figure 4. Typical hill slope relationships between bedrock hollows, convergent headwall, and inner gorge (Drawing: Jack Powell, WDNR, 2003)

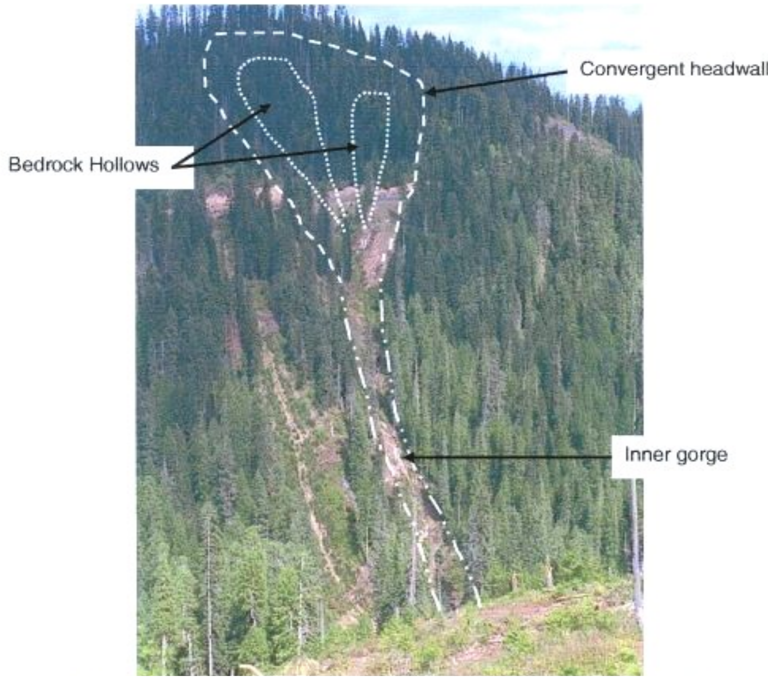

Figure 5. Common hill slope relationship: bedrock hollows in convergent headwalls Draining to inner gorges (Photo and drawing: Scott Marshall, IDL, 2005)

Bedrock hollows are also called colluvium-filled bedrock hollows, zero-order basins, swales, bedrock depressions, or simply hollows. Not all hollows contain bedrock so the term "bedrock" hollow can be a misnomer. Hollows are commonly spoon-shaped areas of convergent topography with concave profiles on hill slopes. They tend to be oriented linear upland down-slope. Their upper ends can extend to the ridge or begin as much as several hundred feet below ridge line. Most hollows are approximately 75 to 200 feet wide at their apex (but they can also be as narrow as several feet across at the top), and narrow to 30 to 60 feet downhill.

Hollows usually terminate where distinct channels begin. This is at the point of channel initiation where water emerges from a slope and has carved an actual incision. Steep bedrock hollows typically undergo episodic evacuation of debris by shallow-rapid mass movement, followed by slow refilling with colluvium that takes years or decades. Unless they have recently experienced movement by a landslide, hollows are partially or completely filled with colluvial soils that are typically deeper than those on the adjacent spurs and planar slopes. Recently evacuated hollows may have water flowing along their axis whereas partially evacuated hollows will have springs until they fill with sufficient colluvium to allow water to flow subsurface.

The common angle of repose for dry, cohesionless materials is about 36^0 (72%), and saturated soils can become unstable at lower gradients. Thus, slopes steeper than about 35^0 (70%) are considered susceptible to shallow debris slides. "Bedrock" hollows are formed on slopes of varying steepness. Hollows with slopes steeper than 70% (approximately 35^0) are potentially unstable in well-consolidated materials, but hollows in poorly consolidated materials may be unstable at lower angles.

Vegetation can provide the critical cohesion on marginally stable slopes and removes water from the soil through evapotranspiration. Leaving trees in steep, landslide-prone bedrock hollows helps maintain rooting strength and should reduce the likelihood of land sliding (Figure 6). However, wind-throw of the residual trees following harvest can be associated with debris slide or debris flow events. In high wind environments, it is essential to harvest in a manner that will limit the susceptibility of the residual trees to wind-throw as well as to reduce the potential for landslides (for example leaving wider strips, pruning or topping trees in the strips, or feathering the edges of reserve strips).

Figure 6. Example of leave areas protecting unstable slopes (Photo: Venice Goetz, WDNR, 2004)

Convergent headwalls are funnel-shaped landforms, broad at the ridge top and terminating where headwaters converge into a single channel. A series of converging bedrock hollows may form the upper part of a convergent headwall. Convergent headwalls are broadly concave both longitudinally and across the slope, but may contain sharp ridges that separate the bedrock hollows or headwater channels.

Convergent headwalls generally range from about 30 to 300 acres. Slope gradients are typically steeper than 70%. Unlike bedrock hollows, which exhibit a wide range of gradients, only very steep convergent landforms with an obvious history of landslides are called convergent headwalls. Soils are thin because landslides are frequent in these landforms. It is the arrangement of bedrock hollows and first-order channels on the landscape that causes a convergent headwall to be a unique mass-wasting feature. The highly convergent shape of the slopes, coupled with thin soils (due to frequent landslides), allows rapid onset of subsurface storm water flow.

Inner gorges are canyons created by a combination of stream down-cutting and mass movement on slope walls. Inner gorges are characterized by steep, straight or concave side-slope walls that commonly have a distinctive break in slope (Figure 7). Debris flows, in part, shape inner gorges by scouring the stream, undercutting side slopes, and/or depositing material within or adjacent to the channel (Figure 8). Inner gorge side-slopes may show evidence of recent landslides, such as obvious landslides, raw un-vegetated slopes, young, even-aged disturbance vegetation, or areas that are convergent in contour and concave in profile. Because of steep slopes and proximity to water, landslide activity in inner gorges is highly likely to deliver sediment to streams or structures downhill. Exceptions can occur where benches of sufficient size to stop moving material exist along the gorge walls, but these are uncommon.

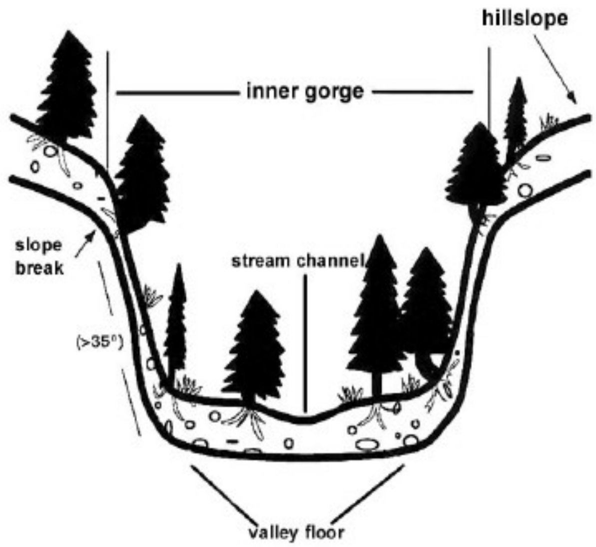

Figure 7. Cross-section of an inner gorge. This view emphasizes the abrupt steepening Below the break-in-slope (Drawing: Benda, et al, 1998)

Figure 8. Photograph showing how debris flows help shape features related to inner Gorges. (For example, V-shaped profile, buried wood, distinctive break in slope along Margins of inner gorge (Photo: Scott Marshall, IDL 2003)

The geometry of inner gorges varies. Steep inner gorge walls can be continuous for great lengths, as along a highly confined stream that is actively down cutting, but there may also be gentler slopes between steeper ones along valley walls. Inner gorges can be asymmetrical with one side being steeper than the other. Stream-eroded valley sides, which can be V-shaped with distinct slope breaks at the top, commonly do not show evidence of recent land sliding as do inner gorges which tend to be V-shaped.

<u>Other Indicators of Slope Instability or Active Movement</u>
In addition to the landforms above, other indicators of slope instability or active movement may include:

(a) topographic and hydrologic
- bare or raw, exposed, un-vegetated soil on the faces of steep slopes
- Boulder piles
- Hummocky or benched surfaces, especially below crescent-shaped headwalls
- Fresh deposits of rock, soil, or other debris at the base of a slope
- Ponding of water in irregular depressions or undrained swampy areas on the hill slope above the valley floor
- Cracks in the surface (across or along slopes, or in roads)
- Seepage lines or springs and soil piping
- Deflected or displaced streams (streams that have moved laterally to accommodate landslide deposits)

(b) vegetational
- jack-strawed, back-rotated, or leaning trees
- Bowed, kinked, or pistol-butted trees
- Split trees
- Water-loving vegetation (horsetail, skunk cabbage, etc.) on slopes
- Other patterns of disturbed vegetation

No one of these indicators necessarily proves that slope movement is happening or imminent, but a combination of several indicators could indicate a potentially unstable site.

Deep-seated landslides are those in which the slide plane or zone of movement is well below the maximum rooting depth of forest trees (generally greater than 10 feet). Deep-seated landslides may extend to hundreds of feet in depth, often including bedrock. Deep-seated landslides can occur almost anywhere on a hill slope and can be as large as several miles across or as small as a fraction of an acre. The larger ones can usually be identified from topographic maps or aerial photographs. Many deep-seated landslides occur in the lower portions of hill slopes and extend directly into stream channels whereas deep-seated landslides confined to upper slopes may not have the ability to deposit material directly into channels.

One common triggering mechanism of deep-seated landslides results from the over-steepening of the toe by natural means such as glacial erosion or fluvial undercutting, fault uplift, or by human-caused excavations. Initiation of such landslides has also been associated with changes in land use, increases in groundwater levels, and the degradation of material strength through natural processes. Movement can be complex, ranging from slow to rapid, and may include small to large displacements.

Deep-seated landslides characteristically occur in weak materials such as thinly layered rocks, unconsolidated sediments, deeply weathered bedrock, or rocks with closely spaced fractures. Deep-seated landslides can also occur where a weak layer or prominent discontinuity is present in otherwise strong rocks, such as clay or sand-rich interbeds in the basalts of central Idaho.

There are three main parts of a deep-seated landslide: the scarps (head and side), along which marginal streams can develop; the body, which is displaced slide material; and the toe, which also consists of displaced materials. The downslope edge of the toe can become oversteepened from stream erosion or from the rotation of the slide mass. A deep-seated landslide may have several of each of these parts because small deep-seated landslides can be found nested within larger slides. These three main parts are shown in Figures 9 and 10. The head-and side-scarps together form an arcuate or horseshoe shaped feature that represents the surface expression of the rupture plane. The body and toe area are usually hummocky and the flow path of streams on these landslide sections may be displaced in odd ways due to differential movement of landslide blocks. The parts of deep-seated landslides that are susceptible to shallow landslides and potential sediment delivery are steep scarps (including marginal stream side slopes) and toe edges.

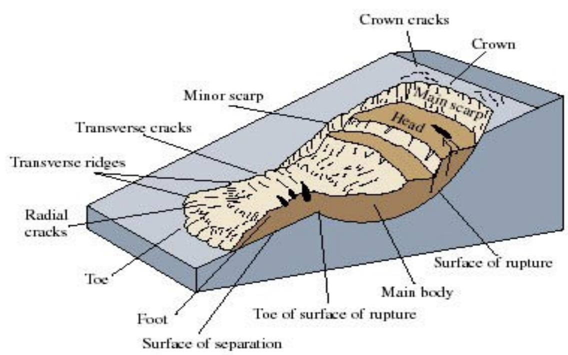

Figure 9. Rotational deep-seated landslide. Rotational displacement of blocks of soil commonly occur at the head of the landslide. Slow flow (an earth flow) may be found at the toe.

Figure 10. Deep-seated landslide showing the head scarp, body, and toe. Cecil Lake, British Columbia.

The sensitivity of any particular landslide to forest practices is highly variable. Deep-seated scarps and toes may be over-steepened and streams draining the displacement material may be subject to debris slide and debris flow initiation in response to harvest or road building. Movement in landslides is usually triggered by accumulations of water at the slide zone, so land use changes that alter the amount or timing of water delivered to a landslide can start or accelerate movement. Generally, avoiding the following practices will prevent most problems: destabilizing the toe by the removal of material during road construction or quarrying; overloading the slopes by dumping spoils on the upper or mid-scarp areas, or compacting the soil in these places which could change subsurface hydrology; and directing additional water into the slide from road drainage or drainage capture. The loss of tree canopy interception of moisture and the reduction in evapotranspiration through timber removal may also initiate movement of the slide.

Groundwater Recharge Areas of (Glacial) Deep-Seated Landslides

Groundwater recharge areas of deep-seated slides are located in the land up-slope that can contribute subsurface water to the landslide. In some cases this can include upslope portions of the landslide itself. Cemented soil horizons, fine-grained soils, and/or the presence of glacial till can be factors controlling the infiltration and flow of groundwater (Vaccaro et al., 1998). Groundwater perching and the characteristics of the overlying groundwater recharge area can be important factors in a deep-seated failure, especially for landslides in glacial sand and other unconsolidated sequences that overlie glacial-lake clay deposits or till (Figure 11).

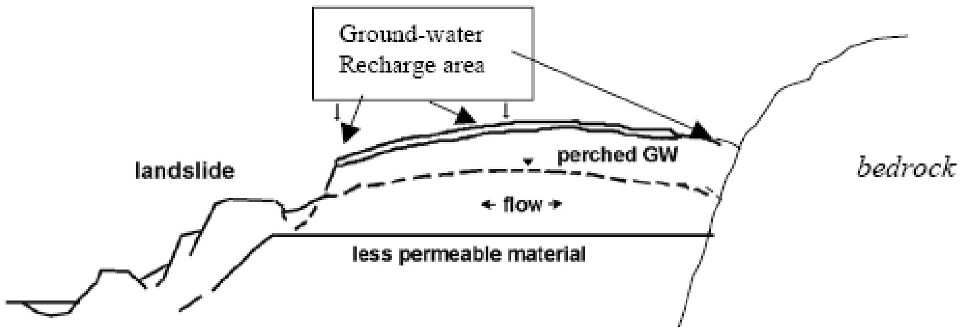

Figure 11. Groundwater recharge area for a glacial deep-seated landslide.

Outer Edges of Meander Bends

Streams can create unstable slopes by undercutting the outer edges of meander bends along valley walls or high terraces of an unconfined meandering stream (Figure 12 and 13). The outer edges of meander bends are susceptible to shallow land sliding including debris avalanching and small-scale slumping, and deep-seated land sliding. The outer edges of meander bends are protected by the stream protection zone (SPZ).

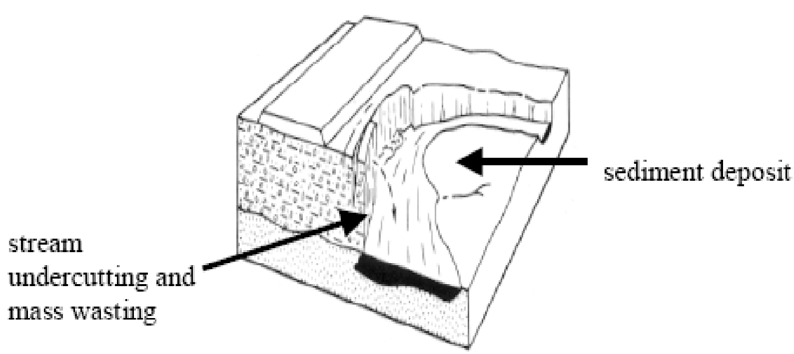

Figure 12. Outer edge of a meander bend showing mass wasting on the outside of the bend and deposition on the inside (adapted from Varnes, 1978).

Figure 13. Landslide on outer edge of meander bend on Pack River in northern Idaho, debris blocked river momentarily (Photo: Scott Marshall, 2002).

DELIVERY

Landslides occur naturally in forested basins and are an important process in the delivery of wood and gravel to streams. Wood and gravel play important roles in creating stream diversity that is essential for fish use as habitat and spawning grounds. When the potential for instability is recognized, the likelihood that sediment and debris would travel far enough to threaten a public resource or public safety should be considered. Many factors are part of that consideration including the initial failure volume of a landslide, the runout distance of a landslide, and landscape geometry.

It is difficult to prescribe guidelines for delivery distances because each situation has a special combination of process and topography. Deep-seated landslides can move anywhere from a few inches to a few miles depending on a friction of the slip plane, the forces pulling the landslides down, and the shear strength resisting those forces. Larger landslides are more likely to be able to move great distances at gentle gradients, but they are also less likely to be significantly affected by forest practices activities.

Timber harvest and road building can cause shallow landslides on steep slopes. Travel distances for such landslides depend on the amount of water contained in or entrained by them. Considering that rain, snowmelt, or some other extreme water inputs trigger the vast majority of landslides in the Pacific Northwest, it should be noted that almost all landslides contain some amount of water that tends to mobilize the soil or rock. Debris slides that do not reach streams usually deposit their debris on the hill slope; and are typically unable to move far across large areas of flat ground. However, since most landslides occur during storm conditions, a large proportion of debris slides do reach flowing channels and create the opportunity to entrain enough water to become debris flows. These flows are quite mobile, and can travel great distances in steep or moderate gradient channels.

Travel distance of a debris flow once it reaches a low-gradient surface is a function of its volume and viscosity. The solid volume of a debris slide or flow deposit is a function of soil depth, distance traveled down the hill slope, and the gradient of the traveled path. The proportion of water is the main control on viscosity. Field or empirical evidence should be used for determining the runout distance.

Even if the main mass of a landslide or debris flow comes to rest without reaching a public resource, there is the possibility that secondary effects may occur. Bare ground exposed by mass movement and disturbed piles of landslide debris can be chronic sources of fine sediment to streams until stabilized by revegetation. If flowing water (seepage, overland flow, or small streams) can entrain significant volumes of fine sediment from such surfaces, the possibility of secondary delivery must be evaluated, along with the likelihood of impact by the initial movement event itself.

To assess the potential for delivery and estimate runout distance, analysts can evaluate the history of landslide runout in the region, use field observations, and/or use geometric relationships appropriate from the scientific literature. In any situation where the potential for delivery is questionable, it is best to have a geotechnical expert examine the situation and evaluate the likelihood of delivery.

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