# Literature Synthesis of the Effects of Forest Practices on Non-Glacial Deep-Seated Landslides and Groundwater Recharge

**By: Daniel Miller**

July 2017

CMER # 2017.07.17

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## Washington State Forest Practices Adaptive Management Program

The Washington State Forest Practices Board (FPB) has established an Adaptive Management Program (AMP) by rule in accordance with the Forests & Fish Report (FFR) and subsequent legislation. The purpose of this program is to:

> *Provide science-based recommendations and technical information to assist the FPB in determining if and when it is necessary or advisable to adjust rules and guidance for aquatic resources to achieve resource goals and objectives. The board may also use this program to adjust other rules and guidance.* (Forest Practices Rules, WAC 222-12-045(1)).

To provide the science needed to support adaptive management, the FPB established the Cooperative Monitoring, Evaluation and Research (CMER) committee as a participant in the program. The FPB empowered CMER to conduct research, effectiveness monitoring, and validation monitoring in accordance with WAC 222-12-045 and Board Manual Section 22.

### Report Type and Disclaimer

This project development report was prepared for the Cooperative Monitoring, Evaluation and Research Committee (CMER), and was intended to support design and implementation of Forest and Fish Adaptive Management research and monitoring studies. The project is part of the Deep-Seated Landslide Program, and was conducted under the oversight of the Upslope Processes Scientific Advisory Group (UPSAG).

This document was reviewed by CMER but was not assessed through the Adaptive Management Program's independent scientific peer review process. CMER has approved this document for distribution as an official CMER document. As a CMER document, CMER is in consensus on the scientific merit of the document. However, any conclusions, interpretations, or recommendations contained within this document are those of the authors and may not reflect the views of all CMER members.



### Proprietary Statement



### Full Reference

Miller, Daniel. 2017. Literature Synthesis of the Effects of Forest Practices on Non-Glacial Deep-Seated Landslides and Groundwater Recharge. Cooperative Monitoring Evaluation and Research Report CMER #2017.07.17 Washington State Forest Practices Adaptive Management Program. Washington Department of Natural Resources, Olympia, WA.

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# Literature Synthesis of the Effects of Forest Practices on Non-Glacial Deep-Seated Landslides and Groundwater Recharge

Prepared for the Upslope Processes Scientific Advisory Group
Cooperative Monitoring, Evaluation, and Research Committee

July 17, 2017

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# 1 Introduction

## 1.1 Why this literature review and synthesis?

As described in the Request for Qualifications and Quotations (RFQQ) No 16-27 distributed by the Washington Department of Natural Resources (DNR) in the fall of 2015, in response to recent deep-seated landslide events, the Forest Practices Board had requested that the Timber Fish and Wildlife Policy Committee (Policy) develop recommendations related to the regulation of forest practices on glacial deposits and their associated groundwater recharge areas. Policy then directed the Upslope Processes Scientific Advisory Group (UPSAG) and the Cooperative Monitoring Evaluation and Research Committee (CMER) to develop a scope of work for a focused literature review and synthesis to update CMER on research assessing the effect offorest practices on groundwater recharge areas and deep-seated landslides in glacial materials. The review and synthesis provide a baseline for UPSAG to further develop an unstable slopes research strategy.

That review and synthesis (Miller, 2016) were completed in 2016 and are available at http://file.dnr.wa.gov/publications/bc_tfw_litsyngdsl_20170202.pdf. Subsequently, additional questions were raised by Policy, CMER, and UPSAG regarding groundwater recharge to non-glacial deep-seated landslides, the reactivation of dormant deep-seated landslides, and the run-out potential for deep-seated landslides. Thus, the literature review and synthesis were expanded to include non-glacial deep-seated landslides. This document is the synthesis of that review, and although this report can stand alone, it is complementary to and expands on information presented in the synthesis for glacial-deep-seated landslides.

Both the glacial and non-glacial reviews and synthesis documents expand on the literature review conducted for CMER by Koler (1992), who noted then the lack of research on deep-seated landslides in areas affected by continental glaciation and on effects of timber harvesting on deep-seated landslides in rock slopes. The current reviews on glacial and non-glacial deep-seated landslides specifically sought published research on these topics.

## 1.2 Deep-Seated Landslides and Washington's Forest Practice Rules

Washington's Forest Practice Rules include provisions to minimize forest-practice-related increases in landslide rates. These provisions are defined in the Washington Administrative Code (WAC), Section 222-16-050(1), which states that proposed activities involving "Timber harvest, or construction of roads, landings, gravel pits, rock quarries, or spoil disposal areas, on potentially unstable slopes or landforms described in (d)(i) of this subsection that has the potential to deliver sediment or debris to a public resource or that has the potential to threaten public safety" are "Class IV-special" forest practices. Class IV-special forest practices require an environmental checklist in compliance with the State Environmental Policy Act (SEPA).

SEPA policies for potentially unstable slopes and landforms are defined in WAC 222-10-030. These policies require certain analyses of potentially unstable slopes and landforms prior to approval of Class IV-special forest practices. These analyses must be performed by a qualified expert[1] and evaluated by Department of Natural Resources staff. The analysis must address the following three issues:

a) The likelihood that the proposed forest practices will cause movement on the potentially unstable slopes or landforms, or contribute to further movement of a potentially unstable slope or landform;
b) The likelihood of delivery of sediment or debris to any public resources, or in a manner that would threaten public safety; and
c) Any possible mitigation for the identified hazards and risks.

The DNR's evaluation must then determine if the proposed forest practices:

a) Are likely to increase the probability of a mass movement on or near the site;
b) Would deliver sediment or debris to a public resource or would deliver sediment or debris in a manner that would threaten public safety; and
c) Such movement and delivery are likely to cause significant adverse impacts.

If it is determined that the proposed forest practice is likely to have a probable significant adverse impact, then SEPA requires that "Specific mitigation measures or conditions must be designed to avoid accelerating rates and magnitudes of mass wasting that could deliver sediment or debris to a public resource or could deliver sediment or debris in a manner that would threaten public safety".

WAC 222-16-050, subsection (1)(d)(i), identifies five sets of potentially unstable slope and landform types, referred to as Rule-Identified Landforms (RIL):

A. Inner gorges, convergent headwalls, or bedrock hollows with slopes steeper than thirty-five degrees (seventy percent);
B. Toes of deep-seated landslides, with slopes steeper than thirty-three degrees (sixty-five percent);
C. Groundwater recharge areas for glacial deep-seated landslides;
D. Outer edges of meander bends along valley walls or high terraces of an unconfined meandering stream; or
E. Any areas containing features indicating the presence of potential slope instability which cumulatively indicate the presence of unstable slopes.

These landforms were identified as potentially unstable based on extensive landslide mapping for watershed analyses. RIL Type B addresses the potential for triggering smaller landslides at the steepened toes of deep-seated landslides, Types D and E may potentially involve deep-seated-landslide features, and Type C explicitly identifies the groundwater recharge area to a glacial-deep-seated landslide as a feature of concern. Groundwater recharge areas to *non*-glacial-deep-seated landslides are not included in the set of rule-identified landforms, but are included in the

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[1] A qualified expert is a person licensed under chapter 18.220 of the Revised Code of Washington as either an engineering geologist or as a hydrogeologist

questions posed by Policy, CMER, and UPSAG for this review of the non-glacial deep-seated landslide literature.

Section 16 of the Forest Practices Board Manual, "Guidelines for Evaluating Potentially Unstable Slopes and Landforms" [2], provides descriptions of different landslide types, criteria for identifying unstable slopes and landforms, and suggestions for analysis methods to assess the likelihood that proposed forest practices will affect landslide movement.

### 1.3 Deep-seated landslides

Although gravity moves material downslope through a variety of processes, our focus is on landslides that involve movement of material above a well-defined failure zone. Landslides are a primary mechanism for movement of slope material downslope. Washington has an abundance of slopes, offering many opportunities for humans to interact with nature's hillslope leveling activities. In general, we seek to avoid those interactions. To do so, we need to know where landslides will occur, how often they will occur, and how our activities will alter those occurrences.

Geologists distinguish between shallow and deep-seated landslides based on the depth below the ground surface to the failure zone. Shallow landslides involve depths down to a couple meters; deep-seated landslides extend from several to tens of meters. The distinction from shallow to deep can be transitional, but it is still useful, because shallow and deep-seated landslides tend to occur in different landscape positions, to involve different mechanisms and rates of movement, to exhibit different responses to human activities, and to pose different types of hazards. Here we focus on deep-seated landslides.

Deep-seated landslides occur over a large range of sizes, from hundreds to millions of cubic meters. Some deep-seated landslides are catastrophic, involving sudden failure and rapid avalanching or flowing of debris downslope. Catastrophic failure of large landslides can pose significant risks to public infrastructure and human lives, such as the March 22, 2014 SR530 (Oso) landslide (Wartman et al., 2016) and the May 2, 2014 Abe Barek landslide in Afghanistan (Zhang et al., 2015). But for many, perhaps most, deep-seated landslides, movement occurs gradually and intermittently over variable time periods: several years to perhaps thousands of years. This gradual and intermittent movement forms characteristic landforms that can persist as recognizable features on the landscape for long periods, from decades for small landslides to thousands of years for large landslides. The occurrence of landslide events over time can thus cause deep-seated landslide features to eventually dominate a landscape. Pierson et al. (2016), for example, found that deep-seated landslide features covered 64% of the surface mapped for an area including the Cascade Landslide Complex along the Columbia River in southwest Washington.

The ubiquitous presence of deep-seated landslides poses a problem for hazard assessment. Most mapped deep-seated landslides show no evidence of recent activity. With their mapping along the Columbia River, Pierson et al. (2016) found that only 12 of the 215 mapped landslides

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[2] http://file.dnr.wa.gov/publications/fp_board_manual_section16.pdf

showed evidence of movement in the past 20 years. Once formed, however, deep-seated landslides remain potentially vulnerable to continued movement, even if they have been inactive for long periods. Reactivation of long inactive deep-seated landslides poses hazards to communities throughout the world (e.g., Bertolini, 2010; Christenson and Ashland, 2006; Massey et al., 2013). Of the thousands of inactive deep-seated landslides that might exist in a region, there may be no reliable indicators of which, if any, are on the verge of becoming active or will respond to human activities.

Washington's forest-practice rules distinguish between glacial and non-glacial deep-seated landslides: those occurring in glacial deposits and those occurring in bedrock or non-glacial deposits. The northern portion of the state was buried under an ice sheet 12,000 years or so ago, leaving river valleys and coast lines filled with the legacy of ice-dammed lakes and outwash streams. Subsequent river incision, channel lateral migration, and wave erosion of these deposits created conditions exceptionally suited to formation of deep-seated landslides. Recognition that glacial environments juxtapose permeable outwash deposits with impermeable lake deposits, which can make glacial-deep-seated landslides acutely sensitive to changes in water balance, motivated the special scrutiny that groundwater recharge zones to glacial deep-seated landslides receive in the forest practice rules.

Deep-seated landslides also occur in bedrock, in soils formed of weathered bedrock, and in volcanic and alluvial deposits, all of which are also found in abundance in our state. These landslides present an even broader array of material properties, landslide types, geologic influences, and potential groundwater interactions than found for glacial deep-seated landslides. They may also pose substantial hazards: many of the examples listed in the "Significant Deep-Seated Landslides in Washington State – 1984 to 2014" posted by the Department of Natural Resources (http://file.dnr.wa.gov/publications/ger_list_large_landslides.pdf) occurred in areas never glaciated.

### 1.4 The literature

A vast literature addresses deep-seated landslide processes, hazard assessment, case studies, and mitigation. A tiny literature addresses influences of forest practices on deep-seated landslides. To cover the issues of concern, we therefore examined a range of topics, expanding from deep-seated landslide studies to include use of stable isotopes to estimate sources of groundwater to the role of bedrock fractures in storm runoff.

We used several strategies to find resources, including keyword searches in Google Scholar and ResearchGate, systematic review of articles in relevant publications (e.g., Geomorphology, Landslides, Water Resources Research, Engineering Geology, Environmental and Engineering Geology), and technical reports provided by the Washington Department of Transportation, and citations provided by the advisory team. The most productive strategies, however, proved to be backward and forward snowballing: using articles cited in papers reviewed, and then using Google Scholar or the publisher's websites to find newer articles that cited those. This is a time-consuming approach, and prone to meandering searches that sometimes lead nowhere, but it also led us to useful articles we could not have found any other way.

There are commonalities between glacial and non-glacial deep-seated landslides, so many of the citations included in the literature review for glacial deep-seated landslides are relevant here. For this review, we focused on studies addressing landslide and groundwater processes in bedrock. The citation list includes over 150 references in addition to the approximately 140 cited in the glacial deep-seated landslide literature review, over 20 of which were published since we finished the glacial deep-seated landslide review.

### 1.5 This document

UPSAG provided a set of issues and a list of questions to address in the synthesis report. CMER and Policy had concerns they wanted addressed as well. We have placed these issues and questions upfront in the next three sections: Section 2, Questions and Answers; Section 3, Knowledge Gaps; and Section 4, Recommendations. These sections are kept intentionally brief. Detailed descriptions of the processes discussed and of the citations where pertinent information can be found are provided in the background material (Sections 5 through 9).

## 2 Questions and answers

Questions and issues posed by UPSAG, CMER, and Policy are addressed here. We have grouped these into similar categories, keeping all the original text of the questions, including redundancies.

### 2.1 What are the triggers for creation and reactivation of non-glacial deep-seated landslides?

Before discussing triggers, it is informative to recognize the two primary modes of movement that characterize downslope displacement of rock and soil: creep and shear.

1. *Creep*. Deformation occurs from the surface downward continuously to a certain depth. Total downslope displacement is greatest at the surface and decreases with depth.
2. *Shear*. Sliding of a mass of material across a slip surface or rupture zone at a specific depth. Deformation occurs within a limited thickness through that zone and displacement above that zone is nearly constant from the surface down to that zone.

All slopes experience some degree of creep. In granular materials, creep is accommodated by sliding across grains. In cohesive materials, such as rock, clay-rich soils, and fine-grained materials compressed by glacial ice (glacial-lacustrine sediments, till), deformation is accommodated by growth, linkage, and slip across a network of cracks (Carey and Petley, 2014). In cohesive rock and soils, creep is a precursor to slope failure (Froude, 2011), and failure occurs if the network of cracks coalesces into a shear zone. Deep-seated landslides are formed by displacement of material across that shear zone. Deep-seated-landslide behavior is thus greatly influenced by the geometry and material properties of the shear zone. Knowledge of the mechanisms that influence shear-zone formation can aid in anticipating subsequent deep-seated landslide behavior.

### *2.1.1 Triggers for creation*

*Progressive failure.* (Section 6.3) Landslides are generally triggered by specific events: a large storm or series of storms, an earthquake, removal of buttressing material at the base of the slope. That triggering event, however, is the culmination of a long sequence of events that progressively weakened zones within the slope. Slope failure occurs when these weakened zones merge to form a continuous shear zone. This process is a consequence of brittle failure (Section 6.2), characteristic of rock and soils at depths of less than about 70 meters. At these depths, crack growth and weathering occur gradually at stress levels well below those required to break intact rock or soil. Stress concentrations caused by topography and geologic structure tend to focus this crack growth into narrow zones that become progressively weaker over time. As portions of a slope weaken, the rate of crack growth accelerates and, at some point, cracks rapidly coalesce to form a shear zone and the slope fails.

Crack growth and progressive slope failure are initiated by changes in stress conditions within a slope. Such changes may be associated with alpine valley glacial retreat, river incision, elevated pore pressure, or excavation for a road cut. Once initiated, the process of progressive failure in soil may occur within days; in rock, it may take thousands of years. Over this time, a slope may undergo periodic stress fluctuations caused by the seasonal rise and fall of groundwater levels, shaking from earthquakes, freeze/thaw cycles, tree root growth in rock fractures, and changing river levels or bank erosion at its toe. A slope may endure such fluctuations for many years, but as it weakens over time, one final seemingly minor change can trigger complete failure and lead to the formation of a deep-seated landslide.

### *2.1.2 Triggers for reactivation*

Topographic features indicative of past deep-seated landslide movement are created by sliding of overlying material across the shear zone. The geometry and material properties of the shear zone thus exert primary controls on landslide behavior and on the response of a landslide to potential triggers for reactivation. Coalescence of a network of cracks to form a shear zone and subsequent movement across the shear zone have broken any cementing or cohesive connections between rock and soil particles within the shear zone, so it is weaker than material above and below. This mechanical breakage also tends to make material within the shear zone finer grained and more prone to chemical weathering (to clay minerals), so that the shear zone tends to have lower permeability than material above and below, potentially causing the shear zone to act as an aquitard that can hold groundwater within a landslide body (e.g., Baum and Reid, 2000).

In discussions of landslide reactivation, it is generally assumed that these properties of the shear zone – lack of cohesion, low permeability – persist when a landslide is inactive. Chemical alterations that occur over time may affect shear-zone properties (Bromhead, 2004). Landslide studies tend to focus on active landslides, so evolution of shear-zone properties during long periods of landslide inactivity is not well explored. Chen and Liu (2014) find that a slip-zone soil from an ancient landslide in China exhibits no cohesion. Re-establishment of some cohesive strength during periods of landslide inactivity is observed in certain clay soils, but laboratory experiments indicate that such strength regain is limited to shallow depths more pertinent to shallow landsliding than to deep seated (Stark and Hussain, 2010). Thus, we assume in the following discussions that the shear-zone properties that control landslide reactivation persist over time when landslides are inactive. This assumption provides a conceptual basis for identifying the processes that might control reactivation.

Development of a shear zone creates a thin, weak layer (from several millimeters to several meters thick) across which large displacements can occur. Although the shear zone is weaker than adjacent material, frictional forces continue to resist movement across the shear zone, and that frictional strength remains nearly the same over time, regardless of how much movement has occurred. Whenever forces acting to move the overlying material – the landslide body – exceed the shear strength of the shear zone, the landslide moves. Triggers for reactivation thus involve factors that reduce resistance of the shear zone: pore-water pressures, which vary in response to rainfall; or changes in geometry of the landslide that alter the balance of those forces.

*Pore pressure*. (Section 9.1.1) Although the material properties that determine frictional strength of the shear zone remain relatively constant over time, pressure exerted by water that fills pore spaces within the shear zone will reduce shear resistance. These pore pressures are proportional to the depth of groundwater[3] within the landslide body. As groundwater depth increases, shear resistance across the shear zone decreases. Groundwater levels vary with seasonal variations in precipitation. Landslide movement initiates when groundwater reaches a level that generates pore pressures that reduce shear strength of the shear zone to a value less than the gravitational (or seismic) forces acting to move material downslope.

Pore pressures within the shear zone may also be influenced by pressure exerted by groundwater impinging on the shear zone from below. High pressures within a confined aquifer intercepted by the shear zone may also trigger landslide movement or make a landslide more sensitive to other factors.

*Rainfall thresholds*. (Section 9.2.1) Groundwater levels within a landslide vary in response to a time series of precipitation events. The magnitude of groundwater variations depends on the magnitude and timing of these events and on the flow paths for water to, within, and out of the landslide. If pathways exist for rapid infiltration and flow of groundwater, such as fissures at the ground surface, sub-surface soil pipes, and fractured bedrock below, groundwater levels may rise and fall in tandem with rainfall events. If fracture porosity is low (but connectivity is high), the rise in groundwater levels may be large, because it takes a relatively small volume of water to fill available pore space. If permeability of landslide debris is low, and if there is a vadose (unsaturated) zone that water must traverse, groundwater response may be slow and muted, but will integrate water inputs over longer time periods, potentially spanning years. Hence, there could be a considerable lag between periods of high rainfall and reactivation of a landslide.

Groundwater response is thus a function of the time series of precipitation events. By comparing time series of precipitation to time series of initiation of landslide movement, researchers have in some cases identified rainfall thresholds for initiation of movement. These thresholds can be

[3] We use the term "groundwater" to refer to water in any saturated zone.

complex functions of antecedent rainfall intensity and duration, which may include estimates of evapotranspiration, and vary from landslide to landslide.

The rainfall threshold for a landslide can also change over time as growth or infilling of cracks and pore spaces and surface-drainage development within a landslide alter its groundwater response.

*Undrained loading*. (Section 9.1.2) High pore pressures can also be induced by compression of material composing the landslide body. Such compression can occur when landslides in adjacent material, such as subsequent failure of the headscarp, deposit additional debris onto the body of a landslide. Compression and shearing of the body can reduce pore-space volume, causing the debris to contract; if pore spaces are filled with water, pore pressure goes up and shear resistance goes down. Water will be squeezed out of the compressed material, but if the material is relatively impermeable, it may take hours, days, even months for the excess pressure to dissipate. This process, referred to as undrained loading, can trigger and maintain movement of landslide debris and has been attributed to reactivation of earthflows in Italy (Bertolini, 2010) and British Columbia (Geertsema et al., 2006). In some cases, the compressive pulse induced by rapid undrained loading causes debris to liquefy and mobilize into a debris flow (Geertsema and Schwab, 2006). Undrained loading can mobilize landslide debris over gentle slope gradients that would be stable otherwise.

*Changes in geometry*. (Section 9.1.3) Events that modify landslide geometry alter the balance of forces within the landslide. Examples include channel incision into the body or at the toe, bank erosion at the toe (Keck, 2017, for example, attributes activation of a deep-seated landslide on the Olympic Peninsula to bank erosion by debris flows from upslope), or excavations for road construction (e.g., Stark et al., 2005a). If these changes act to reduce the forces that are resisting movement, such as any buttressing at the toe, they may trigger motion directly or reduce the pore-pressure and rainfall thresholds for triggering movement.Precipitation, runoff, and groundwater recharge

The abundance and movement of subsurface water plays a primary role in driving downslope movement and failure (Coates, 1990). Deep-seated landslide motion is largely controlled by pore pressures at a basal and lateral shear zone across which landslide movement occurs. These pore pressures are proportional to the depth of groundwater stored within the landslide, so information regarding the sources of water to a landslide is key to success in anticipating landslide behavior (Section 8).

### *2.1.3 Are groundwater recharge areas associated with non-glacial deep-seated landslides?*

Yes, some landslides are affected by recharge from beyond the landslide boundary; see Section 8. Water from precipitation and snowmelt has been observed to flow to deep-seated landslides via four potential pathways:

1. *Direct infiltration* from rain and snowmelt on the body of the landslide.
2. *Surface runoff* from areas upslope (e.g., Baum and Reid, 1995; Proffer, 1992).

3. *Subsurface runoff* from areas upslope via groundwater flow through shallow transient aquifers formed in soil and fractured bedrock perched above underlying less permeable substrates (e.g., Binet et al., 2007b; Vallet et al., 2015a).
4. *Groundwater* flow through perennial aquifers (e.g., Cervi et al., 2012).

Pathways 2 through 4 involve contributing areas outside of the landslide boundary and can account for a substantial portion of the water inflow. The proportion of inflowing water from each source may vary substantially from site to site, depending on surface topography and subsurface conditions, such as the degree to which shallow, transient aquifers formed by stormflow are hydrologically connected to a landslide. Proffer (1992) found that subsurface runoff provided 55% of the water to a large landslide in California; Cervi et al. (2012) found that groundwater accounted for 64% of the water inflow to a landslide in Italy. We have found no studies that attempt to determine the proportion of water from each of these sources for landslides in the Pacific Northwest.

### *2.1.4 How do groundwater recharge areas affect deep-seated landslides?*

Landslide movement can be initiated and accelerated by the increasing pore pressures associated with rising groundwater levels within a landslide body. The effect of inflowing water originating from outside a landslide boundary depends on the degree to which that water increases groundwater levels within the landslide. Groundwater levels within a landslide body reflect a potentially complex interaction between the time series of precipitation and snow-melt events, the inflow rates and transit times for water inflows from each of the four pathways listed above, and the rate of water outflow from the landslide

*Inflow*. The total volume of inflow to a landslide depends on the volumes from each of the four flow pathways listed above. Those volumes are determined by the size of each contributing area (or recharge area for groundwater), the amount of precipitation falling on that area, and the proportion of water that flows through that pathway to a landslide.

*Outflow*. The response of groundwater within a landslide to inflowing water from each flow path depends on the rate and timing of water inflow relative to the rate of water outflow. Water exits a landslide by: seepage through the basal shear zone; outflows of groundwater to the surface at streams, springs, and seeps; and evapotranspiration back to the atmosphere. If water drains from the landslide quickly, then groundwater levels within the landslide may rise and fall in concert with fluctuating water inflows. If water drains from the landslide slowly, groundwater volume within the landslide will accumulate inflows over some potentially long (years) period of time (e.g., Baum and Reid, 2000). Development of incised channels over the body of a landslide may facilitate drainage of water from the landslide by intersecting the water table, which also limits the local height of the water table to the elevation of the base of the channel.

These factors vary from landslide to landslide, but commonalities in climate, geology, and topography may create similar patterns of groundwater response for deep-seated landslides across a region. However, effects on landslide stability of source areas for water originating from outside a landslide boundary may differ for landslides in different regions or in different geologic or topographic settings. We have found no studies that explore these differences.

### 2.1.5 How do forest practices affect these groundwater recharge areas?

*Forest Roads*. Road surfaces generate surface runoff and road cut banks can intercept shallow subsurface flow (Wemple and Jones, 2003), altering the distribution and timing of water flows between the four flow pathways. Consequences for each flow path depends primarily on where water from the road is discharged back to the forest floor: roads may act to divert water away from or onto areas contributing runoff and groundwater recharge to a deep-seated landslide. Recent work on the La Conchita landslide in California, for example, suggests that surface drainage changes due to building a dirt road contributed to the 2005 catastrophic landslide reactivation (Pradel, 2014).

*Timber Harvest*. Timber harvest reduces evapotranspiration[4], which increases water available for infiltration, runoff, and recharge. Studies of water-vapor fluxes (e.g., Brümmer et al., 2012) show that evapotranspiration from forests varies from near 90% of total precipitation in arid regions to about 20% in humid regions, and that loss of forest cover can substantially reduce evapotranspiration. Jassal et al. (2009), for example, measured a 30% reduction in evapotranspiration between older (> 50yrs, ~400mm/yr) and younger (< 10yrs, ~250mm/yr) Douglas Fir stands on Vancouver Island. Paired watershed studies indicate that harvest-related reductions in evapotranspiration translate to equivalent increases in water yield from runoff and recharge (e.g., Hubbart et al., 2007; Keppeler and Ziemer, 1990; Rothacher, 1970; Stednick, 1996). Harvest-related increases in water yield decrease over time as forests regrow, but may persist for up to five decades (Burt et al., 2015).

Increases in water yield for contributing areas (for both runoff and groundwater recharge) associated with timber harvest can be estimated from water-balance models. Examples include relatively simple spreadsheet- and GIS-based calculations (Harbor, 1994; Westenbroek et al., 2010) to more complex, but more complete, spatially distributed transient models, such as the Distributed Hydrology Soil-Vegetation Model[5] (e.g., Du et al., 2016).

### 2.1.6 Are there methodologies that have been used to delineate groundwater recharge areas to non-glacial deep-seated landslides?

Given that a deep-seated landslide responds to the inputs from all water-flow pathways, it is useful to expand this question to also include contributing areas that generate surface and subsurface runoff to the landslide (in addition to groundwater recharge). The primary methods include use of topographic divides and field mapping, isotopic tracers, and numerical modeling.

- *Topographic divides and field mapping*. (Section 8.5.1) Water flows from higher to lower elevations, so all runoff and groundwater inputs to a landslide originate from water infiltrated upslope.

[4] See the appendix for the glacial deep-seated landslide literature review synthesis.

[5] http://www.hydro.washington.edu/Lettenmaier/Models/DHSVM/

Surface runoff originates from areas within the topographically defined surface drainage to a landslide. Surface runoff includes channels that drain to the landslide that may receive subsurface runoff and groundwater-derived baseflow from beyond topographic divides.

Subsurface runoff occurs within near-surface high-permeability zones: soils overlying low-permeability substrates and highly fractured bedrock. Field observations can, to some extent, determine if such zones are present and delineate their extent. Extensional features, such as tension cracks, uphill-facing scarps, and closed depressions, may indicate locations of rapid water infiltration to near-surface fractured bedrock.

Groundwater recharge zones are difficult to identify precisely, because the groundwater divide may not correspond with the surface drainage divide. However, groundwater source areas may be roughly identified by tracing steepest-descent paths from upslope points, extending from ridge tops to perennial stream channels (Vallet et al., 2015a).

- *Isotopic tracers*. (Section 8.5.2) Water contains stable isotopes of oxygen and hydrogen. The abundance of these isotopes in rainwater varies systematically with elevation. Hence, the abundance of these isotopes found in subsurface water can be used to infer the elevation where that water first infiltrated into the ground. This method does not explicitly delineate the recharge area, but it does indicate if recharge occurs at elevations upslope of a landslide boundary.
- *Numerical modeling*. (Section 8.5.4) Groundwater models are used to estimate flow paths and delineate recharge areas. Use of these models requires specifying the three-dimensional distribution of subsurface permeability, which may be inferred by extrapolating surface information, but accurate determination of which requires substantial investment in subsurface exploration. Such models can provide hypotheses about the extent of the recharge area that can then be tested with field observations. Partial validation of model results requires subsurface measurements of groundwater levels.

## 2.2 Material properties

### *2.2.1 How do the properties of geologic materials affect non-glacial deep-seated landslide style of movement and runout distance?*

Although deep-seated landslides occur within a large range of material types and geologic settings, almost all exhibit certain general characteristics:

- Deep-seated landslide movement occurs primarily by sliding over a well-defined shear zone (e.g., Bromhead, 2004). Development of a shear zone breaks any cementing matrix between rock or soil particles, with subsequent loss of cohesion. Mechanical breakage of particles during shear displacements can cause material within the shear zone to have a finer grain size, and thus lower permeability, than material above and below the zone.
- The shear zone is weak relative to adjacent material. It has little or no cohesion and resists movement through friction. Residual strength varies with rock and soil types, but loss of cohesion causes the range to be small relative to the range of peak strength of those

materials. Residual friction angles for shear-zone soils are typically in the range of 10 to 35 degrees (Chen and Liu, 2014; Stark et al., 2005b; Stark and Hussain, 2013).

- Strength of the shear zone is inversely proportional to pore pressures within the shear zone. Increases in pore pressure cause a reduction in strength.
- The shear zone tends to have low permeability relative to adjacent material. The body of a landslide bounded by the shear zone may thus contain a groundwater system isolated to some degree from adjacent areas. Inputs to this system occur by direct infiltration of precipitation into the body, surface and subsurface runoff into the body from areas upslope, and groundwater seepage upward through the shear zone from below. Outflows from this system occur by downward seepage through the shear zone and surface drainage where the water table intersects the ground surface.

These factors create similarities across deep-seated landslide types. Variations in material properties generate differences in landslide behavior.

- Permeability of material composing the landslide body influences how groundwater responds within the landslide. Competent materials can maintain open fractures that allow rapid transit of water vertically and laterally through a landslide body. Pore-pressure responses to precipitation may be rapid. Clays developed from weathering processes clog fractures and other pathways for water, so as landslide deposits age, permeability tends to decrease, and pore-pressure responses tend to be delayed and dispersed, integrating seasonal or longer trends in precipitation. This evolution from higher to lower permeability occurs more rapidly in weaker, more readily weathered rock types.
- As bedrock weathers in place to form a residual soil, the upper soil layer (to the A Horizon) may become permeable with a deeper, lower-permeability zone developing (to the B Horizon) where clay minerals accumulate (Lambe, 1996). Perched groundwater can form above the lower-permeability zone (above the B Horizon). In deep residual soils, formation of a perched shallow aquifer can initiate deep-seated landslides. These low-permeability layers can also act as aquitards for groundwater flowing upward from below.
- Low-permeability material is more subject to reactivation by undrained loading (Section 9.1.2), because of the long time for pore-pressure increases caused by compression of the material to dissipate. Hence, reactivation by undrained loading is most likely in clay-rich and fine-grained debris. In fine-grained materials, undrained loading can also trigger liquefaction and high mobility (long-runout distances).

### *2.2.2 Potential categories of various non-glacial materials that react differently to forest practices than other materials, such as depth, geologic map unit, stratigraphy, slope, precipitation zone, permeability, proximity and juxtaposition to stream channels*

We found no studies that address differences in the reaction of deep-seated landslides to forest practices and other factors. The only published study we found that directly examines the effects of forest practices on deep-seated landslide behavior is that of Swanston et al. (1988), which documented accelerated movement of an earthflow in weathered sedimentary rocks following clear-cut harvest in western Oregon. One study is insufficient for drawing broad inferences, but forest practices affect the same physical processes as natural increases in water flows and changes in slope geometry, so observed deep-seated landslide responses to natural events can show which materials and what landslide characteristics might influence landslide reactions to forest practices.

Analyses of landslide inventories show that the number and relative area of deep-seated landslides varies with material properties and slope characteristics. For an inventory in the southern Cascades of Washington, Dragovich et al. (1993a) found the highest deep-seated landslide densities in intrusive igneous rock types and the lowest in interbedded volcanic-sedimentary rocks. They found higher deep-seated landslide densities where bedding surfaces intercepted the ground surface at an angle, rather than on dip (bedding planes parallel to surface) and scarp (bedding planes perpendicular to surface) slopes. For the urban corridor in Cowlitz county, (Wegmann, 2006) found the highest landslide densities in volcanic tuffs of the Toutle, Troutdale, and Cowlitz formations. Gerstel and Badger (2002) cite the Lincoln Creek Formation, a fine-grained sedimentary rock in southwest Washington, as a “notorious bad actor”. These studies show that landslide susceptibility varies with rock type, stratigraphic sequence, and bedding orientations relative to surface topography, and that these relationships vary regionally. We can infer that combinations of these factors that are more prone to landslide formation will also be more sensitive to forest practices, but this inference has not been tested for deep-seated landslides through empirical studies.

### 2.3 What are the characteristics of large landslides that may predispose them to composite failure.

Deep-seated landslides commonly take place in temporal and spatial sequences. Specific terminology is used to describe different scenarios. Cruden and Varnes (1996) define *composite* landslides as involving different types of movement in different parts of a landslide; *complex* landslides as involving different types of movement in sequence; and a *multiple* landslide as involving repeated movements of the same type, which may share a common shear zone. Cronin (1992) defines a *compound* landslide as “a landslide that has calved or become segmented into smaller, secondary landslides”. Chapter 16 of the Board Manual states that “Some compound deep-seated landslides found in glaciated and non-glaciated terrain have the potential to become highly mobile failures” and cites the 2014 SR 530 (Oso) landslide as an example.

In recognition of the potential for composite landsliding, hazard assessments should examine the potential of future landslide occurrence and the potential consequences of such occurrence. Methods to improve prediction of reactivation of existing landslides are discussed below in Section 2.7.2. These methods should be applied recognizing potential for composite, complex, and compound types of behavior. For example, the spatial density of shallow landslides tends to be greater within deep-seated landslide features (including dormant and relict landslides) than in adjacent areas (e.g., Dragovich et al., 1993b; Wegmann, 2006). Perhaps the density of smaller, deep-seated landslides is also greater within larger deep-seated landslides than in adjacent areas.

### 2.4 What are the characteristics of large landslides that may predispose them to long rapid runout?

Long, rapid runout occurs when landslide debris attains high velocity (on the order of 5m/sec or more, Hungr, 2007). A rock or soil slope can lose strength almost instantaneously when it initially fails. Subsequent movement of landslide debris is constrained by shear resistance across the resulting failure surface (shear zone) and capacity of the landslide debris to resist deformation as it moves. If the failure surface is steep, or if material above the failure surface loses strength as it deforms (loose soils, fracturing rock), upon failure, landslide debris may fall nearly unconstrained downslope. Many long runout landslides are thus associated with initial failures.

On reactivated landslides, high velocities and long runout can occur with loss of strength (liquefaction) caused by suddenly elevated pore pressures within the shear zone. Such elevated pore pressures may be caused by undrained loading (Section 9.1.2), such as when a rock slide deposits onto saturated, fine-grained materials (as described for the Muskwa landslide in British Columbia by Geertsema et al., 2006; Geertsema and Schwab, 2006), or when materials contract during shear deformation (Iverson, 2005; Iverson et al., 2000). Contraction causes a reduction in pore-space volume with a consequent increase in pore pressure and reduction in shear strength. Reduced shear strength results in greater displacement and more contraction. This feedback can result in unconstrained acceleration of landslide debris. Contraction during shearing deformation is observed in loose soils.

Geertsema et al. (2006) document recent (last 40 years), large, long-runout landslides in British Columbia and document the types of landslides and geomorphic settings involved. For non-glacial landslides, these involve rock falls and rock slides on cirque walls, sedimentary dip slopes, and mountain slopes undergoing deep-seated gravitational deformation (see Section 6.1). Some of these landslides evolved into debris flows or debris avalanches, and some initiated movement on downslope earthflows. Most of these bedrock landslides were interpreted as initial failures, and not re-activations of existing landslides. Geertsema and Schwab (2006) discuss these landslide types in the context of assessing landslide hazards for forest practices.

Long-runout landslides described in the literature often involve a sequence of events, such as where a rockslide evolves into a flow or avalanche, entraining downslope material and growing in size, or deposits onto existing landslide debris and triggers movement through undrained loading (Geertsema et al., 2006; Hungr, 2007). In most cases, the initiating event involved an apparently new failure. Deep-seated landslide deposits can also dam streams; failure of the landslide dam can then generate a dam-break flood or debris flow, as attributed to deposits along Jones Creek at Acme, Washington.[6]

#### *2.4.1 What methods might improve prediction?*

Landslide runout is observed to increase with increasing landslide relief or fall height (elevation difference from head to toe), landslide volume, and steepness and topography of the depositional

[6] http://www.whatcomcounty.us/DocumentCenter/View/11552

zone (Hunter and Fell, 2003; Iverson et al., 1998; Legros, 2002). These empirical relationships may vary for landslides in different materials and geomorphic settings (Perkins et al., 2016). Local calibrations can be made from regional mapping of landslide height-runout length relationships (Hattanji and Moriwaki, 2009; Perkins et al., 2016), so that measures of slope heights can be used to predict potential runout lengths.

## 2.5 What are the best tools to assess runout potential for deep-seated landslides?

For regional hazard assessments and initial screening, empirical methods regionally calibrated to mapped deposit extents provide the most applicable tool. Hattanji and Moriwaki (2009) illustrate this approach for areas in Japan. Perkins et al. (2016) describe use of LiDAR for mapping runout extent for landslides in northwest Washington, and point out that length-to-height ratios differ for different materials. Mapping of landslide scar and deposit geometries from high-resolution digital elevation models (DEMs) and field surveys could be used to calibrate empirical models for representative rock types and glacial deposits across Washington. Scatter within the data can be used to infer exceedance probabilities for runout extent (McDougall, 2017). Resulting statistical models can be translated to maps of probability of runout extent. Such empirical approaches assume that past runout extent provides an accurate indicator for future runout potential. For potential flow-type landslides, if the initial volume can be estimated, the lahar-runout model developed by Iverson et al. (1998) may be used to account for downslope variations in valley topography (Griswold and Iverson, 2008; Schilling, 1998).

Physical models for landslide runout, such as those by Hungr (1995) or Iverson and George (2016), can also be applied to deep-seated landslides. Physical models are, however, extremely dependent on the specified material properties and initial and boundary conditions, which are poorly known for most cases. For hazard assessment, regionally calibrated empirical models provide greater applicability.

## 2.6 Sensitivity to forest practices

### *2.6.1 What are the impacts of forest-practice activity on non-glacial deep-seated landslide movement?*

We found only one study that sought to examine effects of forest practices on deep-seated landslides: Swanston et al. (1988) monitored sites on and adjacent to an active earthflow in western Oregon before and after clear-cut harvesting over the earthflow. They found acceleration of movement over a portion of the earth flow following timber harvest, from about 3.4mm/yr to 20.5mm/yr. This is a six-fold increase[7]; however, these rates are nearly imperceptible, corresponding to the slowest velocity class (extremely slow) in a widely used classification

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[7] The authors report total surface displacement at borehole B-2 of 68mm over the 10-year monitoring period (October 1975 to April 1984), 41mm of which occurred during a period of accelerated movement over two years (winter 1977 to winter 1979). Average rate of movement during the non-accelerated period was (68-41)mm/8yr = 3.375mm/yr; average rate during the period of accelerated movement was 41mm/2yr = 20.5mm/yr. This is a six-fold increase. In their conclusions, the authors state a 14% increase, but it is not clear how they came up with that number.

(Hungr et al., 2014). The accelerated rate persisted for a little over two years before returning to pre-logging levels.

We found no studies that investigated a cause and effect relationship between forest practices and the reactivation of inactive deep-seated landslides in the Pacific Northwest. This is not necessarily an indication that such relationships do not exist; rather, it may indicate the difficulty in discerning the effects of forest practices amid the noise of natural variability. More than 40 years ago, Swanston and Swanson (1976) inferred that timber-harvest reductions in evapotranspiration, as manifest through increased water yield, could have substantial impacts on deep-seated landslide movement. Given the paucity of studies, we must continue to rely on inference.

Deep-seated landslides are reactivated when pore pressures at the shear zone exceed some threshold (e.g., Iverson and Major, 1987), the rate of landslide movement may increase with increasing pore pressure (e.g., Hong et al., 2005), and cessation of movement occurs when pore pressures drop below some threshold value. Pore-pressure fluctuations are driven by temporal variability in supply of water to a landslide. Water is supplied through four pathways: direct infiltration of precipitation into the landslide, surface runoff from upslope, subsurface runoff from upslope, and groundwater seepage from below the shear zone.

Removal of tree canopy by forest harvest reduces evapotranspiration with a consequent increase in infiltration, runoff, and groundwater recharge. This increase ranges from 10% to 15% of total precipitation in the Pacific Northwest (see review in Miller, 2016); it can persist for a decade or more, but decreases over time as forests regrow. By increasing water supply and associated pore pressures, timber harvest may trigger reactivation of a deep-seated landslide, can increase the rate of movement of a deep-seated landslide, and may increase the total time that a landslide remains active.

The potential that forest practices will have any of these impacts depends on:

- the increase in water supply to a landslide caused by forest practices,
- the sequence of precipitation events over the time that water supply is increased,
- the influence of changes in water supply to groundwater levels in the landslide, and
- the sensitivity of the affected landslide to changes in pore pressure.

Capabilities to evaluate each of these exist, but we have found no studies that examine all of them in the context of deep-seated landslides and forest practices in the Pacific Northwest.

### *2.6.2 Does harvesting of the recharge area of a non-glacial deep-seated landslide promote its instability?*

Potentially yes. Harvesting reduces evapotranspiration, which can increase runoff and recharge. If harvest occurs in areas upslope of the landslide boundary that provide water to the landslide by surface runoff, subsurface runoff, or groundwater flow, the harvest can increase the amount of water that flows to the landslide, which can increase groundwater levels within the landslide, increase pore pressures, and reduce resistance to shear forces across the shear zone; that is, the harvest can promote instability. This reduction in stability could persist until the forest has regrown sufficiently that evapotranspiration is recovered to pre-harvest levels. Eddy covariance measurements indicate that pre-harvest levels of evapotranspiration may be achieved in about 15 years for a site on Vancouver Island (Jassal et al., 2009).

However, paired-catchment studies of water yield suggest that effects of harvest on basin hydrology can persist somewhat longer. Jones and Post (2004) present a long-term (multi-decade) analysis of 14 paired-catchment studies and find that increases in annual water yield after clear-cut harvesting of conifer forests persist for 30 to 40 (or more) years. Increases in water yield following harvest were larger in snow-dominated basins, and also depended on the age of the original stand. Burt et al. (2015) recently updated analysis of paired-catchment data from the H.J. Andrews experimental forest in western Oregon and also found that effects of harvest persisted for over four decades, in that a 40-year-old stand loses less water by evapotranspiration and more in runoff than an old-growth stand. Du et al. (2016) recently used hydrologic simulations to evaluate effects of different harvest strategies on water yield for the Mica Creek Experimental Watershed in snow-dominated northern Idaho. They also find that complete recovery after clear-cut harvesting to a stable baseline of hydrologic yield may take 45 to 50 years. These studies all find that the rate of hydrologic recovery can vary with site-specific factors, such as the rate of stand regrowth and the proportion of precipitation that falls as snow, but they also show that complete hydrologic recovery may take decades. The implications for deep-seated landsliding have not been explored.

Whether the reduction in stability triggers landslide movement depends on the time series of precipitation (and snow-melt) events over the period of recovery, on the change in water yield associated with harvest, on the sensitivity of the landslide to pore pressure increases, and on factors that may increase that sensitivity, such as erosion of the landslide toe. Increased pore pressures will also render the landslide more sensitive to other events over the period of recovery, such as seismic shaking.

### *2.6.3 Are there differences in response to forest practices versus natural influences?*

This question can be addressed over a range of spatial and temporal scales. The response of an individual deep-seated landslide to forest practices over its extent and within its source area for runoff and groundwater recharge forms one end of this range. This defines the spatial scale at which rule-identified landforms and forest-practice applications are evaluated. At this scale, changes in groundwater levels associated with timber harvest are of similar magnitude as changes associated with natural variability in precipitation. Harvest-related increases are, however, overprinted on that natural variability, which may result in groundwater levels exceeding those that would occur naturally if harvest occurs prior to an exceptionally wet period (see section 9.2.2).

Likewise, at the scale of an individual landslide, excavations for road cuts may be of similar magnitude as natural erosion of channels crossing or adjacent to a landslide. However, road fills and drainage diversions can potentially create conditions that would not occur naturally.

We can also view responses to forest practices at a landscape scale, encompassing an entire population of landslides. This is the scale at which cumulative effects are evaluated (MacDonald, 2000; Reid, 2010) and over which watershed analyses are conducted (Section 11 of the Forest Practices Board Manual). Schedule L-1 of the Forest and Fish Report specifies a performance target for mass wasting sediment delivered to streams of “no increase over natural background rates from harvest on a landscape scale on high risk sites”.

Forest practices and natural disturbances (fire, disease, wind) alter or remove forest cover with consequent increases in runoff and recharge that can persist for decades (for effects of wildfire, see review in Neary et al., 2005). Differences exist in the frequency and spatial extent of natural versus human-caused changes in forest cover. In general, for the west side of the Cascades (with a humid climate and long fire recurrence intervals) harvests occur more frequently than fire. The cumulative effect over time is an increase in the proportion of the landscape covered with young forest stands. For the Coast Range of Oregon, Teensma et al. (1991), for example, found that the proportion of area in stands less than 50 years old increased from 10.5% to 40.8% (including burned areas in both cases) between 1920 and 1940. Most of this change was attributed to harvest, not fire. The proportion of area in stands less than 30 years old was, as of 2004, about 43% (Wimberly et al., 2004). In a fire simulation for the Coast Range, Wimberly et al. (2000) estimated that the proportion of area in stands less than 30 years old would vary between about 10% to 30% (those are the 5% and 95% quantiles), with a median of 17%.

An increase in the proportion of area in young stands may translate to an increase in the proportion of deep-seated landslides and their source areas for water covered by young stands. Young, hydrologically mature stands may actually transpire more water than older stands during the growing season, as found by Moore et al. (2004) in a comparison of 40-year and 450-year-old riparian, conifer-dominated stands in western Oregon. However, interception tends to dominate water losses to evapotranspiration for conifer forests in the Pacific Northwest, so that total evapotranspiration is reduced in young forests (< 15 years stand age based on measurements by Jassal, 2009). When integrated over large areas, an increase in the proportion of area occupied by young forest stands may thus translate to an increase in the time that deep-seated landslides are exposed to increased water supply (see discussion in Section 9.2.2). This increase may translate to increased rates of landslide activity (Benda et al., 1998). These inferences are speculative; there are no studies that examine the landscape-wide effects that changes in forest cover may have on deep-seated landslide activity, but given the landscape-scale mandate of the Forest and Fish Report, it is important to consider these potential effects. Further analysis may demonstrate that the cumulative effects of forest management on deep-seated landslide rates are insignificant, but we will not know until such analyses are made.

Management influences on forest cover differ east of the Cascades. There, fire suppression has resulted in an overall increase in forested area (Hessburg et al., 2000).

Natural erosional processes alter hillslope and deep-seated landslide geometry. Channel incision, river-bank erosion, and smaller shallow and deep-seated landslides occurring within larger deep-seated landslides; all change slope geometry and alter the balance of forces within a slope that can reduce stability of a deep-seated landslide. Dragovich et al. (1993a), for example, found that 37% of observed deep-seated landslides were associated with undercut slopes along streams confined by steep slopes for an area in the southern Washington Cascades. Cut slopes and side cast associated with forest road construction, borrow pits and mines, and any other excavation or material-dumping also change slope geometry and alter the balance of forces, which can reduce stability of deep-seated landslides. Natural alterations to slope geometry are focused in specific topographic locations: valley floors, bedrock hollows, along steep channels. Human-caused alterations to slope geometry can occur anywhere. For example, roads follow contours crossing slopes at all relative elevations, from valley floor to ridge top. Thus, the spatial distribution of human-caused reductions in landslide stability is different than the distribution of natural reductions.

Natural processes create spatial and temporal variability in the factors that affect deep-seated landslide activity. Forest practices alter these distributions. We have found no examples in the literature where consequences for the spatial and temporal distribution of deep-seated landslide activity have been explored.

### *2.6.4 What is the relative influence of forest practices compared to natural factors?*

Precipitation amounts vary substantially across time and space. Annual precipitation, even when spatially averaged over low-land Puget Sound, varies by nearly plus or minus 40% year-to-year from the long-term average (see Section 9.2.2). Following clear-cut and patch-cut timber harvest, annual water yield increases about 6mm for every percentage point of a basin harvested (Moore and Wondzell, 2005). Mean annual precipitation over low-land Puget Sound averages about 1,100 mm/yr. Depending on the proportion of contributing area cut, these numbers suggest that harvest-related increases in runoff and recharge can be of the same magnitude as annual variability in precipitation. These harvest-related increases are overprinted on annual variability, and the combination of above-average precipitation plus harvest-related increased water yield could cause water inputs to a landslide beyond what might occur naturally. An average-precipitation year plus increased water yield from harvest may not trigger reactivation of a landslide; an unusually wet year alone may not trigger reactivation, but the combination of an unusually wet year and harvest-related increased water yield may be enough to reactivate a landslide.

Effects of natural disturbances are also over printed on temporal variability in precipitation. Wild fire can cause increases in water yield of similar magnitude as timber harvest. However, as described above, the cumulative effects of forest practices can differ from those of natural disturbances.

## 2.7 Assessment of forest practices role in landslide susceptibility

### *2.7.1 Can relative levels of response to forest practices be predicted by key characteristics of non-glacial deep-seated landslides and/or their groundwater recharge areas?*

Deep-seated landslide properties and spatial distribution are governed by regional geology, topography, and climate, so it is plausible that levels of response could be predicted by key characteristics. Such characteristics have been identified for shallow landslides; e.g., the Rule-

Identified-Landforms of WAC 222-16-050[8]. We are aware of no efforts that have systematically sought relationships between deep-seated landslide activity, forest practices, and landslide characteristics. However, as described in Section 2.3.2 above, observed correlations between landslide density (the proportion of area covered by landslide features) and levels of landslide activity with geology, topography, and climate might be used to infer relative levels of response to forest practices. Ultimately, however, detection of a potential management signal requires analysis of forest-practice treatments in addition to effects of natural factors.

### *2.7.2 What are the best methods to assess reactivation potential from dormantdeep-seated landslides?*

*Geotechnical models* to assess deep-seated landslide stability are well developed and widely applied (e.g., Duncan et al., 2014; Turner and Schuster, 1996). The confidence one can place in results from these models is directly related to the type, amount, and precision of the input data. High confidence requires abundant sub-surface and monitoring data. Such data are not generally available for regional assessments or for forested landscapes, although they could be collected.

*Statistical analyses*. Landslide potential can also be characterized in terms of statistics for a population of landslides; e.g., using correlations of landslide occurrence with landscape and storm attributes. Such methods have been developed and are widely applied (Corominas et al., 2014; Pardeshi et al., 2013), but primarily for shallow landslides. Deep-seated landslide locations can be identified; they have been routinely mapped using aerial photography combined with field verification (e.g., Dragovich et al., 1993a; Gerstel, 1999), and high-resolution elevation data derived from LiDAR has been incorporated into recent mapping methodologies (Burns and Madin, 2009; McKenna et al., 2008; Pierson et al., 2016; Schulz, 2007). However, deep-seated landslide features may persist for thousands of years, and many identified and mapped landslides show no evidence of activity over the time frame relevant to the morphologic and vegetation indicators of movement, which span hundreds of years. Mapping landslide locations shows where landslides have taken place; however, it does not show where landslide movement is *likely* to occur.

*Age distribution*. Probability of future landslide activity can also be estimated from the frequency of past landslide events. Dating of landslide deposits across a population of deep-seated landslides provides a measure of both the frequency of landslide activity and how that frequency varied over time (Ballantyne et al., 2014a; Booth et al., 2017). In current practice, relative landslide age is commonly estimated via simple field observations of surface morphology (e.g., hummocky versus undulating), drainage and soil development, and characteristics of current and historic timber stands. (Keaton and DeGraff, 1996; Table 2 in Chapter 16 of the Forest Practices Board Manual)[9]. The use of Lidar-derived digital elevation models (DEMs) as a means to categorize landslides as active/recent, dormant distinct, dormant indistinct, or relict (as in Table 2 of the Board Manual) could be utilized for recognizing deep-seated landslides that may be more

[8] http://apps.leg.wa.gov/wac/default.aspx?cite=222-16-050

[9] http://file.dnr.wa.gov/publications/bc_fpb_manual_section16.pdf

susceptible to future movement. LaHusen et al. (2016) and Booth et al. (2017) calibrated surface characteristics of landslide deposits determined using Lidar bare-earth DEMs to $^{14}C$ ages of wood in landslide deposits. Such methods may provide calibrations for using Lidar to estimate the age distribution for a population of landslides.

Landslide age distribution, however, does not tell us which landslides are more likely to become active or respond to forest practices. By identifying which landslides are or have recently been active, the statistical techniques developed for shallow landslides could be applied to identify characteristics associated with deep-seated landslide activity, reactivation, and potential historic triggers. Such a statistical approach may be, at least initially, the best method to apply for assessing reactivation potential in the context of forest practices, but it needs to be developed and tested. This approach could incorporate regional geologic, topographic, fluvial, and climatic attributes that contribute to deep-seated landslide behavior, as mentioned in the previous section. It could also incorporate geotechnical estimates of stability.

Geotechnical models can be applied using the limited data available. Without information on subsurface conditions, model results for individual landslides have large uncertainty, but when applied over a population of landslides, they may reveal interactions between topography, geology, and climate that provide useful indicators of landslide sensitivity to environmental perturbations. Broad spatial application of simple geotechnical models have been used to assess potential spatial variability in slope stability (e.g., Mergili et al., 2014) and sensitivity to change (e.g., Miller, 1995). Spatially distributed hydrologic models have been used to estimate runoff and recharge (e.g., Du et al., 2016). Such models can provide relative estimates of stability and sensitivity for all landslides in a sampled population; these estimates can be tested against empirical observations of landslide activity using the same statistical methods cited above. Such a combination of physical modeling and empirical correlation of model results to observed landslide behavior may provide more reliable predictions of activity level and reactivation potential than either physical models or empirical correlation alone.

## 2.8 Mitigation measures and basis for their determination

Landslide mitigation measures are widely applied and well documented in the literature (e.g., Turner and Schuster, 1996). The Washington Department of Transportation (WSDOT) has extensive experience with landslide mitigation. WSDOT mitigation design is predicated by subsurface data collection and stability analyses; such data are not typically collected for forest-practices slope stability assessments, although they could be.

A large variety of mitigation strategies have been devised and implemented. These commonly involve three approaches: drainage systems that either prevent water flow into landslides or drain water from within the landslide, physically reshaping the slope (commonly by toe buttressing), and strengthening the slope by installing engineered materials.

Another potential mitigation strategy is simply to maintain forest cover within the source areas for water flowing to a deep-seated landslide. This is the basis for including the groundwater recharge area to glacial deep-seated landslides as a rule-identified landform in WAC 222-16-050. As described above, timber harvest increases water yield for one or more decades, which may increase water supply to a deep-seated landslide over that period. If that increase occurs during a period of high precipitation, the potential for triggering or accelerating movement on deep-seated landslides increases. However, long climate records reveal decadal trends in precipitation and temperature. Burt et al. (2015) suggest that recent improvements in characterization of these climate trends might allow for timing of harvests to minimize impacts on water yield. Burt et al. were addressing hydrologic effects on stream flow, but the same strategy might be applied to slope stability. Harvest in source areas for water to deep-seated landslides that are potentially sensitive to increased water yield could be done during the low-precipitation portion of these decadal cycles.

# 3 Knowledge Gaps

It is a bit surprising that, other than Swanston et al.'s 1988 study, there is apparently no published work investigating effects of forest practices on deep-seated landslides. This paucity is not for lack of recognition of potential effects. In a review paper titled "Timber harvesting, mass erosion, and steepland forest geomorphology in the Pacific Northwest", Swanston and Swanson (1976) state:

> "Although the impact of clearcutting alone on slump-earthflow movement has not been demonstrated quantitatively, several pieces of evidence suggest that it may be significant. In massive, deep-seated failures, lateral and vertical anchoring of tree-root systems is negligible. However, hydrologic impacts appear to be important. Increased moisture availability due to reduced evapotranspiration will increase the volume of water not utilized by the vegetation. This water is therefore free to pass through the rooting zone to deeper levels of the earthflow. Although the hydrology of slump-earthflow has not yet been investigated, hydrology research on small watersheds suggests that this effect may be substantial."

Subsequent work in the following 40 years has clearly demonstrated the hydrologic impacts of timber harvesting in increased water yields, so why have only Swanston et al. (1988) examined effects on deep-seated landsliding directly? Several factors might contribute to the lack of studies:

- There are many deep-seated landslides, the majority of which are inactive. A query of the Washington DNR landslide database[10] shows that, of the 4,640 deep-seated landslides mapped in the Landslide Hazard Zonation Project (UPSAG, 2006), 573 are classified as "Activated, reactivated, recent", and these account for only 3.7% of the total mapped deep-seated landslide area. Most deep-seated landslides are inactive and probably have no response to forest practices.
- The morphologic and vegetative indicators of landslide activity and sensitivity to forest practices are diverse and may be ambiguous, particularly for slow-moving landslides.

[10] Available at http://www.dnr.wa.gov/programs-and-services/geology/publications-and-data/gis-data-and-databases

- Physical factors affecting deep-seated landslide behavior are diverse and difficult to characterize. These include geometry and strength of the shear zone and groundwater response to precipitation.
- Groundwater levels may respond to cumulative precipitation over time periods spanning days to years. Thus cause (precipitation) and effect (landslide movement) may be separated in time, and the sequence of precipitation events that can trigger landslide movement may not be easily recognized.
- Hydraulic conductivity of a landslide body may increase or decrease over time. This changes landslide response to precipitation.
- Deep-seated landslides respond to natural processes, such as variations in precipitation, channel incision and lateral migration, and seismic shaking. Effects of forest practices are over-printed on these responses. To separate forest-practice effects from natural variability requires monitoring over long time periods (Swanston et al. reported on observations over ten years) and potentially over many sites.

The relatively high costs of instrumenting sites to monitor precipitation, groundwater, and landslide movement; the decade (or more) long time over which data must be collected, and the high potential that the results will be inconclusive are all high motivation for any aspiring researcher to avoid such studies.

Contrast this with the situation for shallow landslides.

- Shallow-landslide scars tend to be rapidly revegetated, so every mapped shallow landslide is recent; there is no ambiguity as to whether a shallow landslide was active or not.
- Shallow landslide occurrences are numerous: the Mass Wasting Effectiveness Monitoring Project (Stewart et al., 2013) mapped 1098 non-road related shallow landslides and 10 deep-seated landslides after the 2007 storm.
- Shallow landslides tend to occur during large precipitation events, whereas deep-seated landslides may respond to cumulative precipitation spanning many events.
- Physical factors affecting shallow landslides can be characterized using simple infinite slope and steady-state groundwater models.

Thus, it is straightforward to rapidly collect abundant data with well-defined timing of landslide events using field and aerial-photograph mapping, and to characterize shallow landslide occurrences using readily calculated topographic attributes of slope and contributing area. Significant correlations of active shallow landslide locations with forest roads and stand conditions exist and interpretations can be actively debated (e.g., Miller and Burnett, 2007; Montgomery et al., 2000; Turner et al., 2010). We found no similar efforts to relate active deep-seated landslide locations with forest roads and timber harvest.

The discussion above highlights several key information gaps or uncertainties not covered in the literature.

## 3.1 Features associated with deep-seated landslide reactivation potential or sensitivity to forest practices have not been identified

The literature recognizes and documents features that are indicative of the *current* or *past* level of landslide activity (see Chapter 16 of the Forest Practices Board Manual, for example), but there is little documentation of features indicative of landslides that are likely to *become* more active, particularly of those likely to become more active in response to forest practices. The information needed to identify these features can be grouped into several types.

### *3.1.1 Factors associated with rate of activity and rate of reactivation of deep-seated landslides in Washington*

Landslide inventories provide information on the location of landslide features. In some cases, the timing and nature of the triggering event can also be recorded. Kirschbaum et al. (2016) are working to add such information to existing inventories in the Pacific Northwest, but this information is unknown for most mapped landslide features. Many inventories, however, provide information on the relative age and level of activity estimated from observations of surface roughness and vegetation. Table 2 in Section 16 of the Forest Practices Board Manual lists four categories for level of activity used for landslide inventories in Washington: active/recent, dormant-distinct, dormant-indistinct, and relict. The Board Manual also lists field indicators of relative activity that define the current standard of practice and are commonly used by field practitioners. Relative age and level of activity have not been systematically compared to other observations in a search for factors that can help predict level of activity. These factors might include rock and soil types, geologic structure, geometry (size, shape, relief, profile) of landslide, hillslope and valley features, metrics of climate, past valley glaciation, and seismicity.

Success in identifying features that correlate with and may help predict level of activity and potential for reactivation of landslides will vary with the confidence in the level of activity that has been associated with each landslide in an inventory. Generally, few landslides in an inventory have field-verified indicators of activity, and even fewer have quantified measures of rate and timing of landslide movement. Without this information, the statistical techniques used for shallow landslides cannot be applied to deep-seated landslides.

### *3.1.2 Factors associated with groundwater response within landslides in Washington to precipitation*

Groundwater response is a key factor for anticipating landslide activity and sensitivity to forest practices. Information on groundwater response requires monitoring of precipitation and subsurface groundwater levels. Such data are available, but have not been used to systematically address questions that can lead to better characterization of landslide hazard and sensitivity to forest practices. These questions include:

1. How do groundwater levels vary in response to hourly, daily, weekly, seasonal, annual, and multi-year variations in precipitation and snow melt?
2. What factors influence the magnitude of these groundwater-level variations?

3. What proportion of groundwater in a landslide originates from runoff and groundwater recharge from areas outside the landslide?
4. What factors influence this proportion of groundwater?
5. How do these proportions vary over time?
6. How does timber harvest alter these proportions?
7. How do water inputs from these areas outside the landslide affect groundwater response?

#### *3.1.3 Factors associated with landslide response to variations of groundwater levels*

These controls are well understood in terms of geotechnical theory, but have not been translated to guidelines for anticipating the magnitude and timing of these variations.

#### *3.1.4 Factors associated with landslide response to patterns of precipitation*

Rainfall thresholds for initiation of landslide movement are widely used for anticipating shallow-landslide activity (e.g., Caine, 1980; Godt et al., 2006). Rainfall thresholds have also been used for anticipating onset of deep-seated landslide activity, although these tend to be landslide specific and must account for antecedent conditions that may span many months (e.g., Floris and Bozzano, 2008; Vallet et al., 2016). Use of rainfall thresholds is appealing, because precipitation is more easily measured than groundwater levels; we can see how it has varied in the past and anticipate how it will vary in the future. Identification of the factors that influence a landslide's response to precipitation (and snow melt), which essentially involves everything listed in the previous three subsections (3.1.1-3.1.3), might help us to explain why some deep-seated landslides are active, but most are not, and to recognize those landslides most likely to become active in the future.

### 3.2 Runout extent for deep-seated landslides in Washington has not been systematically characterized

The range of runout extents in Washington has not been determined. The variation of runout extent with landslide attributes, such as size (i.e., volume), relief, type, material, and relative position on the slope has not been systematically characterized. Recent work by Perkins et al. (2016) using LiDAR-derived DEMs to map runout extent in glacial sediments in northwest Washington shows that such efforts are feasible.

### 3.3 Accuracy of current methods for assessing landslide hazard and sensitivity to forest practices is unknown

Section 16 of the Forest Practices Board Manual provides guidelines for identifying potentially unstable slopes, but there is no protocol for determining or recording the success of the resulting determinations. This require a specific prescription effectiveness study, which has not been done, and would be required before reasonable protocols could be developed

# 4 Recommendations

In his review of an early draft of this document, Ted Turner pointed out that:

> "First and foremost, we need to determine if there is a problem. Have forest practices had any significant influence on rates of deep-seated landslide activity? Are our current practices effective within the context of acceptable risk?"

Previous studies have not addressed these issues, and any future studies will be hindered by the difficulties listed at the beginning of the last section, reiterated here:

- Deep-seated landslide features cover a significant portion of the landscape in some regions, but the proportion of those landslides that will respond to forest practices is probably quite small.
- Indicators of sensitivity to forest practices may be subtle and ambiguous.
- Without borehole and monitoring data, physical characteristics by which to assess sensitivity to forest practices must be inferred. These characteristics include geometry of the landslide shear zone and response of groundwater to precipitation. Estimates of landslide stability based on geotechnical models therefore have large uncertainty.
- A deep-seated landslide may respond to cumulative precipitation integrated over time spans of days to years. Relationships between precipitation, reductions of evapotranspiration, and landslide movement can therefore be complex and difficult to identify.
- Hydraulic conductivity of the landslide body can change over time, thereby altering landslide response to precipitation and sensitivity to forest practices.
- Forest practice effects are overprinted on large natural variability in precipitation patterns, on changes in geometry associated with natural toe erosion, and on seismic shaking from earthquakes.

UPSAG reviewers and the technical advisory team all highlight the need for detailed observations at the spatial and temporal scales of a harvest unit. We agree with that, but given the points listed above, studies focused on individual landslides will need to include a large number of landslides to discern forest-practice effects. We therefore advocate as a first step examination of the entire population of deep-seated landslides using statistical techniques, like those applied for shallow landslides, but modified as needed for this application. This statistical analysis can aid in defining a sampling strategy for choosing sites for more detailed analysis.

Many of our recommendations, therefore, involve use of existing information and collection of new information to characterize the population of deep-seated landslides across the state. Statistical techniques to identify potential correlations between landslide activity and forest practices require identification of landslides that are active and, ideally, quantification of the level of activity, so we include strategies for identifying and quantifying level of activity across a population of deep-seated landslides. Field observations and monitoring will still be required to quantify physical relationships, but we think that a prior statistical analysis will provide information critical to design and implementation of effective field efforts.

## 4.1 Leverage existing information

### *4.1.1 Combine existing landslide inventories with other available data to seek statistical correlations between estimated level of activity and attributes of the landscape and climate*

Statistical analyses involving landscape features are most effectively done using digital data with a Geographic Information System (GIS), so we focus here on data available in digital format.

Digital landslide inventory data are compiled and available from the Department of Natural Resources (http://www.dnr.wa.gov/publications/ger_portal_landslides_landforms.zip). These include inventories from the Landslide Hazards Zonation Project, which were collected using a consistent protocol and included categories for activity level and confidence (UPSAG, 2006). Slaughter (2015) describes inventory data available with this compilation and discusses issues with consistency, accuracy, and resolution. The Division of Geology and Earth Resources within the Department of Natural Resources has initiated a landslide mapping program using newly acquired LiDAR data and applying methods described by Burns and Madin (2009) and (Burns and Mickelson, 2016). Results from a pilot project in Pierce County should soon be available (Slaughter and Mickelson, 2016) and mapping will continue as additional LiDAR data are collected. King County has also applied these techniques to map deep-seated landslides along river corridors; these data are available at http://www.kingcounty.gov/services/environment/water-and-land/flooding/maps/river-landslide-hazards.aspx.

Digital data to apply to these landslide inventories include:

- *Geologic mapping* at 1:24,000 and 1:100,000 scales. These provide rock type, bedding orientations, fold orientations, and fault locations.
- *Soils mapping* at 1:24,000 and 1:100,000 scales. Soils maps include estimates of soil depth and permeability.
- *Weather station and climate data*, such as the summaries of temperature and precipitation provided by the PRISM Climate Group (www.prism.oregonstate.edu/).
- *Digital Elevation Models (DEMs)*, particularly LiDAR-derived bare-earth DEMs (available from the DNR www.dnr.wa.gov/lidar and the Puget Sound LiDAR Consortium pugetsoundlidar.ess.washington.edu/).

Methods exist to extract a variety of topographic attributes from DEMs. This information includes attributes of landslides themselves:

- *Size*. The size distribution for a population of landslides is found to vary with controls on landslide formation and activity (Catani et al., 2016). Differences in the size distribution as a function of activity level may indicate differences in current and past controls on regional landslide activity.

- *Surface roughness*. Measures of surface roughness may correlate with ages of landslide deposits, as LaHusen et al. (2016) and Booth et al. (2017) demonstrate for landslides in the North Fork Stillaguamish valley.
- *Surface morphology*. Dewitte et al. (2010), for example, found that surface gradient, aspect, and profile curvature provided useful predictors for reactivation potential of deep-seated landslides in weathered sedimentary rocks in Belgium.
- *Surface displacement*. Comparison of changes in ground-surface elevation over time using Lidar-derived DEMs from different years can identify slope movement and changes in slope geometry (e.g., Cavalli et al., 2016; Prokešová et al., 2014).

Attributes of the area where landslides occur can also be mapped from DEMs:

- *Contributing area*. This delineates source areas for storm runoff to a landslide and an approximation of the groundwater recharge area. Geology and topography of the contributing area can also be determined.
- *Drainage density, closed depressions, lineaments*. These provide indicators of permeability and pathways for enhanced infiltration of water.

These data can be used to address questions to help identify features associated with landslide occurrence.

1) How do landslide characteristics vary across the state?
   - How well does the spatial distribution of the landslide inventories include the range of topographic, geologic, and climatic conditions across the state?
   - How does deep-seated landslide density (proportion of total area occupied by landslides) vary across landslide provinces (Thorsen, 1989a)?
   - Within a province, how does landslide density vary with rock type and topographic attributes (such as valley relief)?

2) Are landslide or landscape characteristics associated with level of landslide activity?
   - What proportion of landslides fall into each activity-level class (active/recent, /dormant-distinct, dormant-indistinct, relict)?
   - Are these proportions related to climate attributes, such as mean annual precipitation, or average snow cover?
   - Does the size distribution of landslides in different activity classes vary?
   - Does surface roughness vary with activity class?
   - Does the ratio of upslope contributing area to landslide size vary with activity class?
   - Do characteristics of the landslide profile, such as scarp gradient and convexity of the deposit, vary with activity class?
   - Does the landscape position (relative elevation between valley floor and ridge top) of the head scarp vary with landslide type and activity class?

These are relatively easy analyses to perform, and the results may further highlight data gaps, reveal other productive avenues of inquiry, or identify areas of uncertainty that may not be resolvable. Examples of such analyses include that of Safran et al. (2011), who examined topographic and geologic controls on large landslides in eastern Oregon; Dewitte et al. (2010), who examined relationships of topography, land cover, and land use to reactivation of deep-seated landslides in Belgium; Crosta et al. (2013), who examined geologic, topographic, climatic, and glacial-history controls on large deep-seated landslides in the European Alps.

### *4.1.2 Use physical models with statistical analyses*

Use available data with physical models to calculate:

1. Level of landslide stability; i.e., a factor of safety.
2. Landslide sensitivity to changes in pore pressure and toe erosion; e.g., how much does the factor of safety change with a unit change in pore pressure?
3. A water budget for the landslide; e.g., proportion of inflow from direct infiltration, runoff from upslope, and groundwater seepage from below.
4. Fluctuations in water supply to the landslide from temporal variability in precipitation.
5. Effect of forest cover on the water budget and temporal fluctuations in water supply.
6. Magnitude of pore pressure fluctuations within a landslide caused by estimated temporal fluctuations in water supply.

Physical models have been widely applied for regional assessment of shallow landslide potential (e.g., Baum et al., 2008; Formetta et al., 2016; Montgomery and Dietrich, 1994). Physical models for site-specific analysis of deep-seated landslides are well developed and broadly applied (e.g., Rocscience SLIDE: www.rocscience.com/rocscience/products/slide) and could be readily used for items 1 and 2 above. Techniques for regional application of physical models to assess deep-seated landslide potential have been developed (e.g., Miller, 1995; Reid et al., 2015), and to link hydrologic and physical models (Brien and Reid, 2008; Miller and Sias, 1998).

Results from the physical models serve as another input for statistical analyses; we can seek correlations between the calculated values for items 1, 2, 4, 5, and 6 above and observed levels of landslide activity.

The data inputs to these models are insufficient to provide high confidence in estimates of landslide stability or sensitivity. Rather, the models are used to integrate available information about each landslide based on our understanding of physical geologic and hydrologic processes. The distribution of calculated values provides an additional way to characterize a population of landslides. Statistical analyses can then be applied to see how calculated values of stability, sensitivity, and precipitation response correlate with observed activity levels.

### *4.1.3 Compile and use data from slope stability assessments of Forest Practice Applications*

Success in using the types of analyses listed above to identify controls on level of landslide activity is limited by the qualitative and approximate nature of the categories of activity level assigned to each landslide. For landslides in many inventories, activity level is based on interpretation from aerial photographs or LiDAR shaded relief images.

Geotechnical assessments of slope stability for forest practices provide field-verified activity level and landslide boundaries. These studies provide a separate sample of landslides for the types of analyses listed above, and they provide a means of assessing the accuracy and completeness of the landslide inventories. These studies also provide an assessment of sensitivity to forest practices and potential for future activity.

Landslides identified and mapped during these assessments should be recorded in a digitized landslide inventory. Information from the field assessment should be included as attributes for the landslide. Subsequent forest practices should be recorded as part of the inventory, as should the data and magnitude of any subsequent movement of the landslide.

#### *4.1.4 Compile and use data from detailed geotechnical investigations*

Potential data sources include the Washington Department of Transportation (see notes from Tom Badger in Appendix A), the Forest Service, county and city investigations (e.g., see the compilation of geotechnical investigations for the Cowlitz County urban corridor in Wegmann, 2006), and geotechnical investigations required for development permits from county and city governments.

Data include bore logs, which provide stratigraphy, depth to shear zones, and groundwater levels. Monitoring may include time series of groundwater levels, landslide movement, and precipitation. Some studies provide in-situ or laboratory measurements of material properties.

Compiled data can then be used to:

- Identify common patterns in landslide geometry and style of behavior.
- Seek correlations of those landslide patterns with the attributes and model predictions from analyses described above in 4.1.1 and 4.1.2.
- Identify common patterns in groundwater-level fluctuations and landslide movement.
- Seek correlations of those groundwater patterns with time series of precipitation and with the attributes and model predictions from analyses described above.
- Evaluate the accuracy of existing landslide inventories – were the studied landslides identified in the inventories?
- Evaluate the success of physical models applied using only surface information (Section 4.1.2) to predict subsurface conditions, groundwater response, and landslide behavior.

### 4.2 New information to collect

#### *4.2.1 Field verification for a subset of sites*

As mentioned above, confidence in assigned activity levels is low for many landslides in existing inventories. Field visits could be done for landslides that appear anomalous in the analyses described above. For example, landslides with attributes characteristic of the active/recent class, but assigned to a different class, could be visited on the ground to verify the assigned class.

### *4.2.2 InSAR analyses for rates of movement*

Interferometric Synthetic Aperture Radar provides a remotely sensed means of measuring landslide surface displacements using satellite imagery. In the Pacific Northwest, InSAR has been used to measure landslide movement for earthflows in California (Handwerger et al., 2013), for landslides across northern California and southern Oregon (Zhao et al., 2012), for bedrock landslides along the Columbia River in southwest Washington (Hu et al., 2016; Tong and Schmidt, 2016), and within the Stillaguamish Basin in Washington (Sun et al., 2015).

Application of InSAR is hindered by steep terrain and vegetation (see for example, Alex Grant's project report for David Schmidt's InSAR class at UW: faculty.washington.edu/dasc/InSAR/alex:main:report), and detects movement only in the satellite line-of-sight direction. Instrumentation and analysis methods to overcome these issues are being rapidly developed (Wasowski and Bovenga, 2014), and study sites can be selected to minimize their influence.

InSAR provides a means to detect and quantify movement for landslides within a selected area over a selected time. In the context of the types of analyses described above, InSAR can serve to:

1) Quantify the rate of activity for landslides within a sample. This will reduce uncertainty in the dependent variable for statistical analyses to identify controls on rate of activity (Bianchini et al., 2013; Oliveira et al., 2014).
2) Test predictions made from analyses based on existing landslide inventory data. This will verify if landslides identified as the most likely to exhibit activity *are* actually the most active.

InSAR can also serve as a data source for monitoring. It has been used to detect changes in landslide movement rates, providing insights to landslide response to variations in precipitation and climate (e.g., Bennett et al., 2016; Handwerger et al., 2013). A pilot study to assess applicability of InSAR for these purposes could also evaluate its use for more regional landslide monitoring and hazard assessment.

### *4.2.3 Instrumentation and monitoring of selected sites*

A strong conceptual framework exists for identifying potential relationships between forest practices, local hydrology, and deep-seated landslide behavior; it is outlined in the background material presented in Sections 6 through 9 and provides the basis for many of the analyses suggested above. However, little empirical data have been collected to directly test these concepts. A logical step, therefore, is to identify appropriate field sites, pose hypotheses about groundwater and landslide responses to future precipitation and forest practices, install arrays of piezometers, inclinometers, surface benchmarks, and precipitation gages, and collect data to test hypotheses and, if needed, modify conceptual frameworks.

Such field studies have been performed to hone understanding of processes affecting shallow landslides. Work by Bill Dietrich and his students near Coos Bay, Oregon, for example, advanced understanding of hydrological processes driving shallow landslides (e.g., Anderson et al., 1997; Montgomery et al., 2002; Montgomery et al., 1997; Torres et al., 1998).

Success of field instrumentation and monitoring studies will depend greatly on site selection and study design. Results of statistical and modeling studies as described above can guide those efforts, providing information for identifying representative field sites and predictive models for posing hypotheses that rigorously test the conceptual models they are based on.

#### *4.2.4 Landslide ages*

LaHusen et al. (2016) and Booth et al. (2017) show that measures of surface roughness of large landslide deposits in the North Fork Stillaguamish valley vary systematically with carbon-14 dates for wood excavated from the deposits. They use this relationship with a model for diffusion of surface features over time to calibrate a model to estimate landslide-deposit age based on a LiDAR-derived measure of surface roughness. Surface roughness provides a potential means of estimating the time-since-occurrence, or even rate of activity, for individual landslides and for determining the age distribution for a population of landslides. These measures are valuable for identifying potential climatic and seismic controls on landslide activity (e.g., Ballantyne et al., 2014b) that might aid in assessing current landslide sensitivity.

It is unclear how broadly applicable surface-roughness-based assessments of landslide age might be (e.g., Goetz et al., 2014); the studies cited above were for landslides in glacial sediments. However, it is possible to find out. Collection and analysis of samples for $^{14}C$ dating is straightforward and inexpensive; it could be included as a basic item in landslide assessments. Analysis such as those cited above for the NF Stillaguamish valley could eventually be performed for other populations of landslides in specific geomorphic and climatic settings across the state.

### 4.3 Retrospective analyses of accuracy of stability assessments

As mentioned previously, the accuracy of past and current methods for assessing deep-seated landslide instability and sensitivity to forest practices rules is unknown.

Wendy Gerstel, a member of the science advisory team for this project, provided the following suggestion for synthesis of post-harvest stability and effectiveness of pre-harvest geotechnical characterization for non-glacial deep-seated landslides (note that the same could be applied for glacial deep-seated landslides):

> "Since the implementation of the Forest Practices Rules in the 1970s, Forest Practice Applications (FPAs) in areas with rule-identified unstable landforms require a geotechnical assessment by a qualified expert to evaluate the potential for harvest-related sediment delivery to streams. Numerous geotechnical assessments have been conducted in these areas and reports written as part of the FPA submittal and review process for FPAs classified both as III and IV Special.

A retrospective study of these reports could be used to evaluate impacts of forest practices on non-glacial deep-seated landslides. Such a study could also shed light on the adequacy of the geotechnical assessments to mitigate harvest impacts on non-glacial deep-seated landslides. The proposed study would review findings and recommendations of the geotechnical assessments, determine whether or not recommendations outlined in the report were applied to the harvest area, and conduct field observations and record any post-harvest slope movement. Study results would provide data necessary to evaluate the effectiveness of the mitigation proposed in the original pre-harvest geotechnical report. Such a retrospective study would also identify any additional data necessary to improve the geotechnical characterization of a site to reduce potential harvest-related impacts to unstable landforms."

### 4.4 Implement GIS-based tools and field-based guidelines to apply results of above analyses

The recommendations above involve use of statistical and physical models with digital GIS and remotely sensed data. Methods developed and employed for these analyses should, to the extent possible, use readily available software implemented with accessible GIS user interfaces. For example, government agencies in Washington primarily use ArcGIS. Analysis methods used and developed should be implemented as ArcGIS tools or add-ins.

These analyses can provide maps that show landslides ranked by potential activity level, potential for reactivation, and sensitivity to forest practices. Such maps can be used as screening tools, identifying sites that require additional scrutiny. That scrutiny will typically require on-the-ground evaluations. Maps can also be produced to show the data elements used for the calculated rankings. These may include mapped landslide boundaries, landslide surface roughness, delineation of the estimated contributing area, upslope geological and topographic features, proximity to streams, and other attributes that should be field verified. Such maps could be created for all inventoried deep-seated landslides and provided as an online resource. Guidelines for on-the-ground evaluation of landslide rankings should also be developed and evaluated by on-the-ground users.

# 5 Background

The high-definition view of terrain provided by LiDAR has highlighted the abundance of deep-seated-landslide features across the landscape. Such features were certainly recognized before LiDAR, but now that they can be seen clearly, many more are seen than before (Figure 1).

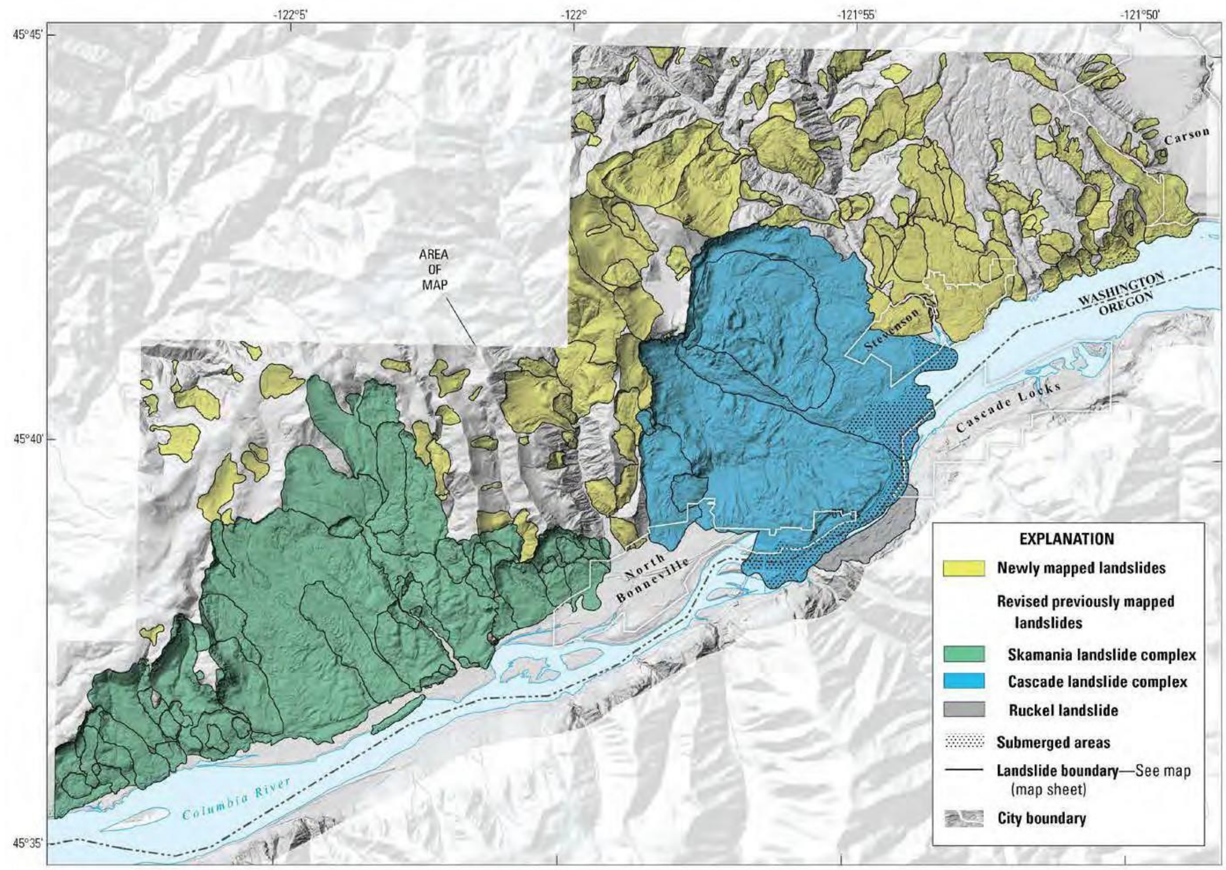



**Figure 1. A newly published map of deep-seated landslides along the Columbia Gorge.**

From Figure 6 in Pierson et al. (2016). The yellow polygons indicate landslides not previously mapped.

The ensemble of deep-seated-landslides records a long history of landslide events. Although we can now see this record clearly, the risks *currently* posed are not necessarily clear at all. Of the 215 landslides shown in Figure 1, twelve are known to be currently active or to have moved in the past two decades. The remainder span a range of estimated ages exceeding 15,000 years. What threat does a 15,000-year-old landslide scar pose? What does this ensemble of landslides tell us about current landslide processes?

Each of these landslides involved an initial failure of a previously intact slope. New areas continue to fail; a rock fall expanded the area of the Cascade Landslide Complex (blue polygons in Figure 1) in 2008 (Randall, 2012). In assessing deep-seated landslide hazards, however, we tend to focus on those landslides that have already occurred (e.g., Burns and Mickelson, 2016).

This is not unreasonable; most landslide activity involves pre-existing landslide features and activity on deep-seated landslides can persist for centuries (Guthrie and Evans, 2007), or resume after long periods of inactivity. Nature has shown that these are sites potentially sensitive to changing conditions, whereas the intact slopes likely to fail may offer fewer clues.

Yet the history we seek to translate to landslide hazard involved an initial failure of an intact slope for every existing landslide. The processes of initial failure provide clues to subsequent behavior, so they are worth exploring. In the next sections, I'll describe concepts of how and why slopes fail. This will provide the basis for then exploring processes driving movement and stability of existing landslide features. It also provides some commonality across landslide types. Deep-seated landslides occur over a huge range of topographic, geologic, and climatic conditions; sizes range from hundreds to millions of cubic meters; they are classified into dozens of different types and styles of movement; and each landslide is unique. Across this vast diversity, processes of landslide initiation generate certain features common to all deep-seated landslides, and these common features can provide a framework for assessing hazard and determining landslide sensitivity to forest practices.

# 6 Initiation of First-time Landslides

Certain events directly trigger failure of an intact slope, sometimes referred to as first-time landslides: earthquakes, for example, and extreme precipitation. Although a specific event may be the trigger, a long history precedes every failure, a history that preconditions a slope to fail in a particular way. Processes of rock formation create heterogeneities, such as alternating strong and weak beds in sedimentary rock. Once formed, tectonism may uplift, fold, and fault rock. Stresses associated with rock formation and emplacement create zones of weakness, referred to in rock mechanics as discontinuities: faults, joints, fractures, cracks, foliation, and cleavage. Discontinuities offer surfaces that can pull apart or slide, and that provide conduits for water to flow into and through a slope. Over geologic time, erosional and tectonic processes may exhume rock from depth, and the landscape we encounter is composed of a weakened rock mass primed to fail. River incision and glacial carving create topographic relief, and gravity then drives failure of slopes, with the style and mode of failure governed by the geologic and geotechnical nature of the materials and the subsurface geometry resulting from that long history of rock formation, tectonism, exhumation, and weathering.

## 6.1 Fractures Resulting from Topography

Topography may be a key factor in the preconditioning of bedrock for development of deep-seated landsliding and in creating features that subsequently control the flow of storm runoff and groundwater to landslides. Topography perturbs gravitational and regional tectonic stresses to create near-surface zones of tensile, shear and compressive stresses (Savage et al., 1985). These stresses can be of sufficient magnitude to initiate and enhance microcrack growth and weathering (Leith et al., 2014a; Molnar, 2004), and ultimately to fracture intact bedrock (Miller and Dunne, 1996; Slim et al., 2015). Seismic refraction surveys show systematic variations in compressional-wave velocity consistent with crack growth caused by the modeled stresses (Clarke and Burbank, 2011; Slim et al., 2015). Stress magnitude and orientation vary with relief, shape, and orientation of local topography and with the regional state of stress (Leith et al., 2014a; Miller and Dunne, 1996; Slim et al., 2015).

Bedrock at depth forms under large overburden pressure; crystals and rock particles form and interlock under high confining stresses. As rock is exhumed, the vertical component of stress is reduced, but lateral confinement maintains high horizontal stresses. This “residual” stress, resulting by exhumation of rock formed at depth, can thereby cause high, near-surface horizontal compressive stresses (Leith et al., 2014a) in any tectonic regime. These stresses are perturbed by local topography. Horizontal compressive stresses are concentrated along valley axes, while horizontal stresses within ridges are reduced. These perturbations favor development of slope-parallel extension fractures through the valley floor and lower side walls (Leith et al., 2014b; Martel, 2006, 2017) and steeply dipping fractures through the upper valley walls and ridge tops (Miller and Dunne, 1996). Such fracture sets are consistent with observed extensive features on upper valley walls and ridge tops associated with “deep-seated gravitational deformation” (Agliardi et al., 2012) and with observed groundwater flow systems associated with deep-seated landslides found in crystalline bedrock in a variety of mountainous landscapes (Binet et al., 2007a; Binet et al., 2007b; Cervi et al., 2012; Guglielmi et al., 2005; Padilla et al., 2014).

Modeling studies indicate that the stress history associated with alpine glacial erosion of valley floors may initiate a period of crack growth and fracture formation following glacial retreat (Guglielmi and Cappa, 2010; Leith et al., 2014a, b). These modeling results are substantiated by dating of extensional features (e.g., Agliardi et al., 2009; Ballantyne et al., 2014a; Beget, 1985). These dating studies also indicate that continued slope deformation can continue for thousands of years following deglaciation, persisting to the present. Upper-slope extensional features indicative of deep-seated gravitational deformation are not, however, limited to glaciated regions, but are found in mountainous regions throughout the world (Pánek and Klimeš, 2016; Pánek et al., 2015).

Gravitational deformation is widespread in mountains of the Pacific Northwest. Features indicative of deep-seated gravitational slope deformation have been reported for crystalline rocks in southwest British Columbia (Bovis and Evans, 1996), volcanic and sedimentary rocks of the North Cascades (Beget, 1985; Thorsen, 1989b), and in sedimentary rocks of the Olympic Mountains (Tabor, 1971).

Development of fracture patterns associated with topographic stress perturbations and deep-seated gravitational slope deformation (“DSGSD”) are recognized as precursors to development of deep-seated landslides (Binet et al., 2007a). Near-surface fracturing increases bedrock porosity and hydraulic conductivity, creating conditions for development of shallow, perched aquifers. Formation of tension cracks, grabens, uphill-facing scarps, and trenches on upper slopes enhance infiltration of rainwater and snowmelt into these aquifers. Formation of surface-parallel cracks and fractures in lower slopes provide conduits for downslope groundwater flow and for progressive slope failure as crack systems coalesce into through-going shear zones. Regional landslide inventories indicate that DSGSD increases susceptibility for both deep-seated and shallow landslides (Capitani et al., 2013; Jomard et al., 2014; Pánek and Klimeš, 2016).

Rates of downslope movement for slopes affected by deep-seated gravitational deformation tend to be episodic and slow (Pánek and Klimeš, 2016). However, these movements can precondition slopes for catastrophic failure (Chigira et al., 2013). Pánek and Klimeš (2016) cite twenty recent catastrophic rockslides and rock avalanches associated with DSGSD features and list possible triggering factors. Of these twenty, seven occurred in British Columbia within the past 50 years.

Uphill-facing scarps, tension cracks, grabens, and trenches provide topographic evidence of deep-seated gravitational slope deformations. These can be identified with field surveys and can be mapped using manual (e.g., Scheiber et al., 2015) and automated (e.g., Hashim et al., 2013; Mallast et al., 2011; Šilhavý et al., 2016) methods from remote sensing data, such as LiDAR DEMs. Crosta et al. (2013), for example, created an extensive inventory of DSGSD features for the European Alps using available satellite images and DEMs. It may also be feasible to identify potential zones with DSGSD based on computer-generated slope profiles (Nonomura and Hasegawa, 2013).

## 6.2 Brittle Materials

Fractures are the visible result when rock breaks. Less visible are the processes of microcrack nucleation, growth, and coalescence that precede the fracture. These micro-scale processes are recognized as precursors to macro-scale failure of slopes in both rock and soils.

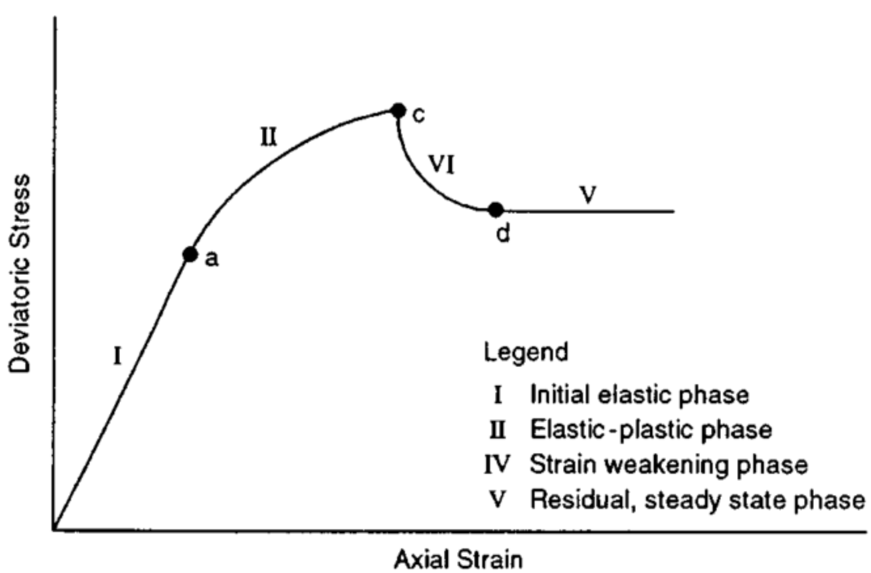



**Figure 2. Stress-strain curve for a brittle material.**

From Petley and Allison (1997). Point "a" indicates initiation of crack growth. Point "c" indicates coalescence of cracks into a through-going shear zone and failure of the sample. The stress at point "c" indicates the peak shear strength of the material. Point "d" indicates the stress supported by friction across the shear zone. This stress level indicates the residual strength of the material.

Micro-crack growth and coalescence explain the brittle behavior of rock and over-consolidated cohesive soil samples in triaxial lab tests (Figure 2). In such a test, a cylindrical sample is compressed from the ends while a confining pressure is applied to the sides. As the force applied to the ends increases, the sample shortens. If the applied force is not too great, the sample will rebound to its original shape when the force is removed. This represents elastic behavior, where the deformation is completely recoverable. However, when the applied force exceeds some threshold value, the rate of deformation increases and, upon removal of the force, the sample remains deformed. This elastic-plastic phase, which involves permanent deformation of the sample, results from the growth and displacement across microcracks within the sample. Eventually, the microcracks coalesce and shear zones develop within the sample. When the force reaches the compressive strength of the rock or soil and enough microcracks have coalesced to form a through-going surface, the sample breaks. The sample then deforms by shearing and further crushing across this surface, with friction determining the force required to cause sliding.

Brittle behavior entails shear failure occurring at some peak strength and development of a distinct shear zone, culminating in precipitous strength loss, followed by ductile deformation. Once a shear zone has formed, displacement across the shear zone is initiated at stress magnitudes less than the peak stress required to cause initial failure; this is the residual strength of the material. Rock and over-consolidated cohesive soils are brittle at low confining pressures. Soils exhibit brittle behavior below confining pressures of about 250kPa, corresponding to a depth of about 15 meters for a soil with bulk density of 1800 kg/m$^3$, and rocks behave brittlely to confining pressures of about 2Mpa, corresponding to a depth of about 70 meters for rock with bulk density of 2800 kg/m$^3$ (Petley and Allison, 1997). Deep-seated landslides commonly occur within these depth ranges.

### 6.3 Progressive Failure

Slopes in soils and rock that behave as brittle materials exhibit progressive failure, in which weakened zones grow progressively over time. Slope displacements monitored prior to failure often indicate a similar process of microcrack growth and coalescence. Displacements are initially slow, but gradually increase prior to failure (Petley et al., 2002), indicative of the elastic-plastic phase of deformation in triaxial tests. Long-term experiments (e.g., Carey and Petley, 2014) demonstrate that deformation of a sample can occur under a constant applied load; that is, that microcracks continue to grow even if the applied stress does not increase. So once a threshold stress is reached, a threshold considerably less than the actual breaking strength of the rock or soil, microcracks grow. Over time, as microcracks coalesce, zones of shear failure form within the slope. Numerical analysis of fracture growth indicate that such shear-failure zones form initially at depth, with no surface indications other than downslope creep (Martel, 2004). Over time, the slope gradually weakens as those internal failure zones expand. Eventually, as shown in Figure 3, an event, such high groundwater levels or an earthquake, perhaps no greater than many other such events endured by the slope in the past, then triggers failure (Petley et al., 2005).

Progressive failure and development of a shear zone lead to a general conceptual picture of the stages of slope deformation, illustrated in Figure 4. Slope movements prior to failure occur through deformation throughout a rock or soil mass; movements post failure – after development of a deep-seated landslide - occur primarily by displacement across a shear zone. Hence, landslide behavior is governed by the geometry and properties of the shear zone. This is key to anticipating landslide behavior. Location and size of the shear zone is greatly influenced by preexisting heterogeneities and discontinuities, particularly for landslides in rock (Stead and Wolter, 2015) and residual soils formed of weathered bedrock (Lambe, 1996). Resistance to movement across the shear zone is controlled by friction, and frictional resistance is reduced by pore pressures.

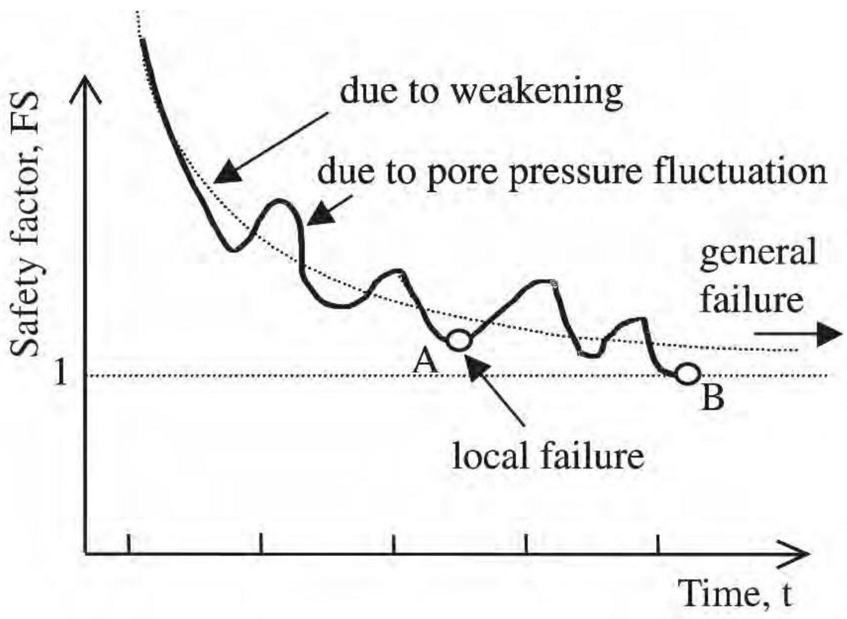



**Figure 3. Progressive failure of a slope.**

Here the safety factor (FS, also called the Factor of Safety), shown on the vertical axis, indicates the ratio of forces acting to move material downslope to those acting to hold it in place. Values greater than one indicate stability. Over time, crack growth and weathering weaken material within the slope, reducing shearing resistance. This causes a gradual reduction in stability, shown by the dashed line. Overprinted on that trend are periodic changes in pore pressures, that also reduce resisting forces, and changes in forces directed downslope, such as seismic shaking. At some point, these fluctuations can cause local zones of failure within the slope and, eventually, complete failure manifest by formation of a through-going shear zone. This figure is from Picarelli et al. (2004), who were examining failure in clay-rich, weathered shale slopes; similar processes are inferred to occur in rock slopes (e.g., Petley and Allison, 1997).

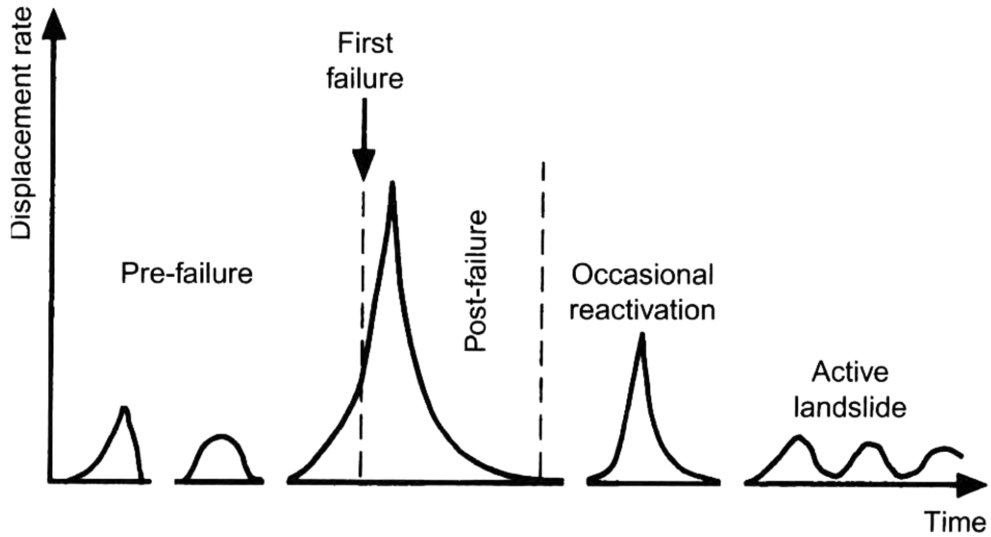



**Figure 4. Stages of slope movement.**
Prior to first failure and development of a shear zone, slope movements occur through creep. Failure is preceded by a period of accelerating creep. Once a shear zone has formed, slope displacement is accommodated primarily by ductile deformation within the shear zone and is controlled primarily by pore pressures at the shear zone. Movement may cease with falling pore pressures or if a landslide adjusts to a more stable geometry. Likewise, movement may reinitiate when pore pressures rise sufficiently. From Leroueil (2001).

## 6.4 Types of landslides

Geologists have devised classification systems to categorize landslides in terms of movement type and materials involved. Table 1 in Chapter 16 of the Washington Forest Practices Board Manual, for example, lists 19 categories; Hungr et al. (2014) identify 32 different landslide types (and provide detailed descriptions of each). Regmi et al. (2015) provide a detailed description of landslide types and processes. In addition to landslide type, it is also instructive to classify landslides in terms of rate of movement. Cruden and Varnes (1996) define seven categories that span six orders of magnitude in velocity.

Although deep-seated landslides span a broad range of types and rates of movement, the landscape features generally associated with deep-seated landsliding, and those most commonly dealt with in the context of forest practices, are formed by development of a thin shear zone, with future landslide movement governed by friction across the shear zone. The characteristics and behavior of any individual landslide are governed by the geometry of that shear zone, determined primarily by the geometry of subsurface fractures and heterogeneities along which it forms (e.g., Badger, 2002; Stead and Wolter, 2015); by fluctuations in pore-water pressure at the shear zone, which alter resistance to movement; by changes in strength of the shear zone as it shears

(Iverson, 2005), and by the degree to which material above the shear zone breaks apart and loses strength as it moves (Hungr et al., 2005).

# 7 Creation and Evolution of Deep-Seated Landslide Features

Deep-seated landslides include a large range of landslide types, involving different materials and rates of movement, with a diverse array of features formed through various histories of landslide activity and evolution (Terzaghi, 1950). Despite this diversity, the mechanisms by which movement occurs creates commonalities across certain landslide types that can be used to understand and anticipate landslide behavior.

## 7.1 Shear zone properties

A deep-seated landslide is created by movement across a shear zone, and subsequent movement occurs primarily across that same shear zone. Development of a shear zone is both the final step in landslide initiation and the determining factor for future landslide behavior. Properties of the shear zone are key to that behavior.

### *7.1.1 Residual strength*

As described previously in the description of brittle failure, a shear zone develops through the growth and coalescence of cracks within soil and rock. Prior to crack development, soil and rock masses resist applied stresses through elastic (fully recoverable) deformation of the mineral grains and any cementing matrix providing cohesive bonds between particles (Phase I of Figure 2). Prior to failure, the strength of intact material determines stability of a slope.

Once cracks form and coalesce into a shear zone, rock and soil resist applied stresses through friction across particle contacts within the shear zone. Deformation occurs through permanent (non-recoverable) sliding across this zone. As shown in Figure 2, the peak stress that can be supported by intact material (point c in Figure 2) is greater than the stress required to drive sliding across the shear zone (point d in Figure 2). Once a shear zone has developed, a rock or soil mass exhibits a residual strength that is less than its intact strength prior to failure (Chen and Liu, 2014; Skempton, 1985).

The shear zone thus provides a weak boundary that mechanically isolates a landslide from adjacent intact material and acts to perpetuate movement within the landslide body (Baum and Reid, 2000).

### *7.1.2 Ductile behavior*

First-time landsliding of an intact slope often involves a sudden decrease of rock or soil strength when cracks coalesce into a shear zone (point c in Figure 2). Subsequent movement of the landslide involves sliding across an existing shear zone (point d in Figure 2). The shear zone exhibits ductile behavior: sliding begins when shear stresses exceed the shear strength of the shear zone, and persists until stresses fall below that level (e.g., Iverson and Major, 1987). The rate of sliding is proportional to the ratio of shear stress to shear strength (e.g., Wong et al., 1995).

### 7.1.3 Effective stress

Frictional resistance to sliding determines shear strength and is proportional to the force driving particles together. Movement across the shear zone occurs when gravitational or earthquake-generated shear forces exceed that frictional resistance. The force driving particles together is proportional to the weight of overlying rock and soil. This weight also determines the shear stresses acting to drive movement across the shear zone.

Water that fills pore spaces within rock and soil exerts pressure that reduces the stress driving particles together, thereby reducing shear strength. The effective stress that determines frictional resistance is thus equal to the weight of overlying soil and rock minus the pore pressure. Materials within a deep-seated landslide shear zone typically have no cohesion, so shear strength can be represented mathematically as

$$\text{Shear strength} = (\sigma\text{-}u)*\tan(\phi),$$

where $\sigma$ is component of gravitation stress normal to the shear zone, u is pore-water pressure, and $\phi$ is the angle of internal friction of material in the shear zone – a measure of its intrinsic frictional resistance (Terzaghi, 1950).

When pore spaces are saturated, water pressure supports the weight of overlying water. Water pressure is thus proportional to the depth of the saturated zone. If pore spaces are not fully saturated, water tension with soil and rock particles supports the weight of the water. Water in the unsaturated zone adds to the weight of soil and rock.

### 7.1.4 Low permeability

Material within the shear zone typically has lower permeability than the overlying material composing the body of the landslide. In such cases, water infiltrating the body of the landslide would tend to pool above the shear zone, creating a saturated layer perched within the landslide body; which, in effect, causes “the slide to fill with water like a bathtub” (Baum and Reid, 2000; Baum et al., 2003). Deep-seated landslides form a leaky “bathtub”, because groundwater can seep downward out of the body through the shear zone (unless the shear zone intersects a deeper groundwater zone, in which case water may seep upward into the landslide) and water drains from the landslide where the water table intersects the surface.

Formation of such a perched, shallow aquifer then influences pore pressures within the shear zone. As depth of the aquifer increases, pore pressures at the shear zone increase, effective stress decreases, and shear resistance of the shear zone is reduced. For many monitored landslides, initiation of movement occurs when pore pressures at the shear zone meet some threshold value. Factors that influence the formation, depth, and persistence of this aquifer profoundly influence the behavior of a landslide. Such factors include the rate and processes by which water is supplied to a landslide, and the rate and processes by which water drains from the landslide. Increases in the amount or rate of water supply may increase groundwater levels; increases in the amount or rate of drainage from the landslide, such as development of incised channels that limit the height of the water table within the landslide body, may reduce groundwater levels.

Pore pressures in the shear zone can also be influenced by confined pressurized aquifers beneath the shear zone (Badger et al., 2011) and through the process of shearing, if drainage is impeded and undrained conditions develop.

#### *7.1.5 Strain softening, strain hardening*

When sliding begins, particles within the shear zone shift position relative to each other. If these shifts enlarge pore spaces, the material dilates. If these shifts reduce pore spaces, the material contracts. Dense materials tend to dilate upon shearing, while loose materials tend to contract. A change in the size of pore spaces within the shear zone will generate a corresponding change in the water pressure within those pore spaces: enlarged pore space lowers pore-water pressure; reduced pore space increases pore-water pressure.

A reduction in pore pressure increases effective stress with a corresponding increase in frictional resistance. If the shear zone dilates upon shearing, its strength increases – it experiences strain hardening. An increase in pore pressure decreases effective stress with a corresponding decrease in shear resistance. If the shear zone contracts upon shearing, its strength decreases – it experiences strain softening. These properties of the shear zone have a profound influence on landslide behavior. Dilation of the shear zone promotes slow and discontinuous landslide movement; contraction of the shear zone causes a positive feedback that promotes runaway acceleration (Iverson, 2005).

If the shear zone dilates, reduction of pore pressures will cause water from surrounding material to flow into the shear zone. If it contracts, increase pore pressures will cause water to flow out of the shear zone. Shear zone permeability can be comparatively low relative to the landslide body above, so the rate of water flow into or out of the shear zone may be slowed, and shear-initiated changes in pore pressure can persist.

#### *7.1.6 Strength recovery*

Field and experimental evidence show that clay-rich, shear-zone materials may regain strength during periods of inactivity under low confining stresses (e.g., Angeli et al., 2016; Hussain and Stark, 2011). Experiments indicate that recovery increases with time and can reach a substantial proportion, up to 70%, of the pre-failure soil strength (Stark and Hussain, 2010). However, strength recovery occurs only under relatively small confining pressure, corresponding to depths less than about five meters, and therefore may apply primarily to shallow landslides or shallow portions of deep-seated landslides.

### 7.2 Landslide Body

The body of a landslide encompasses a portion of a previously intact slope, so it starts initially having the strength and permeability of that intact material. That material tends to break up as it moves downslope, reducing its overall strength and altering its permeability. These changes have profound influences on landslide behavior. The degree of break-up depends on the amount of deformation required to slide over the shear zone, which may have bends and kinks that compress or extend over-riding material, and on the distance traveled downslope. Thus, landslide characteristics and factors controlling landslide behavior may change both over time with landslide activity and with position along the axis of the landslide.

### *7.2.1 Fracture induced permeability*

Landslides in cohesive materials tend to involve initial movement of blocks of material, which fracture as they move downslope. Fractures and tension cracks within the landslide body open pathways for water and increase hydraulic conductivity within the landslide. This network of fractures and tension cracks is sometimes referred to as macro porosity. The location, orientation, and abundance of fractures developed within the body of a landslide (Figure 5) thus influence the rate at which water infiltrates from the surface, the rate at which saturated zones form and expand as water infiltrates, and the rate at which groundwater and pore-pressure fluctuations propagate through the landslide body (e.g., Debieche et al., 2012; Krzeminska, 2012; Malet et al., 2005; Proffer, 1992). In-situ permeability tests at La' Cita in northern Italy, a landslide formed in heavily tectonized marine sedimentary rocks, found that hydraulic conductivity within the landslide body is an order of magnitude greater than that within adjacent fractured bedrock (Cervi et al., 2012; Ronchetti et al., 2009).

Fracture density also determines the amount of water that can be stored within a given volume of rock. Geochemical analyses at La Clapiére landslide (Binet et al., 2007a), located in igneous and metamorphic crystalline rocks of south east France, find that specific yield[11] within the landslide body is two orders of magnitude greater than in the surrounding bedrock.

These and similar observations at other landslides show that material within a landslide body tends to be more heavily fractured than surrounding material, with correspondingly greater porosity and permeability, and with associated greater specific yield (water storage per unit volume) and hydraulic conductivity. Pore-pressure responses to precipitation are therefore different inside and outside of a landslide. Upslope of the La' Cita landslide, Cervi et al. (2012) observed large pore-pressured increases in response to precipitation, with a lag of about a week. Within the upper portion of the landslide, they observed large seasonal variations in pore pressure, but almost no response to precipitation events. They attributed this lack of response to higher specific yield and conductivity within the landslide body. Larger specific yield – larger storage volume – generates a smaller increase in groundwater level, because the same volume of water can be stored with a smaller increase in water level. Higher conductivity allows incoming water to drain rapidly downslope. This difference in response, however, varies widely from site to site. At a landslide in marine sediments in California, Proffer (1992) found large and rapid groundwater-level response to precipitation within the landslide, while areas outside of the landslide had very gradual groundwater-level responses, with a lag time of several months. The important point is that groundwater levels within a landslide can respond to temporal patterns of precipitation differently than those outside of the landslide.

[11] Specific yield is the volume of water released, per unit area, for a unit depth decrease of the water table. It is a measure of the amount of water stored per unit volume of rock.

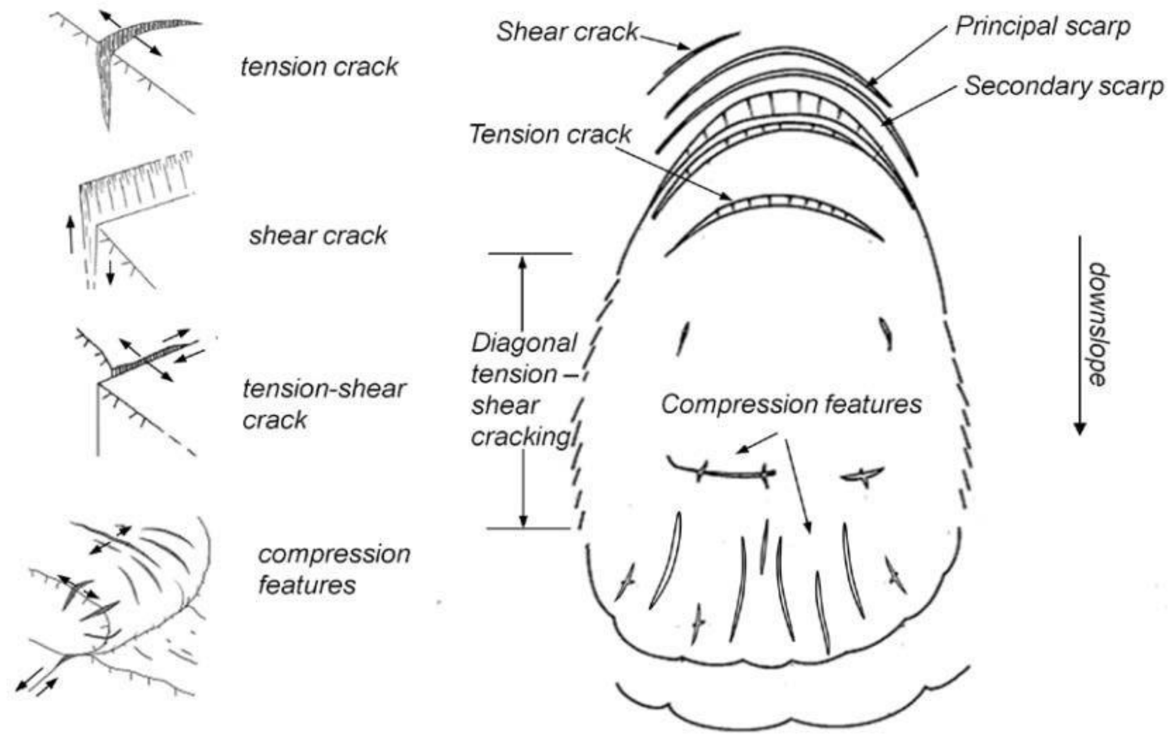



**Figure 5. Cracks form as a landslide moves downslope and spreads laterally.**

These cracks provide pathways for infiltration of water (Keaton and DeGraff, 1996; Krzeminska, 2012; Stumpf et al., 2013).

Fractures provide the pathways for downslope flow of water and corresponding propagation of increased pore pressure. The Johnson Creek Landslide, located in sedimentary rocks on the west coast of Oregon, provides an example (Priest et al., 2008). Rainwater infiltrated the soil overlying the landslide body at a rate of about 5 cm/hr. The water table (top of the saturated zone) was closest to the ground surface at a graben (down-dropped block) formed at the head (top) of the landslide, so it was there that infiltrating water first created a rise in groundwater-level associated pore pressures. Piezometer arrays installed over the body of the landslide show that the pore-pressure rise initiated at the head of the landslide then propagated laterally through saturated material above the shear zone. This pressure wave traveled at 140-250 cm/hr through the upper portion of the slide, and increased to 350 cm/hr and greater through the middle portion of the slide. These high rates of pressure transmission through the body of the landslide, arriving downslope well before infiltrating water from above, were attributed to fracture-induced high effective conductivity.

### *7.2.2 Growth of fractures with landslide displacement*

Monitoring at Johnson Creek landslide also demonstrated that pore-pressure responses to rainfall vary over time, attributed to changes in fracture patterns that occur as the landslide moves and deforms. Temporal variability in the response of groundwater levels within a landslide body is found in many active landslides where monitoring instrumentation has been installed. At the Rosone landslide in Italy, for example, Binet et al. (2007b) found that the increase in groundwater level associated with a given infiltrated volume of water became significantly smaller after a period of landslide activity. They attributed this change to increased hydraulic conductivity associated with new fractures formed by deformation of the landslide body.

Geochemical analyses provide further evidence of fracture growth associated with landslide activity. For example, in examining the geochemistry of spring water emanating from slopes in two Alpine valleys in gneiss, Binet et al. (2009) found that water draining areas with active slope movements were enriched in sulfates derived from oxidative dissolution of pyrite on newly exposed crack and fracture surfaces.

### *7.2.3 Downslope evolution, weathering*

Many deep-seated landslides involve rotational or translational failure of relatively intact blocks. Those blocks then move gradually downslope, breaking apart as they move. The degree of disintegration can increase with the distance moved, so the blockiness and texture of landslide debris may vary with distance along the axis of the landslide.

Fracturing caused by movement and associated deformation of landslide debris provides access of water to fresh surfaces, facilitating weathering and consequent clay-enrichment of the landslide debris. Disintegration and weathering of the debris tends to reduce porosity and permeability, causing the debris to become less blocky and more fine grained. Landslides that involve gradual downslope movement of debris thus often have intact blocks near the head and more disintegrated debris toward the bottom.

This downslope evolution creates a downslope variation in material properties. For example, in examining large rock-slide/earth-flow landslides in marine sedimentary rocks of northern Italy, Ronchetti et al. (2010) report an order-of-magnitude decrease in average hydraulic conductivity from the head to the toe of the landslides. These changes result in associated differences in pore-pressure responses to precipitation. In the head, groundwater levels rise in response to precipitation with a lag time of 1-8 days and seasonal variation of about two meters. In the toe, groundwater response to precipitation is muted, with seasonal variations of less than a meter.

### *7.2.4 Up and down-slope expansion*

Landslide extent can grow upslope (retrogress) over time by new failures into intact material at the head of the landslide. Such failures typically involve blocks of material that slide downward along steeply dipping shear zones, which may be curved so that the block rotates. When a block fails, lateral support is removed from intact material above and stresses within the slope change, potentially creating conditions for progressive development of a new shear zone for the next block in line. The body of many landslides are composed of a series of these blocks, each moving gradually downslope and in progressively greater states of disintegration. In some cases, the blocks disintegrate upon failure, creating a rock or debris avalanche, as occurred at the head scarp of the Red Bluff landslide (Cascade Landslide Complex, Columbia River gorge) in 2008

(Randall, 2012). Deposition of debris from these headscarp failures onto existing landslide debris can reactivate movement within the landslide body (Bertolini, 2010).

Landslides can also extend downslope as material in the toe of the landslide moves, overriding whatever is not pushed out of the way. Rivers occupy valley floors, so the toes of many landslides abut river banks. Material at the deep-seated landslide toe may then fail via periodic shallow landslides, the river removes failed material, and the process continues conveyer-belt like as the landslide moves downslope.

### *7.2.5 Compound landslides*

Disaggregation of landslide debris as it moves downslope, together with changes in topography generated by landslide movement or erosion by streams, offer opportunities for development of additional shear zones and formation of landslides within landslides. Cronin (1992) describes why these secondary landslides may exhibit substantially different behavior than the host landslide and why they may be substantially more sensitive to periods of high precipitation and other factors that can trigger landslide movement.

As discussed previously, presence of a low-permeability shear zone overlain by highly fractured and variably weathered debris can generate an isolated aquifer within the body of a landslide, with groundwater flow patterns controlled by geometry of the shear zone and by the degree of fracturing and weathering within the landslide body. Landslide behavior is strongly influenced by the depth and persistence of the aquifer that forms above the low-permeability shear zone. Behavior of a secondary landslide is influenced by the geometry of its shear zone and material properties of its overlying debris, which may result in substantially different responses to precipitation, stream erosion, seismic shaking, or whatever perturbations that trigger landslide movement than exhibited by the host landslide.

The host landslide has a groundwater system influenced by its shear zone, the infiltration and preferential flow pathways, and hydraulic conductivity of its body. Secondary deep-seated landslides thus form within a groundwater environment greatly influenced by the host landslide. By creating isolated aquifers within the host landslide body, the secondary landslides likewise can alter the groundwater environment of the host landslide. Substantial interaction may occur between the two, which are probably site and event specific. Cronin (1992) provides the only example found in the literature that explicitly examines the implications of these factors for stability of compound landslides, and he concludes that secondary landslides are generally less stable than the host landslide.

The presence of multiple, overlying shear zones provides opportunities for movement on one to trigger movement on another. The Nile Valley landslide along the Naches River in Washington in 2009, for example, involved initial movement of landslide debris over a shallower shear zone (20 to 35 m depth) and subsequent movement over a deeper (60 to 85 m depth) shear zone within an underlying bedrock sequence (Badger and Smith, 2010; Badger et al., 2011).

Development of secondary shear surfaces is an important process in re-activation and downslope displacement of large earthflows in northern Italy (Bertolini, 2010). In these cases, retrogressive rock avalanches/flows at the headscarp add weight to landslide debris at the head of the earthflow. This compresses the debris and causes a transient increase in pore pressures within the debris (undrained loading, Hutchinson and Bhandari, 1971). The rise in pore pressure triggers localized downslope movement across the shear zone. Inertia prevents complete reactivation of the earthflow; rather, the localized movement is transferred upward into the body of the earthflow along a new or existing shear zone to thrust material upward and over downslope debris. This loads the downslope debris, triggering a similar response, which may be repeated sequentially downslope, creating a series of imbricate thrusts that may ultimately extend throughout the length of the earth flow (Figure 6).

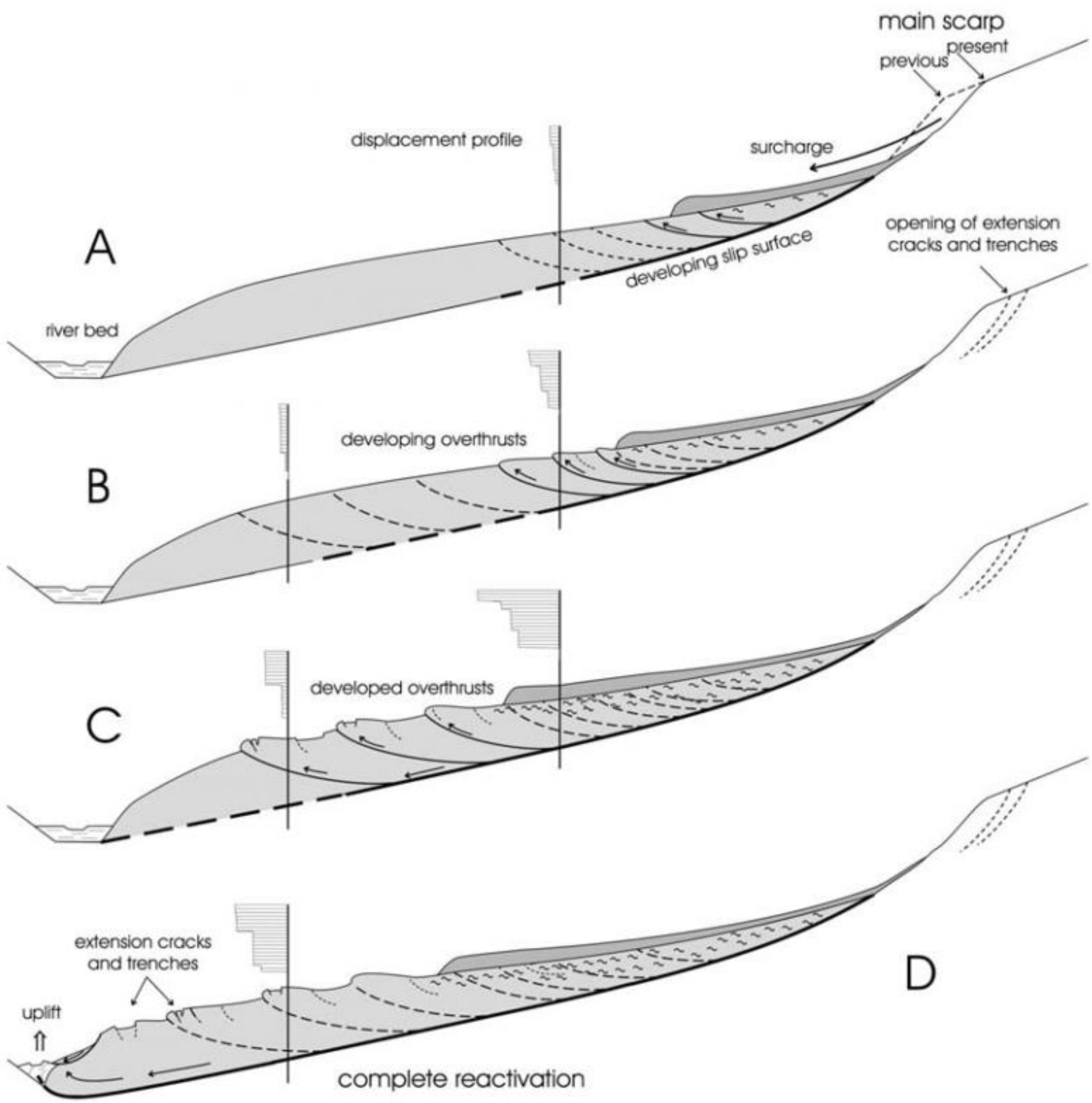



**Figure 6. Progressive downslope compound landslide development by undrained loading.**
From Bertolini (2010).

#### *7.2.6 Implications for hazard assessment*

Changes in specific yield and hydraulic conductivity, both by opening of fractures during periods of activity and healing of fractures (through deposition of precipitates and weathering products) during periods of inactivity, alter the pore-pressure response to precipitation, and alter thresholds for initiation of landslide movement.

# 8 Water

Water contained within the body of a landslide has four potential sources:

1) infiltration of precipitation onto the landslide,
2) surface runoff from areas upslope (e.g., streams, gullies) draining into the landslide,
3) subsurface runoff from areas upslope draining into the landslide through the head and lateral scarps, and
4) groundwater seeping through the shear zone from below.

Groundwater levels within the landslide aquifer(s) and landslide response to precipitation depend on the relative amount and timing of water flow from each of these sources. Groundwater levels also depend on the amount and timing of water outflows from the landslide body. Outflows occur through

1) seepage downward through the shear zone, if material below is unsaturated.
2) seepage to the surface to feed overland and channelized flow draining the landslide, and
3) evapotranspiration of soil water.

## 8.1 Landslide water budget

We have found few studies that attempted to quantify the relative contribution from each source for water inflow to a landslide. Proffer (1992) developed a water budget for a landslide in southern California. She used five years of monitoring data that spanned years with average precipitation and years with double the average. Her results are shown in Figure 7.

In this case, runoff from areas upslope provided a substantial portion of the water inflow to the landslide. Loss of water from the landslide occurred by seeps at the toe, which accounted for 81% of the water inflow, and by evapotranspiration and downward seepage through the shear zone. None of the water inflow was attributed to upward seepage of deeper groundwater through the shear zone.

In contrast, in constructing a water budget based on geochemical analyses for water draining the Ca’ Lita landslide in northern Italy, Cervi et al. (2012) attributed 64% of the water in the shallow landslide aquifer to seepage from deep groundwater.

Pore pressures at the shear zone exert a primary control on deep-seated landslide behavior. These pore pressures are a function of the quantity and flow patterns of groundwater within the landslide. The quantity and timing of water flow into the landslide influence the quantity and flow patterns of groundwater. As shown in the examples above, the quantity of water from each of the four sources of water to a landslide can vary substantially. The timing of water influx also

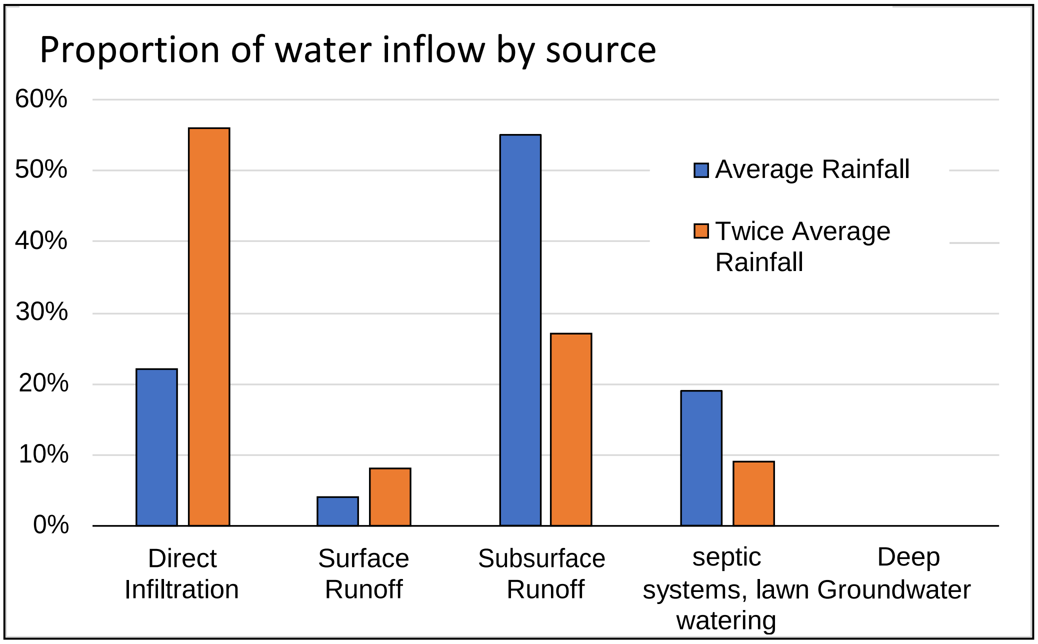



**Figure 7. Landslide water budget.**

Abalone Cove Landslide, Southern California. From Proffer (1992).

differs for each source. Three of the four sources, surface and subsurface runoff from upslope and groundwater inflow from below, originate from precipitation and snow melt outside the perimeter of the landslide. The extent of this area, its size relative to the size of the landslide, and the proportion of water falling on this area that flows as surface runoff, as subsurface runoff, and as groundwater to the landslide are all important controls on landslide behavior. These factors are unique for each landslide, but the physical factors that govern these processes can be observed and potentially characterized to provide quantitative, or at least qualitative, estimates of where the water controlling landslide behavior comes from.

### 8.2 Runoff versus persistent groundwater

Runoff involves the relatively rapid (hours to days) drainage of water from hillslopes to streams during and following rainfall and snow melt events. Rapid drainage occurs through both overland flow and subsurface flow. Forest soils tend to have high infiltration capacity and hydraulic conductivity[12], so precipitation readily infiltrates the ground surface, except in areas lacking soil (rock outcrops, road surfaces), or where soil has been compacted (Harr, 1977). Subsurface flow occurs in both unsaturated and saturated pore spaces. Flow through unsaturated zones occurs along a film of water coating soil particles and crack surfaces. In unsaturated zones, water movement is primarily downward. Rapid drainage by saturated flow, also called storm

[12] Harr (1977) measured saturated hydraulic conductivities for a forest soil in the H.J. Andrews Experimental forest in western Oregon that ranged from 412cm/hr near the surface (30cm depth) to 22cm/hr at 150 cm depth.

flow, interflow, and throughflow, occurs when transient zones of saturated soil or fractured rock form during and after precipitation events (Figure 8). These zones form where soil or rock permeability decreases with depth and the rate of infiltrating water seeping downward exceeds the infiltration capacity. In these saturated zones, groundwater can move laterally and flows downslope roughly parallel to the ground surface. The depth below the ground surface at which such transient saturated zones form during and after rainfall events depends on the variation of hydraulic conductivity with depth.

Although permeability generally decreases with depth, it does not go to zero, and water continues to seep downward to recharge a perennial zone of saturation at deeper depths. This is the persistent groundwater zone (Figure 8).

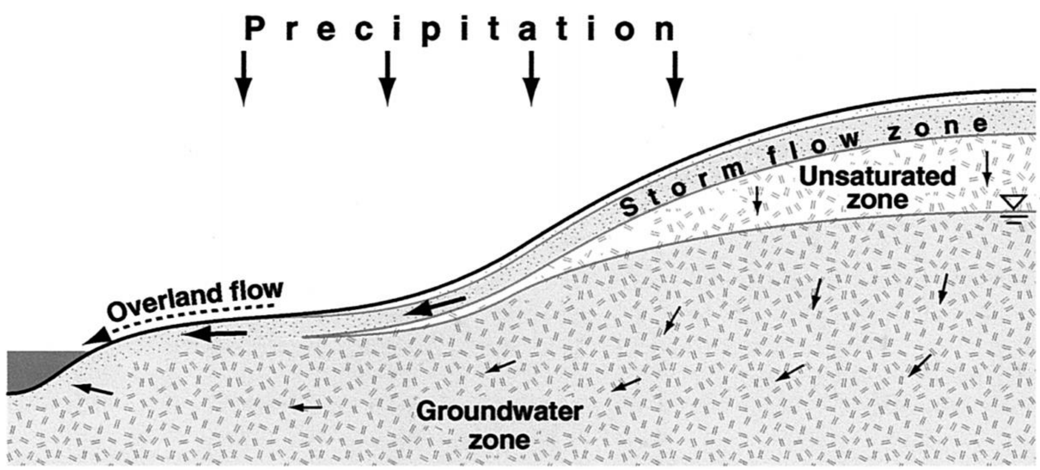



**Figure 8. Water flow pathways.**
Gray coloring indicates saturation. From Scanlon et al. (2000).

Storm runoff has traditionally been associated with shallow subsurface flow through saturated zones that form above the soil-bedrock interface. Many studies, however, find that a substantial portion of storm runoff can occur within the bedrock (Figure 9), where highly fractured zones near the surface provide conduits for rapid transit of water downslope (Gabrielli et al., 2012; Kosugi et al., 2011; Montgomery et al., 2002; Padilla et al., 2014). These highly fractured zones may extend from a few meters to tens of meters below the ground surface.

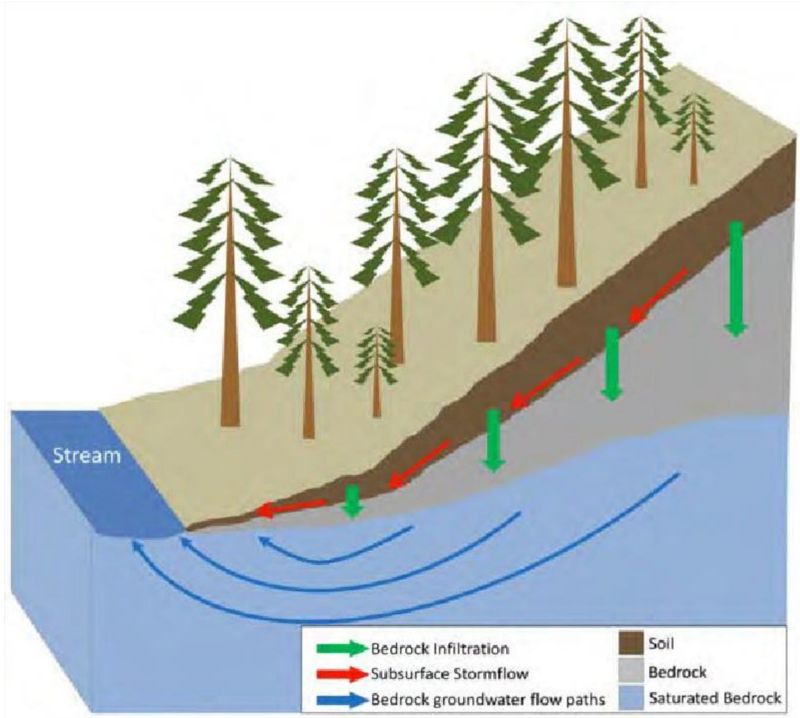
Groundwater in both the transient and persistent groundwater zones flows from areas of high elevation to areas of low elevation. Where the top of the saturated zone – the water table – intersects the ground surface, water seeps out to form overland flow and to feed stream flow. Summer base flow in streams is maintained by outflow from the deep groundwater zone.

Transient and persistent groundwater are both recharged by infiltrating rainwater or snowmelt. Depth to the water table for both zones thus changes over time and with location as the amount of infiltrating water varies. The response of groundwater to precipitation events generally differs for these two zones. Water must traverse the unsaturated zone before contributing to groundwater, so there is a lag in response to precipitation that increases with increasing depth to the water table. Soil and fractured rock porosity and permeability are high at shallow depths, so groundwater here can drain downslope relatively quickly. Porosity and permeability tend to decrease with depth, so a given amount of water saturates a greater volume of material and drainage to areas downslope occurs more slowly. Hence, at shallow depths, groundwater levels tend to rise and fall with each precipitation event; at deeper depths, groundwater levels tend to rise and fall slowly, responding to the cumulative rainfall over multiple events.

Details of this conceptual model are described in all hydrology textbooks. A good synopsis can be found with the online course available from Utah State University at http://hydrology.usu.edu/RRP/.

Water flow from the surface through the unsaturated zone can be bypassed by fissures and other preferential flow paths, causing a more rapid response of groundwater at depth (e.g., Bogaard and Greco, 2016; Krzeminska et al., 2012; Shao et al., 2016). Steeply dipping opening-mode fractures, as formed under surface-parallel tension induced by extension, can provide a direct pathway for water inflow to bedrock. Padilla et al. (2014), for example, observed a large (8.5m) and rapid groundwater response to precipitation in the upper ten meters of bedrock, whereas the response in overlying soil was much less (0.7m). The soil and the shallow bedrock contained independent saturated zones. That in the soil was fed by direct downward infiltration of water seeping through the soil; that in the bedrock was fed by water draining into steeply dipping extensional fractures upslope that then drained into a heavily fractured zone of bedrock (Figure 10). Similar responses have been observed in monitored slopes in Washington at Snoqualmie Pass and at Aberdeen Bluffs (T. Badger, personal communication).

A deep-seated landslide can be placed in the context of this conceptual model. A low-permeability shear zone, or shear zones in a compound landslide, further hinder downward seepage of infiltrating rainfall or snowmelt and thus promote formation of a persistent saturated zone within the landslide body (Baum and Reid, 2000). Shallow subsurface flow can enter the landslide along its upslope boundary, and further contribute to groundwater within the landslide. If deep groundwater intersects the shear zone, water can seep up through the shear zone into the landslide body.



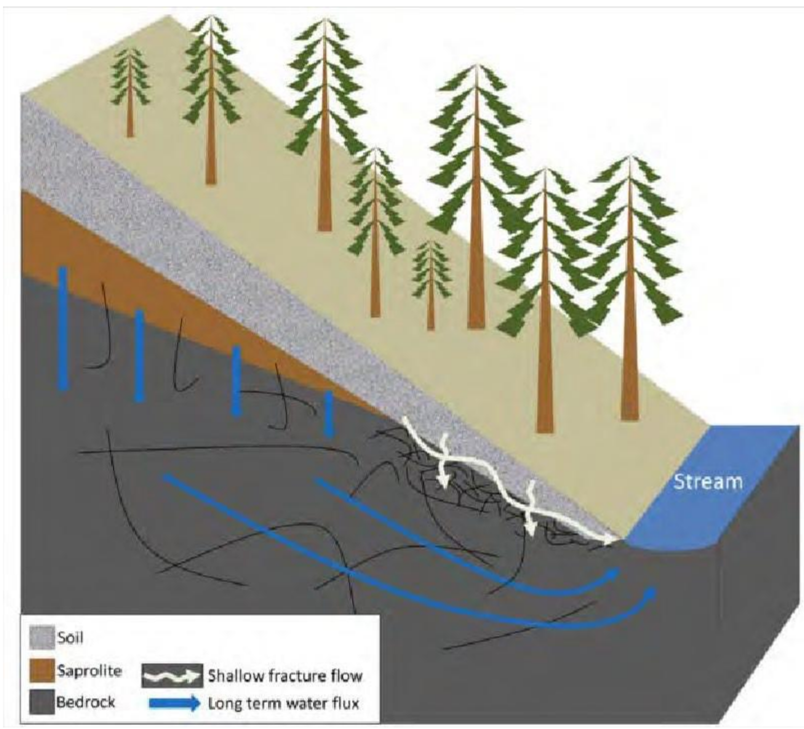




**Figure 9. Conceptual models of bedrock flow paths.**
Based on observations in New Zealand (upper) and Oregon (lower). In the New Zealand case, storm flow develops in the soil above the bedrock and infiltration through the soil recharges a deep, bedrock aquifer. In the Oregon case, infiltration through the soil feeds stormflow through a shallow, highly fractured zone in the bedrock and recharges a deeper bedrock aquifer. Gabrielli et al. (2012).

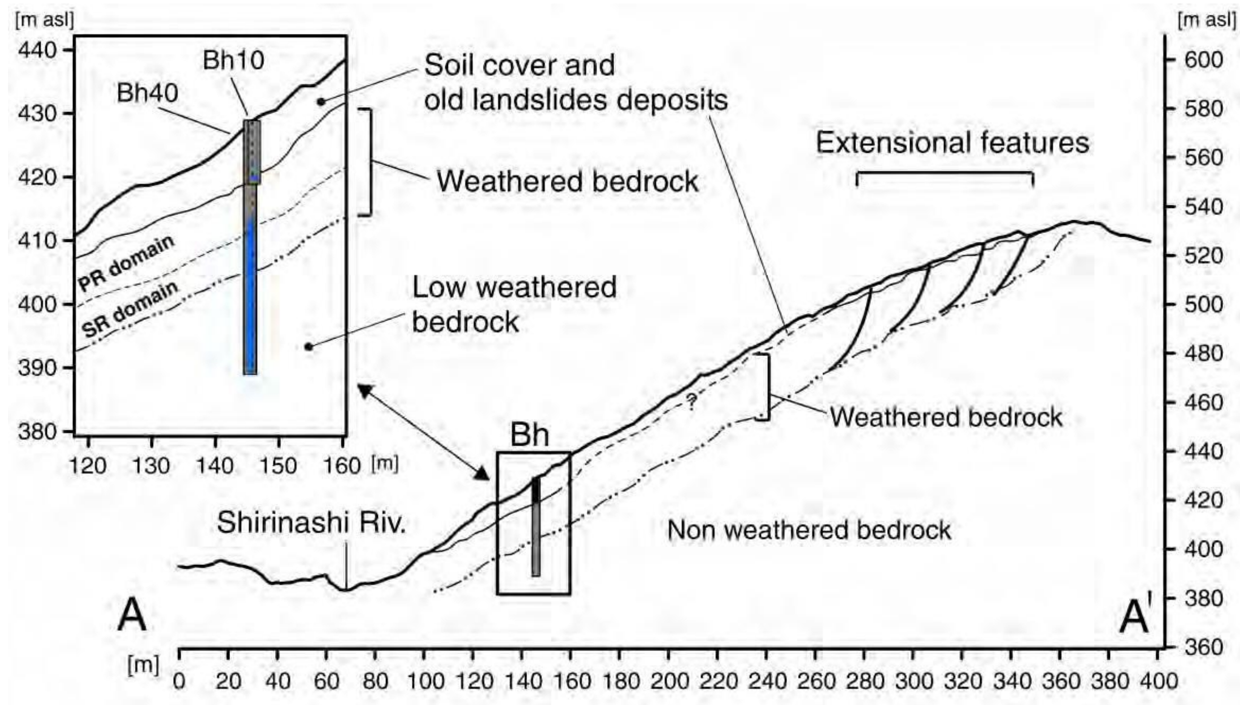



**Figure 10. Bedrock aquifer fed by upslope fractures.**

Shallow (Bh10) and deep (Bh40) boreholes installed in a hillslope in Japan revealed a perched aquifer above the soil-bedrock interface and a lower, independent bedrock aquifer. The perched soil aquifer was recharged by downward infiltration of rainwater through the soil; the upper portion of the bedrock aquifer was recharged by water draining into extensional features upslope that directed water into a highly fractured zone that allowed rapid drainage of water downslope. From Padilla et al. (2014).

### 8.3 Fractures

Fractures provide the pore spaces and connections for groundwater storage and movement in bedrock. Factors that influence patterns of bedrock fracturing will likewise influence patterns of groundwater flow.

Folding of rock layers under tectonic forcing creates localized fracture sets oriented relative to the axis of the fold (Singhal and Gupta, 2010), such as opening mode fractures that form parallel to the axis of an anticline (Badger, 2002). Geologic structure plays an important role in determining water-flow pathways through bedrock. However, many investigations examining hydrologic properties of bedrock find a highly fractured zone that extends from the bedrock surface downwards for several meters to tens of meters. These fractured zones are not necessarily related to geology, but rather to near-surface weathering processes. For bedrock, the topography itself imposes a primary control on these weathering processes.

As discussed in Section 6.1, topographic perturbations of gravitational and regional tectonic stresses can be of sufficient magnitude to fracture rock (Miller and Dunne, 1996). These topographic stresses are consistent with observed near-surface zones of high fracture density (Figure 11) and with fracture patterns associated with morphology referred to as sackung (Ambrosi and Crosta, 2006; Kinakin and Stead, 2005; Pánek et al., 2015) and deep-seated gravitational slope deformation (Jaboyedoff et al., 2013). These morphologies are found in mountainous terrain throughout the world (Crosta et al., 2013; Pánek et al., 2015; Tsou et al., 2015; Varnes et al., 1989), including the Olympics and Cascades (Beget, 1985; Bovis and Evans, 1996; Tabor, 1971; Thorsen, 1989b).

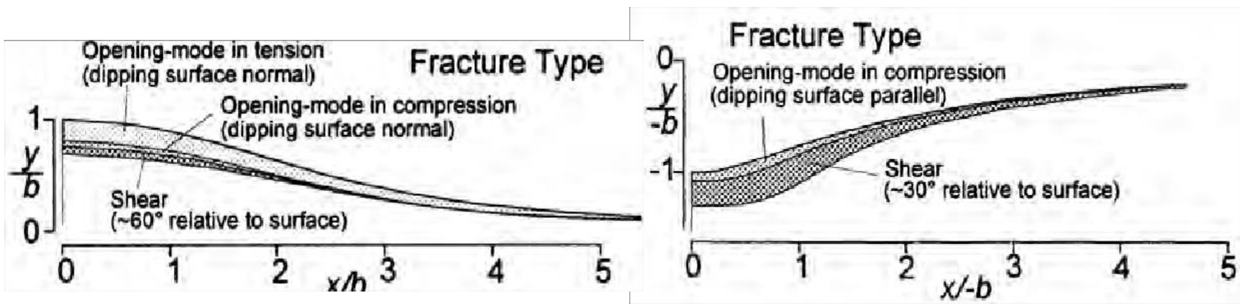



**Figure 11. Fracture type and orientation induced by topographic stresses.**

Fracture type and orientation consistent with stresses in an isolated ridge (left) and valley (right) under regional compression. Horizontal and vertical distances (x and y) are normalized by the relief (b) of the ridge or valley. From Miller and Dunne (1996).

Topographically induced fractures provide conduits for infiltration from the surface and create near-surface porosity and permeability that, as described above, exert a primary control on runoff and groundwater flow processes. Studies seeking to identify sources of water at landslide sites in the Alps (Binet et al., 2007b; Guglielmi et al., 2002; Pisani et al., 2010; Vallet et al., 2015a) identify shallow aquifers formed in bedrock through zones of "unloading" and "decompression" characterized by contour-parallel extension features (fissures, scarps, trenches). These aquifers may be transient, with some portion of water lost to seepage downward into less fractured, lower permeability bedrock below, and some portion flowing laterally downslope through the aquifer, in some cases to drain to deep-seated landslides formed on the lower portion of the slopes.

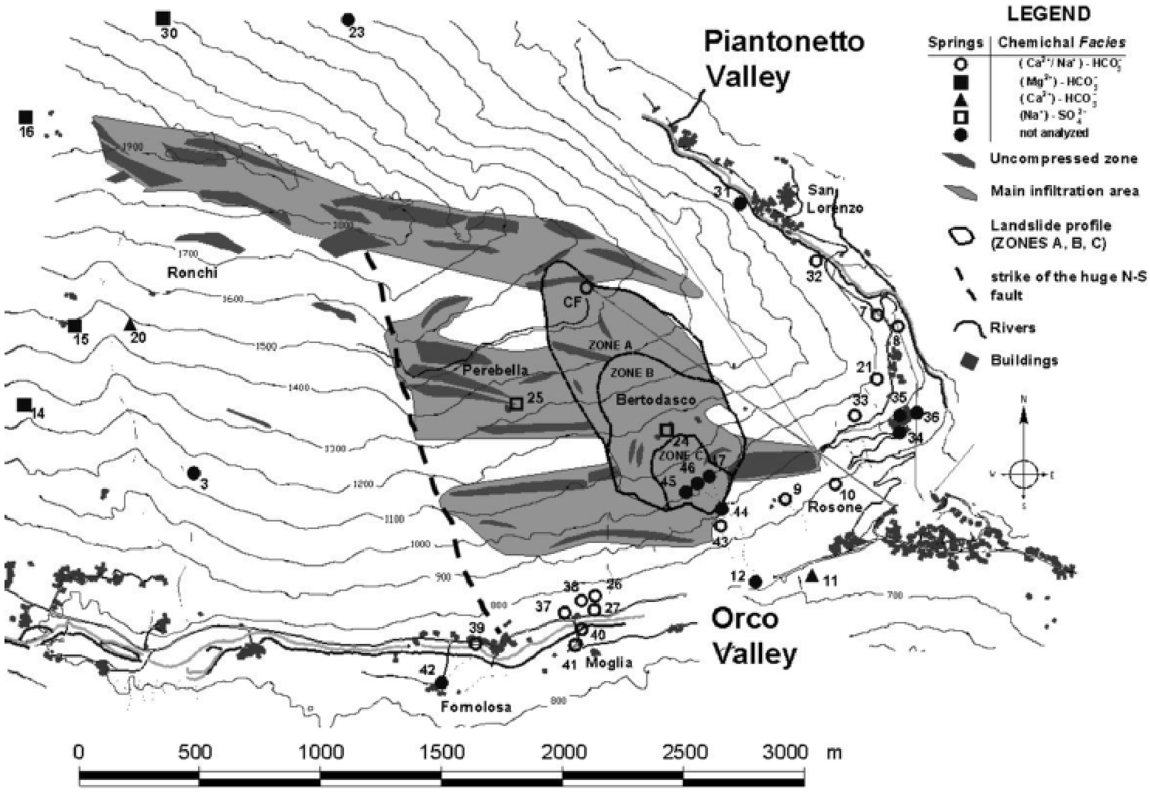



**Figure 12. Fracture-induced recharge area.**
Recharge areas for springs on the Rosone slope in Italy. Lighter shaded zones indicate areas of infiltration. Infiltration extending west along the ridge top flows through a shallow, perched aquifer to the landslide, indicated by the thick black lines. Binet et al. (2007b) and Pisani et al. (2010).

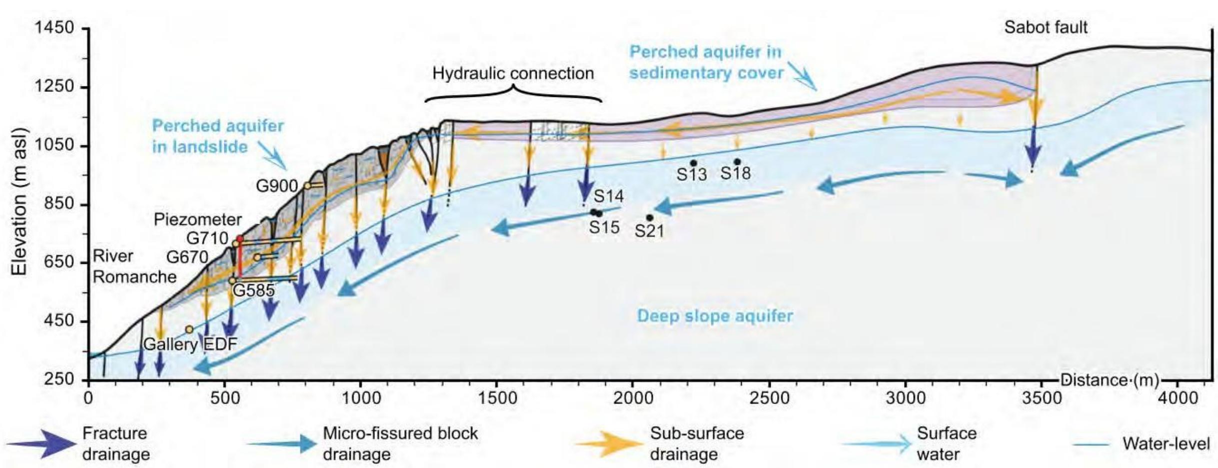



**Figure 13. Perched aquifer with lateral flow to landslide.**
A cross section through the Séchilienne landslide in France, showing formation of a shallow perched aquifer with lateral flow to the landslide. From Vallet et al. (2015a).

## 8.4 Deep groundwater

We have used the term "deep groundwater" to refer to a zone of perennial saturation to differentiate it from transient zones that form within soil and near-surface highly fractured bedrock. Deep groundwater is not always deep; the water table may intersect the ground surface to form seeps and springs, typically near the base of slopes and at valley floors, where it provides base flow to streams. Groundwater can also flow within confined aquifers, where a higher-permeability layer is overlain by a lower-permeability layer – an aquitard – so that water within the more permeable material is pressurized. When a well is bored into a confined aquifer, the water level will rise above the confining layer.

Several studies document interactions of deep groundwater with deep-seated landslides. This may occur either where the water table of an unconfined aquifer intercepts the landslide shear zone, or where a landslide shear zone cuts into or is affected by pore pressure from an underlying confined aquifer (Badger et al., 2011; Cervi et al., 2012; Ronchetti et al., 2009). In either case, if pore pressures within the underlying confined aquifer exceed those in the shear zone, seepage will occur upward into the shear zone and landslide stability will be further reduced (Hodge and Freeze, 1977; Ronchetti et al., 2009).

Pore pressure at a point within a deep groundwater aquifer will tend to increase as recharge to the aquifer increases. The extent of the recharge area to an aquifer is determined by regional patterns of groundwater flow. These patterns cannot be observed directly, and so are inferred from indirect evidence and numerical modeling.

Several types of observations provide indirect evidence of groundwater flow patterns. Measurements of water table level or water pressures (pressure head) obtained from arrays of monitoring wells or piezometers are used to infer water flow directions – water flows from high to low pressure head. The abundance of heavy isotopes of oxygen ($\delta^{18}O$) and hydrogen ($\delta^{2}H$) are observed to vary with elevation, so the abundance of these isotopes in sampled groundwater or spring water can provide an estimate of the elevation from which the water originated. The geochemistry of sampled water provides clues as to the rock types traversed. Chemical tracers are used to identify recharge locations for sampled water. Geophysical methods can be used to infer subsurface water content.

None of these methods, however, provide a complete picture of actual flow paths. Complete two- and three-dimensional estimates of groundwater flow patterns can be obtained using computer modeling. The indicators listed above can be used to calibrate and validate groundwater models for specific sites (e.g., Smerdon et al., 2009; Tiedeman et al., 1997). Even without such data, however, groundwater models can provide insight about how geologic structure and topography influence groundwater flow and about how changes in recharge might influence groundwater levels.

Tóth (2009) and numerous references therein explore influences of structure and topography on groundwater flow patterns at length, but primarily in a two-dimensional, cross-sectional context. Welch and Allen (2014) look specifically at how patterns of surface fracturing affect modeled flow paths, also in a cross-sectional view. Welch and Allen (2012) and Welch et al. (2012) use two- and three-dimensional models to examine flow paths for both generic and actual basin topographies. These and other studies highlight the following points:

- Groundwater flow is topographically driven, with recharge occurring at higher elevations and discharge at lower elevations. However, groundwater and surface drainage divides do not necessarily coincide (Winter et al., 2003, discuss this issue at length). Within a mountain front or valley side, recharge at high elevations may follow a deep flow path to the valley floor and bypass smaller tributary basins. This is illustrated in cross section in Figure 14.

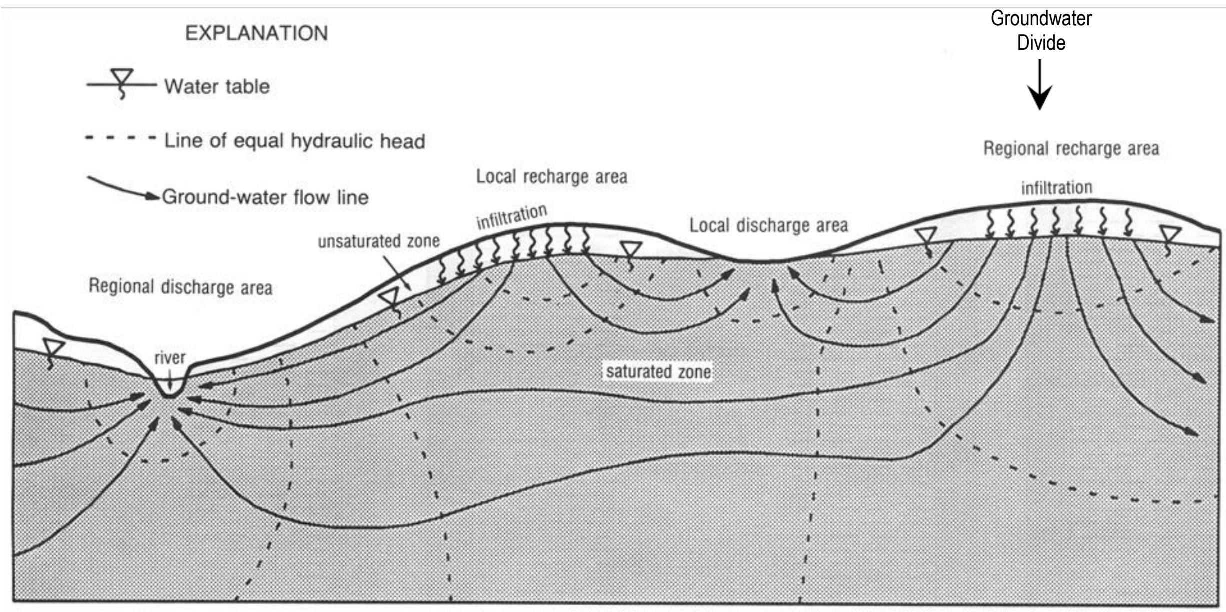



**Figure 14**. Conceptual cross section showing nested patterns of groundwater flow

Figure 15, from a study by Tiedeman et al. (1998), illustrates this in plain view for groundwater recharge to a lake in New Hampshire. Based on well data and numeric flow modeling, the recharge area for groundwater flow to the lake extends beyond the surface drainage basin. To further illustrate, Figure 16, from 3-D groundwater modeling by Welch and Allen (2012) for a basin in south eastern British Columbia, shows flow paths for groundwater discharge points along a mainstem river (Mission Creek, red lines), a tributary channel to the mainstem (Daves Creek, green lines), and from low-order channels that flow to the tributary channel (blue lines). Recharge for groundwater flow to Mission Creek extends to the upper-most elevations in the Daves Creek basin with flow paths that pass underneath Daves Creek and its tributaries. Likewise, flow paths for groundwater feeding Daves Creek flows underneath its tributary basins. Groundwater recharge to the base of a slope near the valley floor, where deep-seated landslides tend to be found, can extend to the ridge top, even though the surface drainage may be intercepted by smaller basins and their channels that drain the valley slopes.

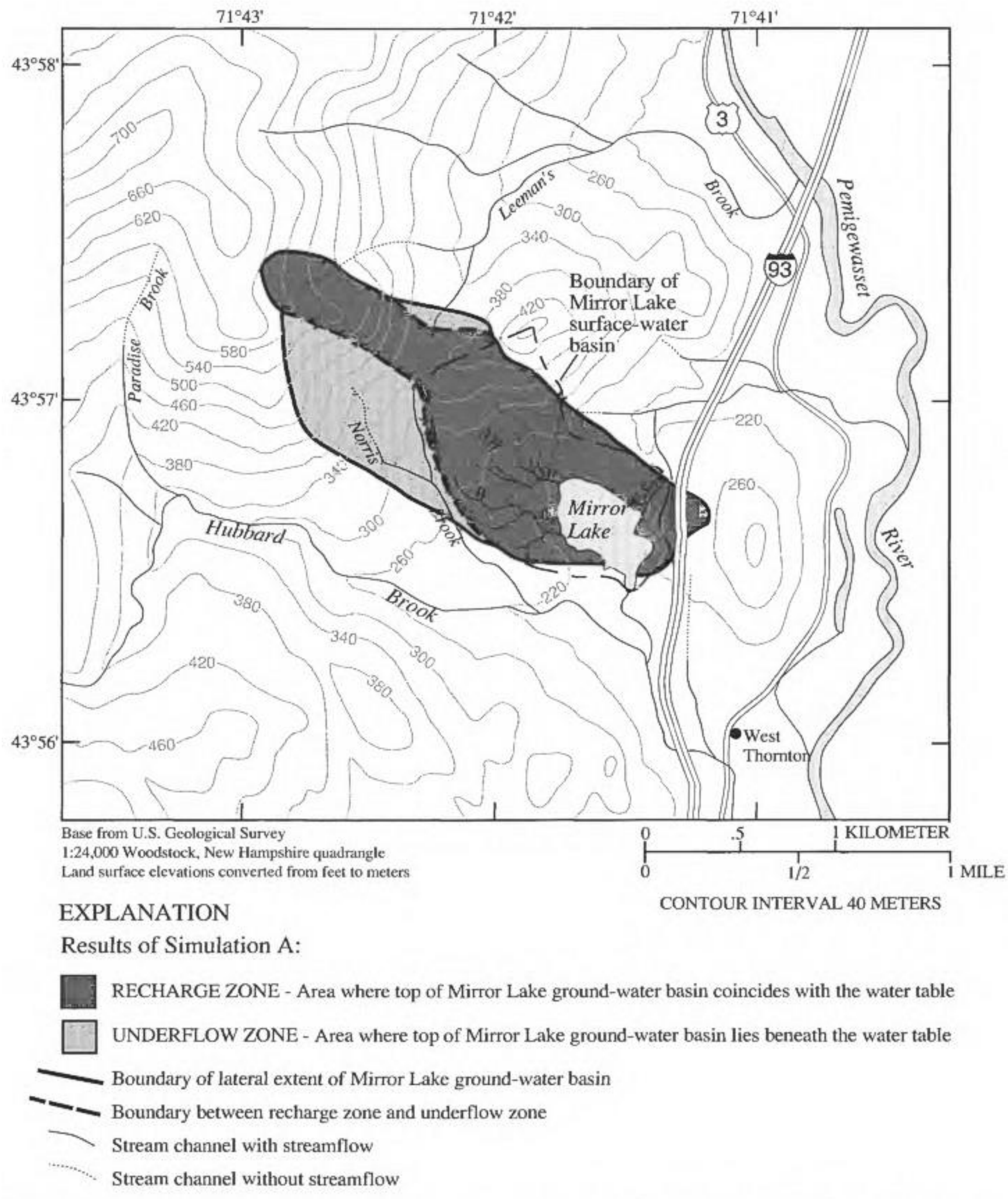



**Figure 15. Recharge area.**
The groundwater basin extends well beyond the surface-drainage basin for this lake in New Hampshire. The underflow zone indicates a nested flow pattern, where deeper flow paths bypass the Norris Creek drainage. From Tiedeman et al. (1997).

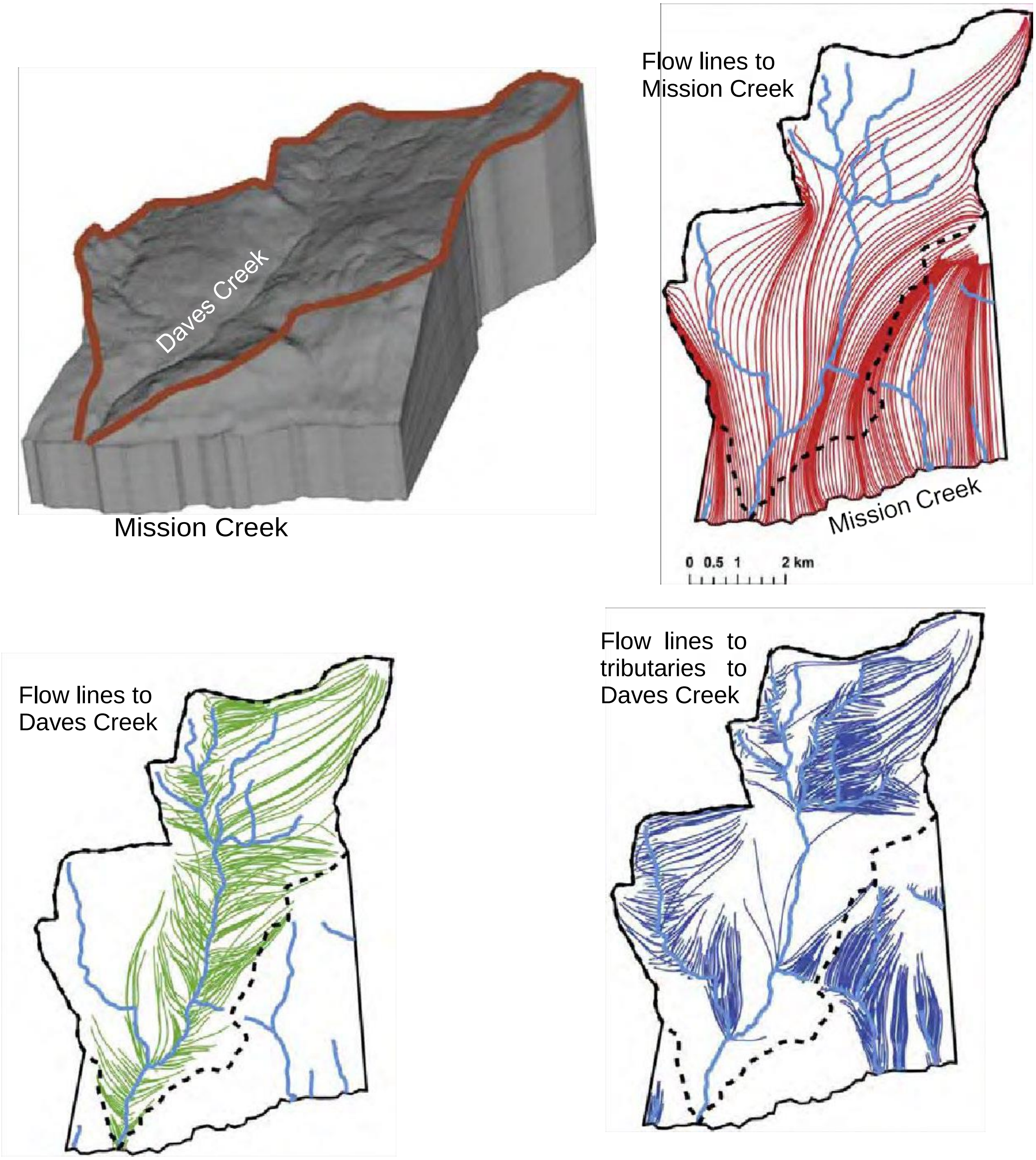



**Figure 16. Modeled groundwater flow lines.**

In the upper right, red lines show flow lines to Mission Creek; in the lower left, green show flow lines to Daves Creek, and in lower right, blue show flow lines to low-order channels tributary to Daves Creek. The flow paths shown for Mission Creek flow underneath those to Daves Creek and its tributaries. Likewise, flow paths to Daves Creek flow underneath those to the tributary channels.

- Recharge-zone extent can change as recharge rate varies. As recharge rate decreases, groundwater levels fall and local groundwater flow within smaller, tributary basins is reduced. If the water table falls below the level of the low-order channels, they dry up. Welch et al. (2012) illustrate this phenomenon for Daves Creek, as shown in Figure 17 copied from their paper.

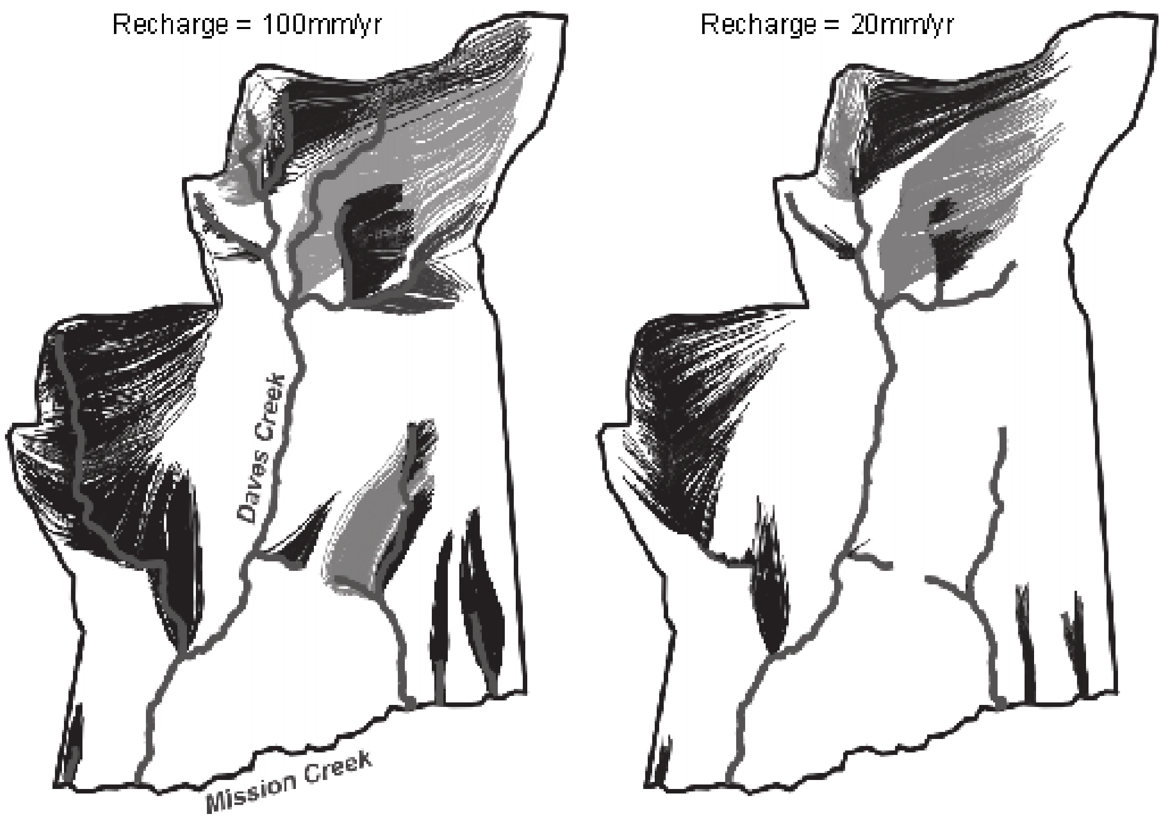



**Figure 17. Variation in recharge alters recharge area.**

The black and gray lines indicate groundwater flow paths to 1st-order channels. As recharge rate is reduced, modeled groundwater levels fall and the area providing recharge to these channels contracts. In some cases, the water table falls entirely below the channel elevations, as seen for channels in the southeast portion of the modeled area. From Welch et al. (2012).

- Changes in the spatial pattern of recharge can change groundwater flow patterns. For example, tree plantations are used in some areas to draw down the water table; a water-table mound forms under areas with no forest cover (Figure 18). Similarly, water tables may become elevated under areas where trees are harvested, thus causing the groundwater divide to shift and the recharge area supplying groundwater to a site downslope to expand.

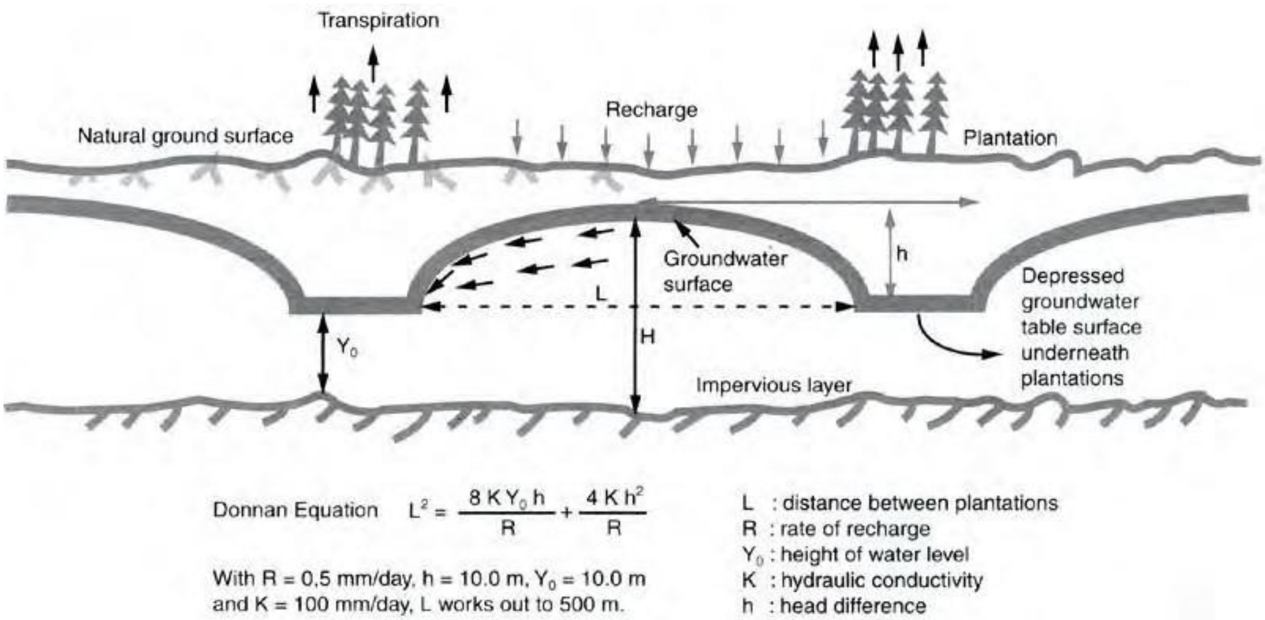



**Figure 18. Use of plantations to draw down the water table.**

The Donnan Equation, generally used to determine spacing between surface drains to keep the groundwater mound below a certain height, is used here to determine spacing between row plantings of trees. From Heuperman et al. (2002).

### 8.5 Methods to delineate source areas of water

To reiterate, there are four primary pathways for water influx to a deep-seated landslide:

1) infiltration of precipitation onto the landslide,
2) surface runoff from areas upslope,
3) shallow subsurface runoff from areas upslope, and
4) groundwater seeping through the shear zone from below.

Surface observations can be used to delineate the first two. The landslide boundary defines the source area for direct precipitation. The topographic surface drainage divide defines the potential source area for surface runoff; mapping of channel courses and field evidence of overland flow can provide unambiguous evidence of the area providing surface runoff to a landslide.

Source areas for shallow subsurface stormflow and deeper groundwater flow are more difficult to delineate. As described above, flow pathways provided by bedrock fractures, nested groundwater flow systems, and temporal or spatial variability in recharge rate can cause groundwater divides to differ from surface drainage divides, so that source areas for subsurface stormflow and groundwater may differ substantially from surface drainage basins. A variety of methods are used to infer sources of subsurface water.

### *8.5.1 Field mapping and remote sensing of extensional features*

Surface exposure of extensional features provide pathways for rapid infiltration of surface water to near-surface, highly fractured bedrock (e.g., Figures 7 and 8 above). Field mapping of extensional features, such as fissures, scarps, and depressions, can thus provide indicators of upslope source areas for subsurface storm flow and deeper groundwater recharge (Guglielmi et al., 2005; Padilla et al., 2014; Pisani et al., 2010).

Extensional features can also be mapped using remotely sensed data. Fissures, scarps, and trenches form linear features – lineaments – that can be mapped from satellite or optical imagery and digital elevation data (Scheiber et al., 2015; Šilhavý et al., 2016). Lineament density (lineament length per unit area) has been used to map groundwater recharge potential (e.g., Mallast et al., 2011; Shaban et al., 2005; Yeh et al., 2008) and in landslide hazard assessment (e.g., Ramli et al., 2010; Simon et al., 2014).

Closed depressions on ridgetops and upper slopes may also indicate extensional features, such as a down-dropped block (graben) created upslope of a laterally sliding large landslide block. Closed depressions can be mapped from high-resolution digital elevation data, such as LiDAR bare-earth DEMs (e.g., http://www.netmaptools.org/Pages/NetMapHelp/3_10_closed_depressions.htm).

### *8.5.2 Stable isotope tracers*

Water contains a small proportion of stable isotopes of hydrogen and oxygen. Deuterium, $^{2}H$, contains a neutron not found in hydrogen ($^{1}H$) and $^{18}O$ contains two more neutrons than the much more abundant $^{16}O$. Water consists of two atoms of hydrogen and one of oxygen. The most abundant form is $^{1}H_2{}^{16}O$; and there are two other stable molecules: $^{1}H^{2}H^{16}O$ and $^{1}H_2{}^{18}O$. These isotopes, because of the extra neutrons, have higher atomic masses than regular water, so that when water evaporates, the relative abundance of the isotopes in the water vapor is slightly depleted, and when water condenses, the relative abundance of the isotopes increases, relative to the vapor. Isotope composition of water is expressed as the ratio of the heavy to the light isotopes relative to a standard set by the International Atomic Energy Agency:

$$\delta^{18}O = ((^{18}O/^{16}O)_{Sample}/(^{18}O/^{16}O)_{Standard} - 1)*1000.$$

Every body of water: lakes, rivers, groundwater, has a unique isotopic composition ($\delta^{18}O$ value) reflecting its source and process of formation. The relative abundance of the isotopes changes only during changes in phase (evaporation, condensation, melting) and remains constant once water has infiltrated the surface to recharge groundwater. Thus, these isotopes can be used to identify groundwater sources (McGuire and McDonnell, 2007).

For example, Peng et al. (2007) used $\delta^{18}O$ values to determine that groundwater flowing to an unstable slope in Taiwan came primarily from seepage of upslope farm ponds, and not from infiltrating rain water (see Peng et al., 2011; Peng et al., 2010 for further examples).

The absolute change in isotope abundance is strongly dependent on temperature, and temperature decreases with increasing elevation. Hence, rainwater becomes increasingly depleted in $^{18}O$ with increasing elevation (McGuire and McDonnell, 2007) and regional relationships can be established between the $\delta^{18}O$ precipitation value and elevation. McGuire et al. (2005), for example, found variations in $\delta^{18}O$ of -0.22‰ to -0.26‰ per 100 meters elevation gain for the Oregon Cascades.

Several studies have used the elevation relationship for $\delta^{18}O$ in rainwater to estimate the elevation of the recharge zone for groundwater flowing to landslides. In the Italian Alps, Binet et al. (2007b) measured a -0.26‰ per 100m elevation gain $\delta^{18}O$ precipitation gradient. With that information, they inferred a mean elevation for recharge of water sampled at springs draining the Rosone landslide of 1800m, at the elevation of tensional features considerably upslope of the landslide (see Figure 9). Guglielmi et al. (2002) found non-linear $\delta^{18}O$ elevation relationships for precipitation in the southern French Alps. In examining spring and groundwater water sampled at the La Clapière and Sèchilienne landslides, they found that the area providing groundwater extends to elevations above the landslides and varies seasonally, with spring snowmelt contributing infiltration at progressively higher elevations. Vallet et al. (2015a) further examined $\delta^{18}O$ relationships for springs near the Sèchilienne landslide, along with geochemical and tracer tests, to delineate topographically controlled and geologic-structure-controlled groundwater flow paths.

### *8.5.3 Geochemical water balance*

The chemistry of groundwater evolves as it flows through soil and rock. Certain chemical components dissolve from the soil and rock surfaces in contact with water; others may be deposited from the water onto those surfaces. The chemical composition of groundwater thus reflects the initial composition of the rain or snow melt that infiltrated the surface, the chemical and mineral composition of the soil and rock surfaces encountered during transit, and the time spent in contact with those surfaces. Water chemistry thus provides clues as to the water origin and travel path.

The chemical composition of water sampled from springs and wells has been used to identify recharge areas and infer the proportion of water from different sources for several large landslides in Europe. Guglielmi et al. (2002) used a combination of stable isotope tracers to estimate groundwater recharge elevations and major ion components of the sampled water to infer flow paths to identify two aquifer sources for water to the La Clapière and Séchilienne landslides in southern France: a shallow, perched aquifer with lateral flow extending to the summit, and a deep aquifer intercepting the landslides near the foot of the slope. Vallet et al. (2015a) expanded the analysis for the Séchilienne landslide, with similar results, using additional monitoring data. Binet et al. (2007a) use the same methodology to identify a similar two-aquifer system for the Rosone landslide in Italy. Ronchetti et al. (2009) use geochemical analyses to infer a groundwater source for the Ca' Lita landslide in Italy, later confirmed by Cervi et al. (2012) using a water balance for the landslide based on groundwater chemistry. Montety et al. (2007) use geochemical analyses water sampled from the Super-Sauze earthflow in France to infer a primary groundwater source for portions of the earthflow.

### *8.5.4 Numerical (computer based) groundwater models*

Groundwater recharge areas can be inferred using groundwater models, as described for studies by Tiedeman et al. (1998) in New Hampshire (Figure 15) and Smerdon et al. (2009) for a basin in British Columbia. Groundwater models require specification of the surface topography, the spatial and temporal distribution of recharge, and the subsurface distribution of hydraulic conductivity and water storage capacity. Detailed groundwater models are thus coupled with spatially distributed soil water-balance models, as done by Smerdon et al. (2009), and parameterized with abundant bore-hole data and in-situ well tests.

Groundwater models can be applied without detailed site-specific data. Welch and Allen (2012, 2014) and Welch et al. (2012), for example, use numeric models to gain insights about topographic (Figure 16 and Figure 17) and structural controls on groundwater flow patterns. Models can be used in this way to test assumptions and pose hypotheses; to assess the validity of using surface drainage divides to infer groundwater recharge zones, for example.

## 8.6 Implications for hazard assessment

Conceptually, we can view a deep-seated landslide as a semi-independent groundwater system, isolated to some degree from adjacent areas by its shear zone. For most landslides, the basal shear zone acts as a partial barrier (an aquitard) to groundwater flow. As described by Baum and Reid (2000), we can think of a landslide as a bathtub[13]. It fills with water from the four sources listed previously: direct infiltration, surface runoff, subsurface runoff, and inflowing groundwater. It drains from seepage downward through the shear zone and from channels, springs, and seeps where the water table within the landslide body intercepts the ground surface. Groundwater levels rise when the rate of inflow exceeds the rate of outflow, and the landslide moves when ground water levels generate pore pressures at the shear zone sufficient to overcome frictional resistance across the shear zone. The magnitude of pore pressure required to trigger movement depends on a numerous factors, including the geometry of the landslide body, and other processes may trigger movement, such as loss of buttressing at the landslide toe, but variations in the rate and timing of water inflows can be an important factor driving the initiation and cessation of landslide movement.

If outflows of water from the landslide body occur at lower rates than inflows, groundwater within a landslide can integrate variable inputs over long time periods. The time taken for groundwater levels within a landslide to respond to changing inputs will depend on the size of the groundwater reservoir (the aquifer) within the landslide, on the relative rates of inputs and outputs, and on the rate of water flux through the landslide body; this response time may span

[13] Baum and Reid (2000) were referring to active earthflows. We've expanded this bath tub analogy to include all deep-seated landslides that involve movement over a shear zone and to include inactive (dormant and relict) landslides under the assumption that properties of the shear zone persist over long periods of inactivity. This assumption is supported by lack of cohesion in a sampled shear-zone soil from an ancient landslide in China (Chen et al., 2014) and laboratory work showing lack of strength regain in clay soils at deeper depths (Stark and Hussien, 2010), but the chemical and mechanical evolution of shear-zone materials over long periods of inactivity has not been widely explored in landslide studies (Bromhead, 2004).

weeks for a small landslide and years for a large landslide. It is worthwhile to consider each source of inflow.

*Direct infiltration*. All water that infiltrates directly from the surface into the body of a landslide adds to the water budget of the landslide. Therefore, any changes in landcover over a landslide that alter the soil-water balance will alter the landslide-water balance. Evapotranspiration from forest cover accounts for a substantial portion of the annual precipitation for Pacific Northwest forests (see discussion and citations in the glacial-deep-seated landslide literature review), so timber harvest or other factors that reduce forest cover within the boundary of a deep-seated landslide will increase water influx to the landslide.

*Surface runoff*. When rainfall intensity exceeds soil infiltration capacity, surface runoff is generated by infiltration-excess overland flow. In the Pacific Northwest, forest soils have relatively large infiltration capacities relative to rainfall intensities (e.g., Harr, 1977), so that infiltration-excess overland flow occurs primarily over rock outcrops and road surfaces and is not generally an important process for generating surface runoff, except where severe fires have reduced soil infiltration capacity (Wondzell and King, 2003).

Stream channels that flow from upslope can also contribute surface runoff to a landslide. Streams flowing across a landslide body may contribute water to the landslide where the water table is lower than the base of the stream (a losing stream), and may drain water from the landslide where the base of the stream intersects the water table (a gaining stream). Water table levels can vary over time, so that some portion of a stream may be losing water to the landslide at sometimes and gaining water from the landslide at other times. If a stream flowing onto a landslide is losing water, activities within the area contributing flow to the channel can increase water inflow to the landslide.

*Subsurface runoff*. As found for the Abalone Cove Landslide in California (Figure 5), subsurface runoff can provide a substantial portion of the water inflow to a deep-seated landslide. Subsurface runoff occurs primarily at shallow depths, so the source area for subsurface runoff will generally match that for surface runoff: it can extend to the topographically defined surface drainage divide. However, as found for the Rosone landslide in the Italian Alps (Figure 10), near-surface bedrock fractures can route shallow subsurface flow laterally from beyond surface divides (Binet et al., 2007b).

Numerous studies document increased peak stream flows following timber harvest, indicating that loss of forest cover increases storm runoff. Increased runoff reflects the increase in antecedent soil moisture content caused by reduced evapotranspiration.

*Groundwater*. Landslide investigations in Italy (Cervi et al., 2012; Ronchetti et al., 2009) and here in Washington (Badger and Smith, 2010; Badger et al., 2011) have encountered groundwater below landslide shear zones with water pressure exceeding the hydrostatic pressure exerted by groundwater levels within the landslide body. In these cases, the shear zone may have cut through a low-permeability layer into or experience pore pressures from an underlying confined aquifer, or may act as a confining layer to upwelling groundwater. In either case, seepage into the shear zone increases pore pressures, adds water to the landslide body, and can reduce landslide stability. Increased groundwater flux to the landslide will further increase pore pressures at the shear zone, and so it is plausible that landcover changes within the groundwater recharge area can affect landslide stability.

*Outflow*. Factors that alter the rate of groundwater outflow from a landslide can also influence groundwater levels and affect landslide stability. For example, development of incised channels over the body of a landslide may facilitate more rapid drainage by intersecting the water table.

# 9 Reactivation

## 9.1 Triggers

### *9.1.1 Pore pressure*

Many studies of reactivated landslides identify rising groundwater levels with associated increased pore pressures at the landslide shear zone as the primary suspect in triggering renewed activity. In these cases, reactivation is associated with large precipitation events or with extended periods of above-average precipitation.

Antecedent moisture and rainfall intensity play an important role in shallow landslide initiation, so that rainfall thresholds for the onset of landsliding can be identified for some areas (e.g., Caine, 1980; Dahal and Hasegawa, 2008; Godt et al., 2006). Similarly, rainfall thresholds have been identified for the onset of movement for deep-seated landslides, although there are few examples in the published literature. Hong et al. (2005), for example, found a threshold in the average intensity and duration of storms for initiating movement of four large bedrock landslides in Japan (Figure 19). Their results suggest that some rough regional threshold may apply, but detailed examination of individual landslides shows that each has a unique threshold (Floris and Bozzano, 2008) and that these thresholds may involve complex temporal rainfall sequences and must account for evapotranspiration (Vallet et al., 2016).

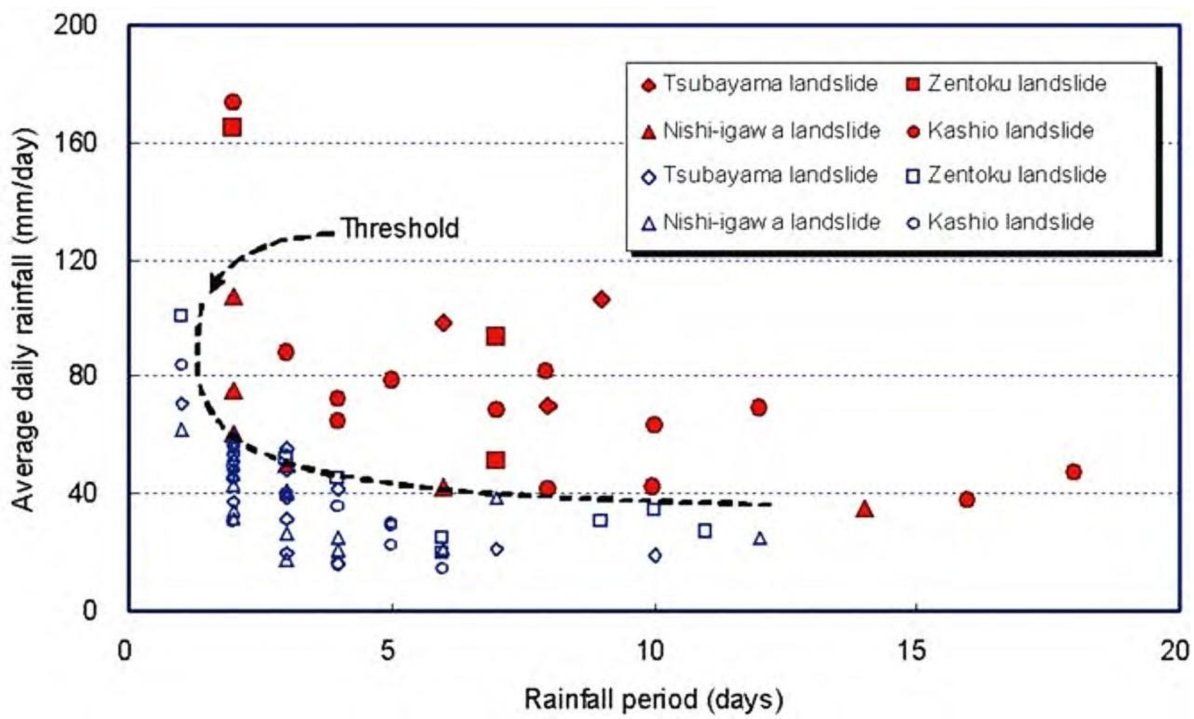



**Figure 19. Rainfall threshold.**
Rainfall threshold for initiation of movement for four deep-seated, bedrock landslides in Japan. Red symbols indicate rainfall events associated with onset of landslide movement; blue symbols indicate events for which no movement occurred. From Hong et al. (2005)

### *9.1.2 Undrained loading.*

One landslide can trigger movement of another when rock or debris avalanches from the headscarp deposit material onto older landslide debris at the foot of the scarp. The added weight of the newly deposited material compresses the debris and this compression can increase pore pressures within the landslide body, particularly if the debris is relatively impermeable so that the increased pore pressures cannot rapidly dissipate. These compression-induced pore pressures can exceed those that would occur solely by full saturation of the landslide debris (Hutchinson and Bhandari, 1971), and can trigger landslide movement. Hutchinson and Bhandari (1971) invoke this mechanism of undrained loading to explain observed movement over shear surfaces of low slope where static stability analyses indicate landslide stability, even when fullysaturated.

Undrained loading triggered by rock avalanches at the headscarp is thought to reactivate earthflows in the northern Apennines of Italy. The initial event triggers a series of imbricate thrusts that progresses downslope (Figure 6). Each thrust loads a downslope portion of the earthflow, which then initiates further undrained loading and triggers the next thrust (Bertolini, 2010; Bertolini and Pizziolo, 2008).

Exceptional runout distances for some landslides is also attributed to undrained loading from material that falls onto saturated landslide debris and causes liquefaction (Geertsema et al., 2006; Geertsema and Schwab, 2006).

#### *9.1.3 Toe erosion:*

Changes in geometry of a landslide body or intact slope alter the balance of forces and can reduce stability. River erosion (Massey et al., 2016) or excavation (Stark et al., 2005a) can reactivate existing landslides and trigger new ones.

Toe erosion may not trigger reactivation directly, but if it results in a reduction in stability, it will make a landslide more sensitive to other perturbations, such as pore-pressure changes (the rainfall threshold for reactivation will change) and earthquakes.

#### *9.1.4 Earthquakes.*

Earthquakes can reactivate existing landslide features and trigger new landslides (Highland, 2003; Pacific Northwest Seismograph Network, 2001; Schulz et al., 2012). Interestingly, inventories of earthquake triggered landslides worldwide indicate that the number of new landslides may exceed that of reactivated landslides in many cases (Keefer, 1984). Topography can focus seismic shaking, so that earthquake-triggered landslides tend to occur at higher topographic positions, nearer ridge crests, than landslides triggered by other processes (Meunier et al., 2008).

Processes that trigger landslide reactivation do not act independently or in isolation. Landslide sensitivity to increases in pore-pressure can be increased by erosion of the landslide toe; potential for seismic shaking to reactivate a landslide is greater when groundwater levels are high.

### 9.2 Reactivation Potential

#### *9.2.1 Rainfall Thresholds*

Evidence of a rainfall threshold for reactivating existing landslide features suggest that some landslides have a rainfall threshold for triggering movement and, therefore, if no evidence of movement exists, that the threshold has not been met for at least as long as such evidence would persist (Hong et al., 2005). This reasoning is often applied in assessing potential for landslide reactivation in evaluation of forest practices; that is, lack of evidence of past landslide activity is interpreted to mean low probability of future activity, and if past harvest within the groundwater recharge zone to a landslide did not trigger activity, future harvest will not likely trigger activity either (e.g., see the Engineering Geologic Recognizance Memorandum for the North Zender Timber sale in Watcom County: http://file.dnr.wa.gov/publications/sepa/amp_sepa_nw_ts_northzender_geo.pdf).

#### *9.2.2 Temporal variability in precipitation*

Temporal variability in the supply of water to a landslide drives the pore-pressure fluctuations that can trigger movement. Water supply is driven by precipitation minus evapotranspiration (effective precipitation) and the proportioning of that water over the four pathways for input to a landslide body (direct infiltration, surface runoff, subsurface runoff, groundwater). Temporal variability in precipitation is substantial. Figure 20 shows a time series of annual precipitation averaged over Puget Sound. Over this extensive area, precipitation year to year can vary from the mean by nearly plus or minus 40%; local variability may be greater. The potential of exceeding a rainfall threshold for triggering landslide movement over any time period varies considerably depending on the sequence of precipitation events during that period.

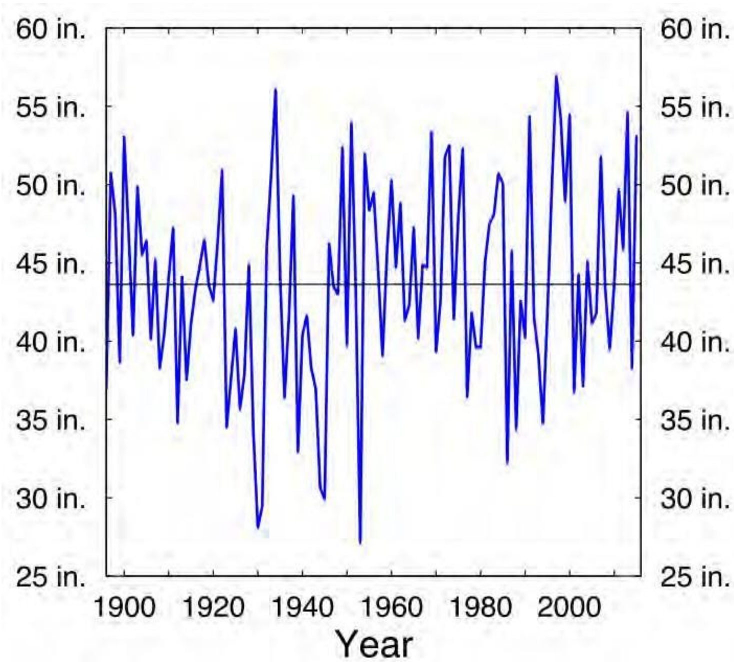



**Figure 20. Annual precipitation over Puget Sound.**

Horizontal line shows average over 1950-1999. From Mauger et al. (2015).

Timber harvest and road construction reduce evapotranspiration. This reduction is evident in direct measurements of evapotranspiration (e.g., Jassal et al., 2009) and through observed increases in water yield following harvest and road construction (e.g., Hubbart et al., 2007; Keppeler and Ziemer, 1990; Rothacher, 1970; Stednick, 1996). Reduced evapotranspiration increases the proportion of precipitation available for input to a landslide. In the Pacific Northwest, the increase in runoff and recharge associated with loss of forest cover ranges from about 10% to 15% of total precipitation (see review in Miller, 2016). This increase can persist for a decade or more (rates of evapotranspiration seem to recover to pre-harvest levels in about 15 years, e.g., Jassal et al. (2009), but harvest effects on water yield can persist for more than 40 years, e.g., Burt et al. (2015).

During that time, increases of water supply to a landslide caused by reduced evapotranspiration are overprinted on changes in supply driven by temporal variability of precipitation. The water available for infiltration, runoff, and recharge (the water yield) is equal to precipitation minus evapotranspiration. Subtracting evapotranspiration from the precipitation values in Figure 20 gives an estimate of water supply. Applied at a particular site, there would be a sudden 10% to 15% upward shift above the average at the time of timber harvest, and that shift would then gradually decrease over a decade or more. The potential for exceeding a rainfall threshold is increased over that period.

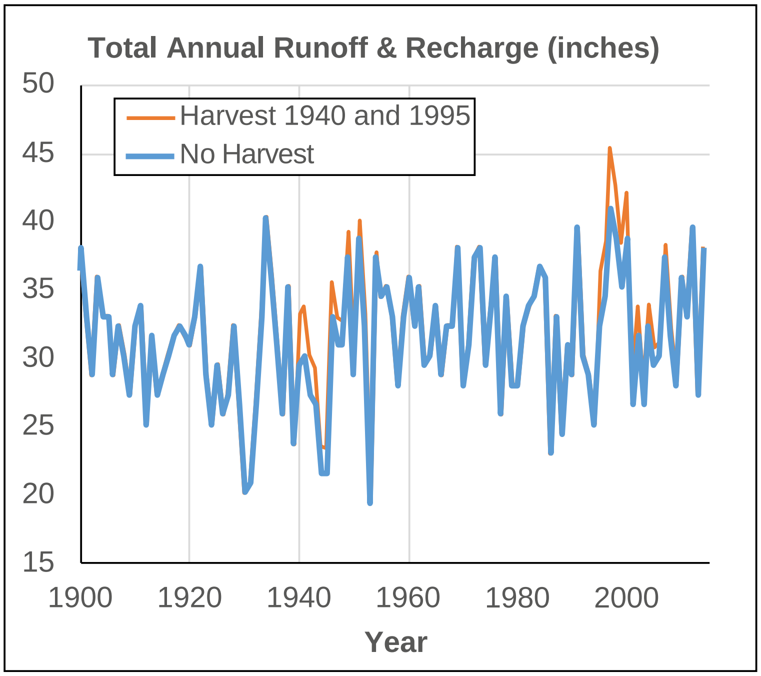



**Figure 21. Water yield with and without harvest.**

This is illustrated in Figure 21. For this figure, we used the annual rainfall depths for Puget Sound shown in Figure 20. Based on measurements from Vancouver Island by Jassal et al. (2009), we set evapotranspiration of a hydrologically mature Douglas Fir stand (~50 years age) at 28% of annual precipitation and that for a recent clearcut at 19%, with recovery to preharvest levels over 15 years. We applied timber harvests in 1940 and 55 years later in 1995. If the threshold for triggering movement on an inactive landslide happened to be 45 inches of water input in one year, the 1940 harvest would have had no effect, but the increased water yield associated with loss of forest cover in 1995 would have reactivated the landslide in 1997.

The potential that the increase in water yield associated with timber harvest will result in exceeding a rainfall threshold for any landslide depends on the sequence of precipitation events over the period affected by the harvest. The probability of exceeding historic antecedent conditions that did not trigger movement requires assessment of temporal variability in precipitation overprinted with increases in water yield associated with forest practices.

Even with such an analysis, lack of past activity may not be a reliable indicator of future behavior. Past patterns of precipitation and evapotranspiration may not be indicative of future patterns. Observed trends over time (Rosenberg et al., 2010) and modeling of future climate (Mauger et al., 2015) both suggest that the frequency of large precipitation events is increasing. Additionally, landslides themselves change over time, and these changes may render them more or less stable and sensitive to large precipitation events.

### 9.3 Temporal variability in landslide properties

Rainfall-triggered landslides that reactivate after some long period of inactivity tend to respond to a period or event with above-average precipitation. Although above average, such triggering periods or events may not be particularly infrequent. The landslide may have withstood more-severe events in the past without activation. Why would a landslide remain stable for long periods, potentially weathering many potentially triggering events, only to start moving after some particular event or sequence of events? For example, Figure 22 shows monthly rainfall for

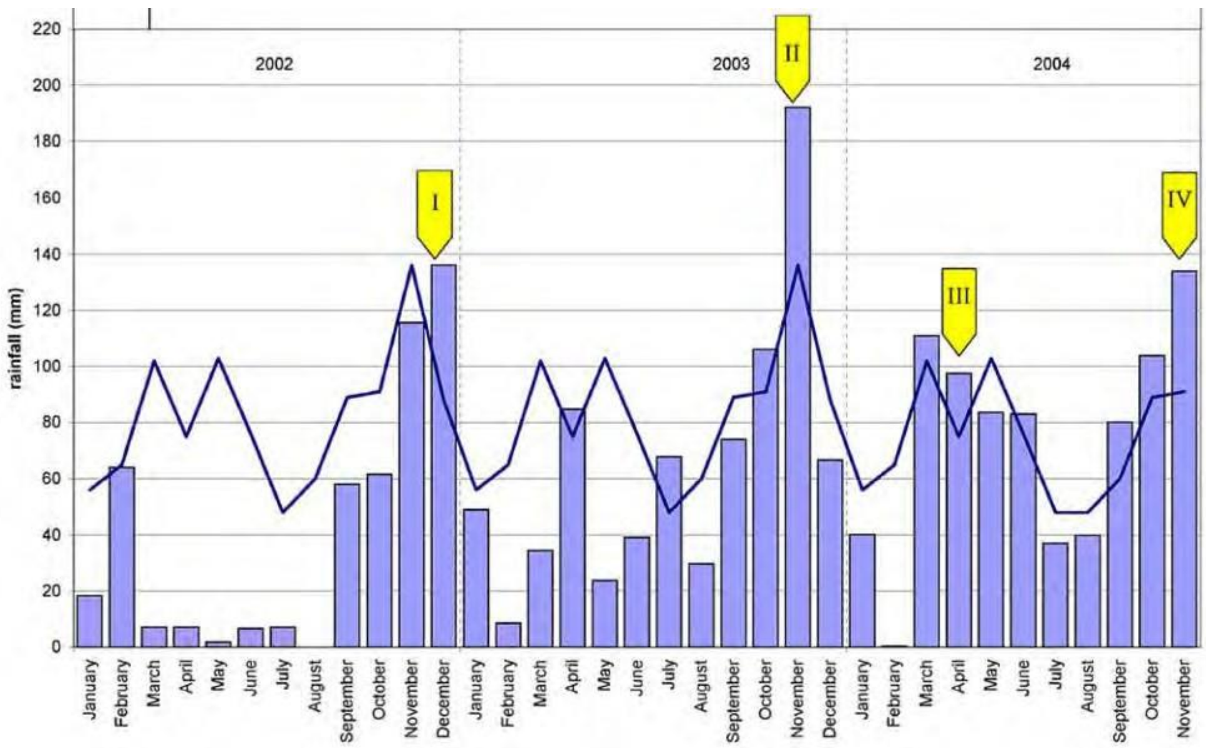



**Figure 22. Rainfall associated with reactivation.**

Blue bars show monthly total precipitation for 2002 through 2004; the blue line shows the average monthly precipitation for the period 1920-1951. Yellow markers indicate initiation of movement on the Ca' Lita landslide. From Borgatti et al. (2006).

2002 – 2004 (blue bars) and average monthly rainfall over the period 1920-1951 for the Ca' Lita landslide in Italy (Borgatti et al., 2006), which reactivated at the end of 2002 after many decades of inactivity. Average monthly rainfall prior to initiation was below average. And if precipitation is the causative factor, why do only a small proportion of the landslides in a region respond to a large precipitation event?

On periodically active landslides, it is observed that rainfall thresholds for landslide initiation can change over time (Priest et al., 2008). The internal plumbing of a landslide inevitably evolves through the growth and sealing of fractures (Binet et al., 2007b) and the development of soil pipes from burrows, root growth, and other processes (Bogaard and Greco, 2016). These changes alter landslide groundwater response to precipitation. Similar changes can occur on inactive landslides. Seasonal variations in water table levels and temperature cause small slope deformations – creep – which may be minor and difficult to detect, but which involve crack growth and modification of subsurface permeability and strength (Cappa et al., 2014; Vallet et al., 2015b).

### 9.4 Examples of reactivation in Washington.

Deep-seated landslide reactivation events in the state are not systematically sought out or cataloged, so there is no representative sample of events available. The events that get noticed are primarily those that cause damage, so geotechnical analyses by the Department of Transportation for mitigating landslide impacts to public roads provides a source for identifying reactivated landslides. The synopsis by T. Badger in the appendix provides a brief list of examples.

In UPSAG's review of the first draft of this document, Casey Hanell added this text:

> "Almost every forest practice proposed now involves a timber stand that is in its second or in some cases third harvest rotation. Both landowner and regulatory engineering geologists are frequently involved in reviewing timber harvest proposals on the ubiquitous non-glacial deep-seated landslides in Washington. Evaluations commonly consist of a review of historical aerial photos to examine previous harvest activity and landslide response. This methodology has limitations as noted above including consideration of climate conditions following harvest. However, these evaluations take place for deep-seated landslides harvested in all of the decades since industrial forestry began, which include all climate conditions in those decades, and a pattern of deep-seated landslide reactivation in response to timber harvest has not emerged. *For certain, some of these assessments have identified deep-seated landslide movement that correlates with previous harvest activity*. In most cases, evidence of deep-seated landslide reactivation in response to previous timber harvest activity is not observed."

This text includes an important statement, which we highlighted with italics, indicating that cases exist where landslide movement correlates with previous harvest activity. As described earlier in this document, evidence suggests that only a small proportion of deep-seated landslides will respond to forest practices. Forest practices are avoided in areas with evidence of such prior response, so these sites tend not to be documented. However, such documentation is needed both to assess the magnitude of the problem (e.g., what proportion of the landscape is involved?) and to identify the characteristics associated with landslide response to forest practices.

In Appendix C, Dan McShane discusses a reactivated landslide in Whatcom County, the "Darrington" landslide along Jones Creek above the town of Acme.

### 9.5 Effects of Forest Practices

Many studies have identified relationships between forest practices and shallow landsliding. Few studies examine connections between forest practices and deep-seated landsliding. As noted in the opening, deep-seated landslides are ubiquitous in Washington. Timber harvest on these deep-seated landslides has also been ubiquitous since industrial logging began around the beginning of the twentieth century. We found no studies that investigate a cause and effect relationship between forest practices and the reactivation of non-glacial deep-seated landslides in the Pacific Northwest. This includes systematic watershed assessment projects conducted during Watershed Analysis and Landslide Hazard Zonation. This does not imply that such connections do not exist.

Abundant evidence shows that forest practices can alter water budgets. Many studies document increases in water yield following forest harvest and road construction (Hubbart et al., 2007; Moore and Wondzell, 2005). Other studies document increased groundwater levels and greater persistence of high groundwater levels following timber harvest (Dhakal and Sidle, 2004; Johnson et al., 2007; Keppeler and Ziemer, 1990; Keppeler et al., 1994). Both effects can act to increase water supply to deep-seated landslides, which can decrease landslide stability.However, the only study we encountered that directly sought to document relationships between forest practices and deep-seated landslide movement in the Pacific Northwest is that by Swanston et al. (1988). They observed an increased rate of movement from about 3.4mm/yr to about 20.5mm/yr on a portion of an active, slow-moving earthflow following clear-cut harvest (including over the entire earthflow itself) that did not correlate with any change in precipitation and was not seen in areas outside the earthflow. This increase persisted for less than three years. The rates of movement observed are very small, involving only about 70mm of total surface displacement on the active portion of the earthflow over ten years. Such minor movement hardly seems to pose a hazard. However, a single case study provides no information about how representative of other earthflows these results might be. All that can be said definitely is that this study did document an six-fold increase in movement rate associated with harvest on the earthflow.

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### 10.1.1.1 Appendix A. Synopsis of WSDOT Geotechnical Investigations of Non-glacial, Deep-seated Landslides

Provided by Tom Badger

A preliminary search of the Washington State Department of Transportation (WSDOT) files for geotechnical reports of non-glacial deep-seated landslides in Western Washington initially yielded eleven comprehensive investigations completed since the late 1990s. These investigations included detailed site descriptions and mapped limits; some historical information on activity; one or more critical geologic sections through the landslide supported by multiple geotechnical borings; multi-year records of continuous piezometer, precipitation, and inclinometer data; and laboratory test results characterizing material index and strength properties. WSDOT files also include dozens of summary correspondence for the reconnaissance and investigation of other landslides, though with less supporting data.

Table 1 presents some of the compiled attributes for the above-mentioned eleven landslide investigations. The landslides are located in terrain underlain by weathered marine sedimentary and volcanic bedrock. Some of the landslides occurred within or immediately downslope of recent timber harvests. Most/all initiated during or shortly following major regional storms, either as first-time failures or, more commonly, as reactivations. Failures were dominantly translational with the failure zone sub-parallel to the ground surface, on slopes often less than 20° (35%), and at depths between 20 and 60 feet. Groundwater response to precipitation and landslide response to groundwater flux have not been evaluated for commonalities and differences between sites.

**Table 1**

| Landslide Name | County | Road Impacted | Dimensions (W /L) (ft) | Landslide Material |
|---|---|---|---|---|
| SR 4 MP 25 Vicinity | Wahkiakum | February 1996 storm-related; possibly impacted road several decades prior | 200/300 | residual soil and landslide debris failing over intact siltstone (Lincoln Ck/Astoria?) |
| SR 6 MP 21.15 | Pacific | December 2007 storm-related | 350/700 | residual soil and landslide debris failing over intact siltstone-sandstone (McIntosh) |
| SR 6 MP 21.95 | Pacific | December 2007 storm-related | 250/150 | residual soil and landslide debris failing over intact siltstone-sandstone (McIntosh) |
| SR 6 MP 27 Vicinity | Pacific | December 2007 storm-related | 350/350 | residual soil and landslide debris failing over intact siltstone (Pe Ell Volcanics?) |
| SR 101 MP 29 Johnson Landing | Pacific | slope deformation first noted 5 years prior to major episode related to February 1996 storm | 250/600 | residual soil/landslide debris/siltstone failing over intact siltstone (Lincoln Ck) |
| SR 101 MP 69.8 | Grays Harbor | January 2006 storm-related; slope deformation noted several years prior | 250/500 | residual soil/landslide debris/siltstone failing over intact siltstone (Lincoln Ck) |
| SR 101 MP 184 Bogachiel | Jefferson | August 2004 reactivation on south end; recurrent deformation since at least 1950s | 2400/1000 | residual soil/landslide debris/siltstone-sandstone failing over intact siltstone-sandstone melange (Hoh Assemblage); alpine glacial deposits failing over melange |
| SR 107 MP 4.7 Montesano | Grays Harbor | December 2005 storm-related | 300/350 | residual soil and landslide debris failing over intact siltstone (Lincoln Ck) |
| SR 109 MP 36.5 Pt. Grenville | Grays Harbor | road deformation since at least 1970s | 400/400 | coastal marine deposits failing over intact siltstone-sandstone (Quinault ?) |
| SR 112 MP 32 Jim Creek East | Clallam | November 1990 storm-related; road deformation years/decades (?) prior | 350/700 | minor glacial and deep residual soils and landslide debris failing over intact siltstone (Twin Rivers) |
| SR 410 MP 64.25 | Pierce | annual deformation of road since at least 1980s; more severe spring 2015 | 1000/700 | landslide deposits (volcanic and glacial-derived) failing over intact andesite (Ohanapecosh) |

**Table 1, continued**.

| Landslide Name | Failure Type | Failure Depth (ft) | Borings | Vegetation |
|---|---|---|---|---|
| SR 4 MP 25 Vicinity | translational | 20 - 40 | 6 | forested above road cut and head scarp |
| SR 6 MP 21.15 | translational | 15 - 35 | 9 | forested |
| SR 6 MP 21.95 | translational | 30 - 40 | 7 | forested |
| SR 6 MP 27 Vicinity | translational | 10 - 25 | 3 | landslide and upslope area recently harvested |
| SR 101 MP 29 Johnson Landing | translational; dip-slope | 20 - 43 | 7 | upslope area and upper 2/3 of landslide harvested within preceding 5 years |
| SR 101 MP 69.8 | translational | 60 - 80 | 11 | forested |
| SR 101 MP 184 Bogachiel | translational-rotational | >100 | >20 | forested; mature |
| SR 107 MP 4.7 Montesano | translational | 27-39 | 9 | entire upslope and landslide area harvested several months prior |
| SR 109 MP 36.5 Pt. Grenville | translational | 16-28 | 7 | forested |
| SR 112 MP 32 Jim Creek East | translational; dip-slope; rotational | 40-60 | >10 | upslope and landslide areas harvested within preceding 5-10 years |
| SR 410 MP 64.25 | translational | 60-66 | 5 | forested |

### 10.1.1.2 Appendix B. Proposed Synthesis of WSDOT Geotechnical Investigations of Non-glacial Deep-seated Landslides

Provided by Tom Badger

In Washington State, timber harvests and forest roads on landslide-prone terrain are typically designed without the benefit of subsurface geologic, hydrogeologic, or geotechnical data. This is the state-of-practice for both public and private timberlands. While these data are costly to acquire, they are judged to be necessary, for example, in the management of highway systems to characterize landslide hazards, evaluate associated public safety risks, and design and construct remedial measures.

Deep-seated landslides in non-glacial (and glacial) geologic units episodically impact highways in Washington State, many occurring in areas that are intensively managed for timber production. The Geotechnical Office of the Washington State Department of Transportation (WSDOT) has historically and continues to perform geotechnical investigations to address these landslide problems. These investigations often incorporate drilling, instrumentation, and laboratory testing to define depth and geometry of the failure surface(s), characterize groundwater regimes and their response to precipitation, and recover and test high-quality soil and rock samples to quantify index and strength properties. All of these data are vital to comprehensively and quantitatively evaluate slope stability and the effects of current and future activities on stability.

Deep-seated landslides occur more frequently in particular geologic units, and therefore regions, in Washington State, for example, the Lincoln Creek Formation in the Willapa Hills and the Twin Rivers Formation along the Straits of Juan de Fuca. In the past several decades, WSDOT has performed numerous landslide investigations in these and other landslide-prone regions. A preliminary search performed for these reports identified nearly a dozen WSDOT investigations of non-glacial deep-seated landslides that include detailed site descriptions; one or more critical geologic cross-sections through the landslide; multi-year records of continuous piezometer, precipitation, and inclinometer data; and laboratory test results. Also within WSDOT's files are dozens of other geotechnical assessments for non-glacial deep-seated landslides and other slope stability problems that contain at least some of the above types of data.

Data from these investigations should be scrutinized, as it may be the only source for subsurface geologic, hydrogeologic, and geotechnical information in the more remote, landslide-prone regions and where forest land is intensively managed for timber production. Potentially available data includes:

- preceding land-use activities including construction of the highway
- preceding climatic conditions
- slope and surface morphology
- geotechnical soil and rock parameters
- geometry and depth of failure
- groundwater regimes and their response to precipitation
- landslide response to groundwater flux
- mitigation effectiveness

A compilation and review of the above data could identify commonalities and differences between and among investigated landslides. Identified commonalities could be extrapolated to other areas of similar geologic setting to help characterize unstable landforms where resources for subsurface investigations are limited or unavailable. In this way, characterization of pre-management conditions would inform and greatly improve the assessment of potential effects of management activities. Existing WSDOT subsurface data could further be used to design a variety of future studies looking at effects and impacts of forest land management.

### 10.1.1.3 Appendix C.

Provided by Dan McShane

UPSAG (Uplands scientific advisory group for CMER) is interested in the potential for forestry activities to reactivate landslides. We are not aware of efforts to systematically evaluate the relationships between deep-seated landslide activity, forest practices, and landslide characteristics. It is very likely that observations by forest practice professionals and other on the ground professionals may be aware of reactivated deep-seated landslides in areas where forest practices have taken place. We aware of deep-seated landslides that reactivated in locations of forest practice activity. Some examples are Racehorse Creek (Crider and others, 2010), the upper Trout Creek drainage northeast of Clinton Knob (46° 25' 09.73" N and 122° 09' 8.31" W), Jones Creek, and Pipeline Slide on Sumas Mountain. The proximal relationship of deep-seated landslide movement and recent forest practices should not be viewed as a direct link between forest practices and slide initiation. While forest practices could be a contributing factor to reactivation or initiation, other explanations and factors covered in the literature synthesis may be as or more important. For example Crider and others (2010) note that "The correspondence between the boundaries of recent logging activities and the boundaries of the rock slide are striking; however, there is no direct evidence linking the slide to timber harvest or road building". Resolving the role of forest practices contribution may not be resolvable, but these slides as well as others may provide valuable insights.

Of the deep-seated slides mentioned above, we further discuss the deep-seated landslides at Jones Creek in Whatcom County where geology assessments took place prior to and after forest practices on and in the vicinity of deep-seated landslides and ongoing monitoring of the deep-seated landslides has been taking place.

**Jones Creek**

The Jones Creek drainage has received considerable geologic attention due to the potential risk to the community of Acme located on the Jones Creek alluvial fan within the South Fork Nooksack River Valley. A debris flood on Jones Creek in 1983 caused considerable damage in Acme (Weden, 1983). Hazard mapping of the alluvial fan by Fox and others (1992) recognized that most of the community of Acme was at risk from debris flow and related hazards. Thorsen and others (1992) recognized that deep-seated bedrock landslides were present within the Darrington Phyllite within the Jones Creek watershed. The phyllite is a mechanically weak rock weathering to slaty chips and clay-rich residues. The phyllite is highly folded and faulted and is prone to failure by deep-seated rotational failures, creep, and slow block glide (Thorsen, 1989).

Raines and others (1996) recognized that the slides in the Darrington Pyllite in the lower watershed had the potential to dam the creek and that much of the alluvial fan material consisted of Darrington Phyllite. They noted erosion and sloughing of the toe of the deep-seated landslides and postulated an acceleration in the movement of these deep-seated slope failures following the channel scour from the 1983 debris flow event. They noted that catastrophic failure of either the toe or the deep-seated mass had the potential to create a large magnitude landslide dam-break event or debris flow.

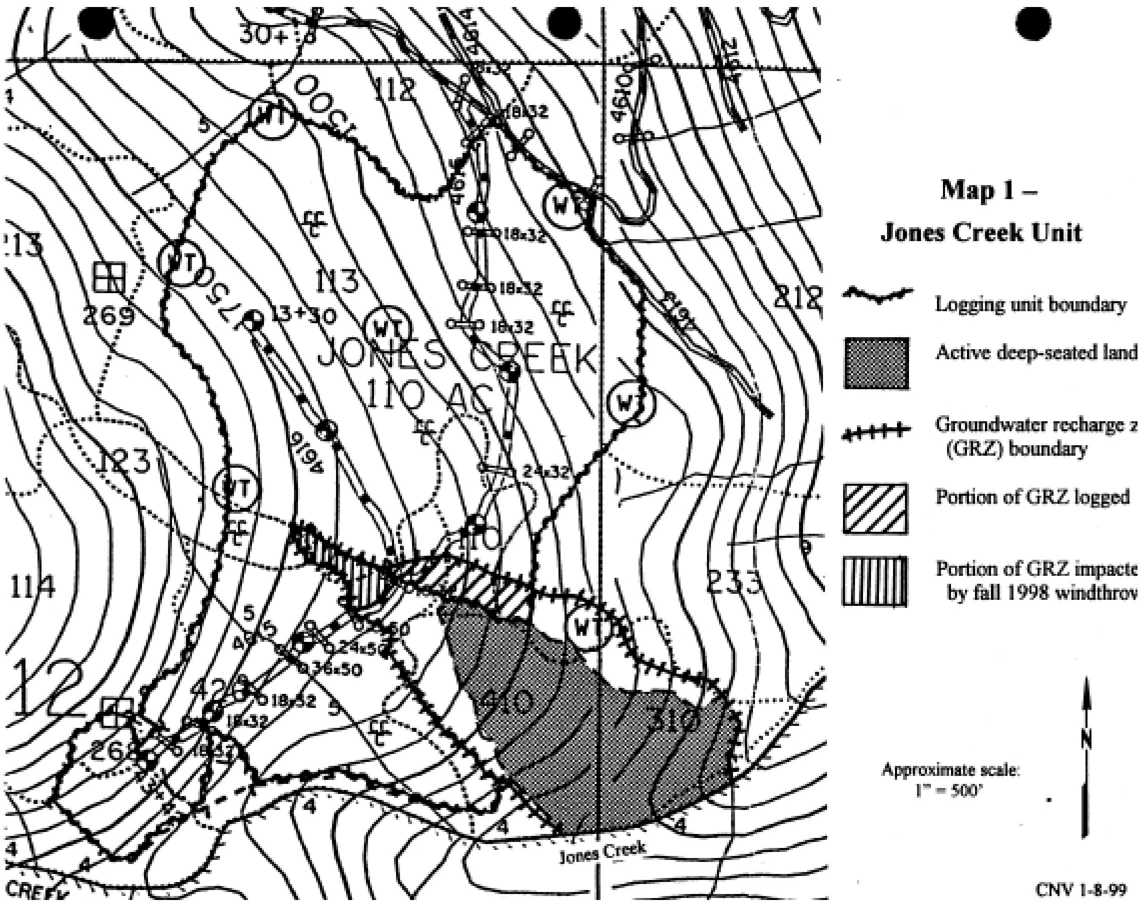
The Acme watershed analysis (Benda and Coho, 1999) included the Jones Creek drainage. The Acme watershed analysis was begun in 1994, but went through a lengthy review process before its completion in 1999. The watershed analysis identified the areas of deep-seated landslides and identified the deep-seated slide areas as MWMU #9. Recommended prescriptions for the deep seated-landslides were as follows:

> No road construction through the active portion of a deep-seated failure.
>
> For roads constructed through the GRZ (groundwater recharge zone), road drainage should be designed to minimize water accumulation in ditches and prevent diversion between sub-drainages.
>
> No logging in the active portion of the slide.
>
> Logging within the GRZ can occur under the following conditions:
>
> - Clearcut logging must leave an uncut buffer along the margin of the active area, covering an area equivalent to 50% of the active portion
> - Selective logging must preserve a minimum relative density of 35 among residual stems >25 years old
> - A detailed study by a geotechnical specialist indicates that slide activity did not increase following prior logging or road construction

In addition to the above prescription recommendations the watershed analysis recommended monitoring, “In an effort to gain a better understanding of the factors which influence its movement, it is recommended that affected landowners adopt and implement a program for monitoring active deep-seated landslides, particularly when harvesting within the GRZ.

Monitoring may include annual site inspections or use of aerial photographs during 5-year reviews of watershed analysis."

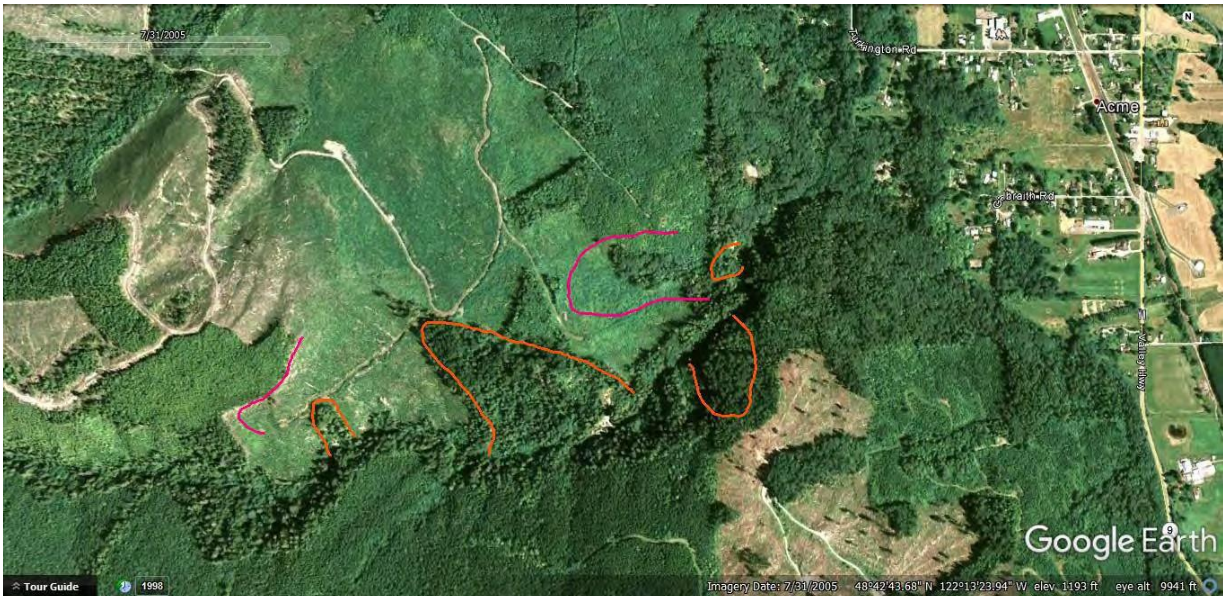
Within Jones Creek watershed harvests have taken place on inactive deep-seated landslides and within groundwater recharge areas of identified active deep-seated landslides.




From Veldhusen (1999)

Veldhusen (1999) noted that post partial harvest in the GRZ, some additional areas of the GRZ lost canopy due to windfall. Veldhusen (1999) estimated the increase water recharge from the windfall loss.

Since harvest activities in the late 1990s, Kerr Wood Liedal (2004) mapped the active deep-seated landslides (as of 2003) and mapped landslide scarps associated with inactive slide areas in the lower Jones Creek drainage as part of their Jones Creek Debris Flow Study. Their mapping of slide areas is transposed on the 2005 aerial image of the Jones Creek drainage.




2005 Google Earth image. Orange lines designate active slide areas and pink designate scraps on dormant slides.

The active slide on the left portion of the image may not have been active to the extent that it was in 2005 at the time of harvest in approximately 1998 and hence it is not buffered and the GRZ was harvested. Kerr Wood Leidal (2004) mapped the slide as active based on their 2003 field work. There was a buffer within the groundwater recharge zone for the large active slide area suggesting that it was active in 1999. Since the 2003 mapping by Kerr Wood Leidal (2004) slide activity has increased and the slide area has expanded.


Google Earth 2016. Note very active slide area. Active slide area also includes forest area within slide.

Since the initial Kerr Wood Leidal (2004) study additional monitoring and documentation of slide activity has been undertaken by Whatcom County. Kerr Wood Leidal (2010) reported wide-spread slope destabilization at several locations in the watershed. They noted that slide activity was causing periodic blocking of the stream. Since 2010, observations of the active slide area have found continued movement and breakup of the slide mass, expansion of the slide and new slide activity on the south side of the creek (John Thompson, Steve Fox, and Paul Pitman, personal communication, and McShane personal observations).

The USGS working with Whatcom County has installed a stream gage on Jones Creek. The gage was set up to be a warning system for the community as a stream blockage from sliding could be a precursor for a large debris flow event. Data from the gage could be used to correlate slope movement, rain fall events and stream flow as well as documenting times when slide movement blocks or restricts the stream.

Slope stability modeling of the slides was conducted by Brayfield (2014). He modeled slope stability of the deep-seated landslides in the lower Jones Creek watershed using various timber harvest scenarios.

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**10.1.1.4**