Hillslopes

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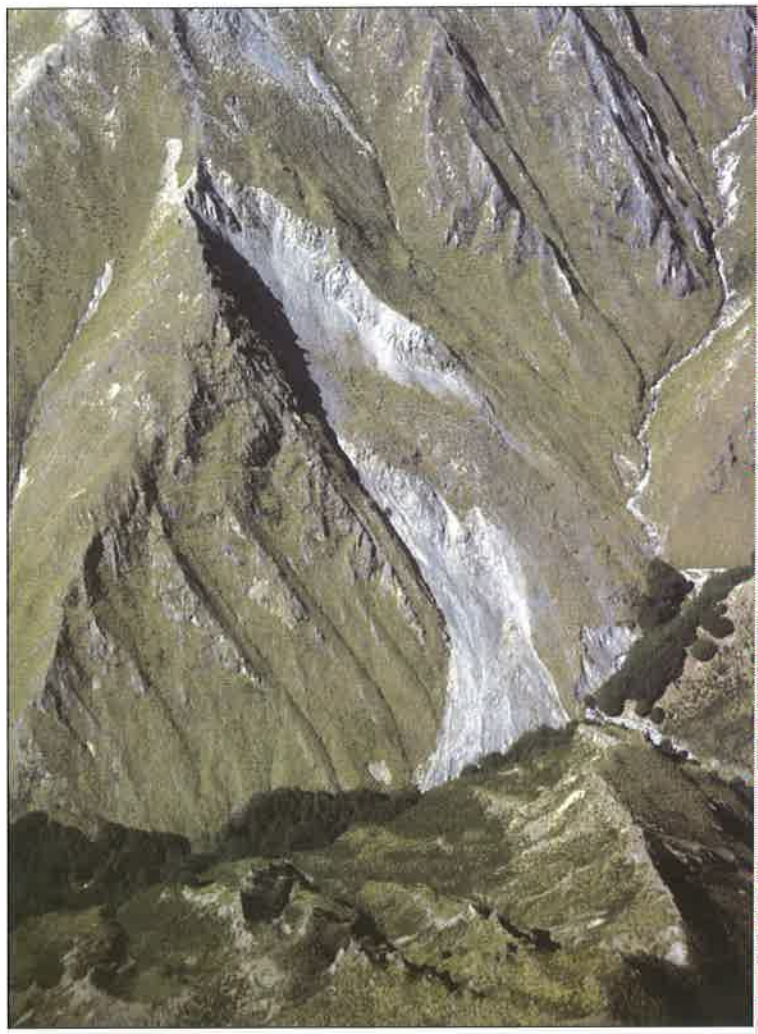


D. Montgomery

Steep, threshold slopes in the mountains of Alaska. In the center of the image, a gray rockslide partially covered with tundra vegetation descends from a ridgeline, entering the river below.

Introduction

Hilltops and hillslopes, the elevated land between valley bottoms, account for much of the landscape; thus, understanding hillslope processes is central to understanding landscape evolution. In general, hillslope topography reflects the nature of the slope-forming materials, the environmental factors that govern processes on inclined surfaces, and the history of specific landscapes.

From a geomorphological perspective, the pace of soil production and sediment delivery determine the sediment supply to channels at the base of slopes and influence fluvial processes far downstream. Eventually, this sediment is transported to depositional basins, coastal plains, and continental margins. Many processes that transport material down hillslopes do so without flowing water. For example, **mass wasting** is hillslope sediment transport during which soil and rock move downslope when the gravitational stress acting on a slope exceeds the slope's ability to resist that stress (its strength). Rates of downslope transport range from the slow movement of soil displaced by burrowing animals, trees falling over, and gravitational creep to catastrophic landslides that can move a neighborhood in an afternoon or destroy a house in less than a minute.

For society, properly assessing the nature of geologic hazards and the environmental impacts of upland land use depends on understanding how our actions influence hillslope processes and the places, styles, and rates at which such processes occur. Hillslope geomorphology has practical implications because upland land use influences the

IN THIS CHAPTER

stability of steep hillslopes and the supply of sediment to river systems. From a biological perspective, the style of hillslope sediment transport defines the geomorphic disturbance regime; rapid catastrophic delivery of material has impacts different from slow, steady sediment movement.

This chapter discusses the properties of slope-forming materials and their influence on hillslope processes, topography, and landforms. We emphasize the key distinction between soil-mantled and bare rock slopes. The chapter also explains how erosion is related to slope steepness, climate, tectonics, and lithology. We begin with the nature of slope-forming materials before discussing controls on hillslope processes and slope form.

Slope-Forming Materials

The properties of slope-forming materials exert a profound influence on both hillslope processes and form. Slopes made of loose, unconsolidated sediment, slopes mantled by soil reinforced by plant roots, and slopes that expose bedrock each offer substantially different resistance to erosion and gravity-induced failure. Because of this, the material that makes up a slope strongly influences the processes that determine that slope's morphology and evolution [**Figure 5.1**].

Rock is made of mineral grains that are bound together by interlocking crystal structures or interstitial cement. Bedrock properties directly influence the morphology of bare rock slopes. Underlying bedrock also influences the stability, soil properties, and topography of soil-mantled slopes. **Saprolite,** deeply weathered bedrock that is still in place, is usually weaker and more prone to failure than unweathered bedrock, because the transformation of primary minerals into secondary minerals involves volume and chemical changes that reduce material strength [**Photograph 5.1**].

Bulk material properties measured on samples brought into the lab are typically used to characterize soil and rock strength, but it is widely recognized that the strength of rock masses is typically determined more by the frequency and orientation of discontinuities than by the properties of intact, unfractured rock. For example, slopes underlain by rock layers that dip parallel to the slope surface are more prone to instability and will adopt a lower gradient

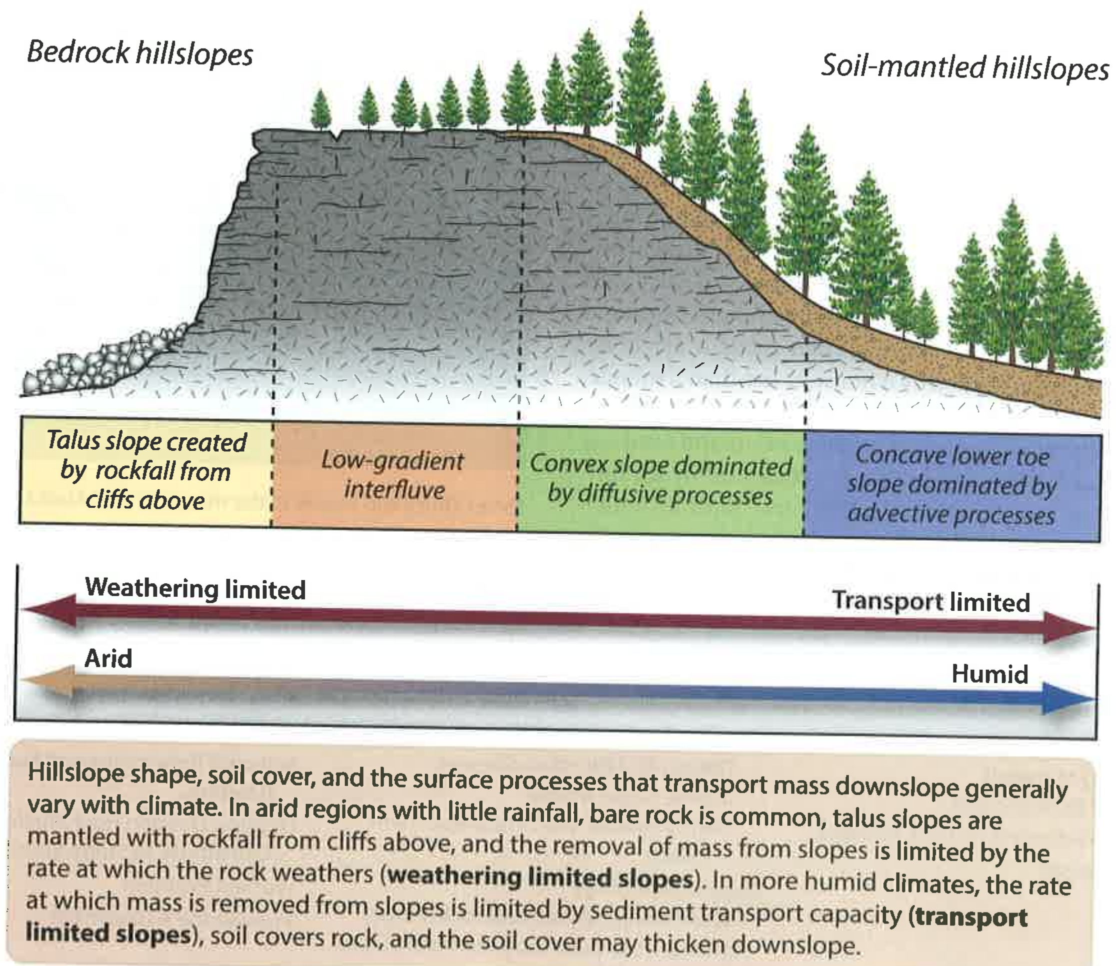


FIGURE 5.1 Hillslope Geometry and Nomenclature. Hillslopes can be categorized by whether they are mantled by soil or regolith, in which case they are known as transport limited, or whether they are characterized by extensive bedrock outcrops (weathering limited slopes). Each type of slope is shaped by a characteristic suite of surface processes.

PHOTOGRAPH 5.1 Saprolite. Weathered biotite gneiss saprolite in Clemson, South Carolina, is bright red from iron oxidation. The original foliation is preserved, although many of the original primary minerals have weathered away and been replaced by secondary minerals.

than slopes underlain by rock with fractures that dip into the slope [**Photograph 5.2**].

Colluvium is the unsorted, mobile or potentially mobile hillslope material that overlies a more stable substrate, including bedrock and consolidated sediments. It generally forms by weathering of underlying material and moves downslope as the result of processes that involve gravity-driven mass wasting, frost wedging, and animal burrowing. Even with bedrock parent material, hillslope soils tend to be colluvial because they are transported downhill by the force of gravity.

In some places, hillslopes are formed of unconsolidated sediments that were deposited by rivers (**alluvium**), glaciers (**till**), or wind (**loess**). In such cases, the material properties of the slope and surficial materials may be similar to those of a soil-mantled bedrock slope, even if little to no soil development has taken place. Surficial deposits that have been overridden, compacted, and thus strengthened by glacial ice are an important exception, because they can form extremely resistant vertical cliffs that behave more like rock outcrops than loose sediment.

PHOTOGRAPH 5.2 Dip Slopes. Dipping quartzite beds at Keurbooms along the southern coast of South Africa. The outcrops fail along bedding plane dip slopes when the toe of the outcrop is undercut by waves.

Strength of Rock and Soil

The rock and soil that make up Earth's surface vary greatly in material strength and ability to resist erosion. Granite is difficult to break with a sledgehammer because the material is cohesive and has high **compressive strength** (resistance to squeezing) and high **tensile strength** (resistance to pulling apart). A loose pile of sand at the base of a slope of weathered granite can be scooped up with a spoon (see Photograph 3.12) because it has no cohesive strength. Deeply weathered saprolite in tectonically stable continental interiors or tectonically shattered bedrock in rapidly uplifting mountains may have less strength than an overlying clay-rich soil B horizon. The strength of slope-forming materials is a dominant influence on slope processes and topography.

The ability of material to resist shearing stress (the sliding of one body over another) is called its **shear strength,** a property that is quantified by three components, two of which are intrinsic material properties—the **angle of internal friction** (a measure of frictional strength commonly referred to as the **friction angle**) and **cohesion** (the tendency of material to stick to itself). The third component, called the effective **normal stress,** is the material's unit weight (the product of bulk density and gravitational acceleration) perpendicular to the slope (a force) considered per unit area of the slope less the **buoyant force** due to any interstitial water. The way that these factors together determine the shear strength of a material is described by the Coulomb equation,

$$SS = C + \sigma' \tan \phi \qquad \text{eq. 5.1}$$

where SS is shear strength, C is cohesion, σ' is effective normal stress, and ϕ is the angle of internal friction [**Figure 5.2**].

The frictional strength of rock and soil (ϕ) arises from the resistance to shear between mineral grains that are in contact across potential failure surfaces. Frictional strength increases in direct proportion to the normal stress holding grain surfaces in contact. The friction angle, or the angle of internal friction (ϕ), corresponds to the slope of

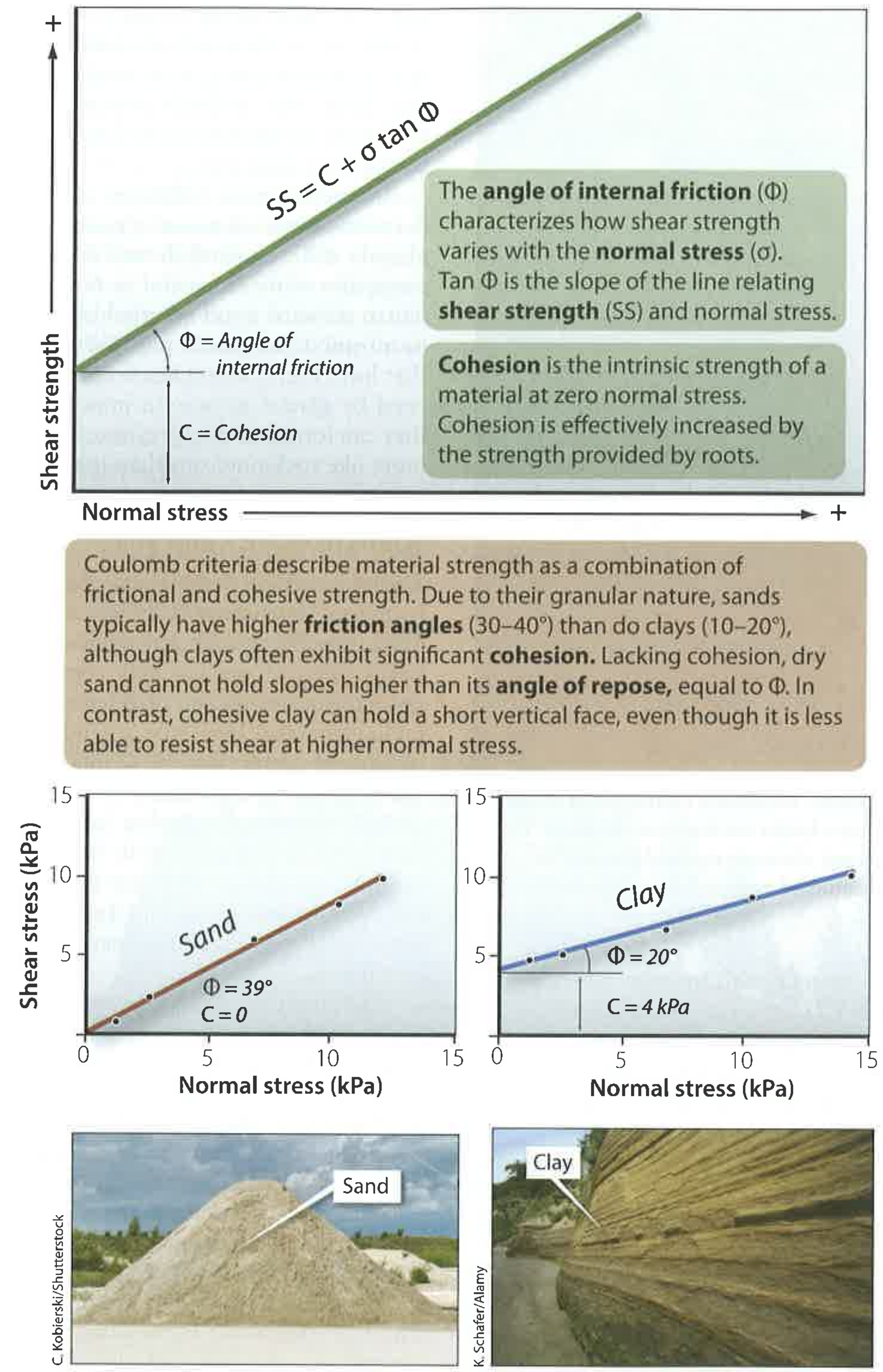


FIGURE 5.2 Shear Strength (Coulomb Criteria). The strength of earth materials has two components: a frictional component that is linearly related to the applied normal stress and a cohesive component that is unrelated to normal stress. Together, these are known as the Coulomb criteria.

the line that describes the relationship between shear strength and the confining stress (which on a hillslope is equal to the effective normal stress).

In other words, as the effective normal stress increases, the shear strength of a material increases at a rate set by the friction angle. In loose, granular materials, the friction angle is usually close to the **angle of repose**, the maximum angle at which a slope of dry, cohesionless material can stand. Rock masses and granular material like sand typically have friction angles of about 30 to 40 degrees; clay has a lower friction angle of 20 to 30 degrees. The great difference in strength between loose sand and rock is due to cohesion; in contrast to mineral grains making up a rock, grains of sand are not bonded to each other and thus have no cohesion.

Friction angles for both soil and rock generally fall in the range of 10 to 40 degrees, but cohesion values of rock are typically many orders of magnitude greater (100,000 kPa) than those of most soils (10–100 kPa, **Table 5.1**). (*Note*: 1 Pa = kilograms per meter per second squared, kg/ms^2.) Consequently, there is a profound difference in strength on many slopes where weaker soil and weathered rock lie above much stronger bedrock. Because of the disaggregation of soil and rock particles once slope failure occurs, the postfailure strength of earth materials often is less than the peak strength before failure. For most hillslope processes, it takes less force to maintain downslope transport than to initiate it. Once it is moving, slope-forming material tends to keep going until it spreads out onto a gentler slope or dissipates its kinetic energy as friction when flowing over, through, or around whatever was in its way.

Cohesion is a measure of the intrinsic strength of a material when there is no normal stress. This corresponds to a material's y-intercept value in Figure 5.2 and varies greatly among slope-forming materials (see Table 5.1). Cohesive strength arises from various types of internal bonding, including the chemical bonds and interlocking fabric between mineral grains that provide substantial strength to crystalline rocks. Electrostatic bonding between the charged surfaces of clay particles and ions in interstitial water enhances the cohesion of clay-rich soils, but these electrostatic forces are much weaker than chemical bonds in rock. Sediments compacted by the weight of now-melted glacial ice often have high cohesion.

Interstitial cements like calcium carbonate also greatly increase rock and soil cohesion. Plant roots can impart an **apparent cohesion** to soils in a manner similar to the way steel rods (rebar) contribute tensile strength to concrete. Root strength can make the difference between slope stability and failure for thin soils on steep slopes. But the apparent cohesion from roots changes over time as trees grow, mature, and die—whether from natural disturbance, succession, fire, and senescence or because of root decay after timber harvesting.

TABLE 5.1

Typical strength of Earth materials

Material	Friction angle (degrees)	Cohesion (kPa*)
Soil		
Sandy soil	30–40	0
Soft organic clay	22–27	5–20
Stiff glacial clay	30–32	70–150
Rock		
Intact sandstone (lab)	35–45	>10,000
Intact shale (lab)	25–35	>1,000
Sandstone (field)	17–21	120–150
Shale (field)	15–25	40–100

*1 Pa = 1 kg/ms^2; 1 kPa = 1000 Pa
lab = laboratory data on small samples
field = data collected from field measurements

Normal stresses—those oriented into the slope—help hold soil on hillslopes. On a dry slope, the normal stress (σ) from the weight of dry rock or soil is supported by the contacts between grains. The normal stress is greater on gentle slopes than on steep slopes because it is proportional to the component of the weight of the soil ($\rho_s\, g\, z_s$, where ρ_s is soil density, g is the acceleration of gravity, and z_s is soil depth) oriented into the slope, and thus is a function of the cosine of the slope angle (θ):

$$\sigma = \rho_s\, g\, z_s \cos\theta \qquad \text{eq. 5.2}$$

However, if water fills the void spaces between rock or soil particles, it reduces the normal stress (σ) by an amount equal to the pore-water pressure (μ). One way to understand this effect is to consider that some of the weight of the overlying material is supported by pressurized water rather than by solid material contacts. The remaining portion of the normal stress that is supported by a rigid network of grain-to-grain contacts is called the **effective normal stress** (σ') when considered per unit area, and this is the normal stress that matters for slope stability:

$$\sigma' = \sigma - \mu \qquad \text{eq. 5.3}$$

In dry soil where $\mu = 0$, the effective normal stress is equal to the applied normal stress (i.e., $\sigma' = \sigma$). In partially saturated soils, the surface tension produced by capillary stresses can increase stability through negative pore pressures, but only to a point. Negative pore pressures are significantly reduced as a soil approaches saturation. Thus, they do not contribute much, if at all, to soil strength at the time it is needed most—in the middle of a soaking rainstorm.

Below the water table, positive pore pressures ($\mu > 0$) reduce the effective normal stress and lower the shear strength of the soil. The higher μ becomes, the greater the reduction in the effective normal stress. Thus, landslides tend to happen during and after rainstorms because even partially saturated soils are much weaker than dry soils. A slope does not, however, need to be completely saturated to fail. Slopes fail when they become saturated enough that material strength is less than the shearing stress, a condition that requires lower pore pressures (μ) on steeper slopes, as we will see below.

Effects of Weathering on Rock Strength

Weathering lowers rock strength over time through physical and chemical alteration of rock properties and by changing slope hydrology. The cohesion of weathered rock, soil, and unconsolidated sediment is generally much lower than that of intact rock, and the development of zones of weakness as weathering proceeds also greatly reduces rock strength. As fresh rock weathers to become soil, porosity and permeability increase, sometimes by orders of magnitude. Such changes proceed preferentially along fissures and fracture zones and produce patterns of variable weathering intensity within the rock (see Figure 3.13).

Many slope failures involve sliding of the surficial soil and sediment mantle over underlying bedrock, because unweathered or lightly weathered rock has much higher cohesion and is much stronger than soil. In landscapes with intense chemical weathering, as is common in humid-tropical regions, a zone of pervasively weathered, virtually cohesionless saprolite that extends deep beneath the soil may slide off a slope as pore pressures rise during storms. Such saprolite may erode rapidly if gullied by surface water drainage or seepage pressures generated by high groundwater tables. The **gullies** of Madagascar are prime examples of hillslope erosion in heavily weathered, saprolitized bedrock [**Photograph 5.3**].

Calculations based on strength values measured in hand specimens of rock suggest that some rock types should be capable of supporting vertical cliffs far taller than any that exist in nature. Why the discrepancy? Over time, weathering and tectonic stresses reduce the strength of even the most resistant rocks at or near Earth's surface and produce the deep-seated bedrock landslides that are common in pervasively fractured, tectonically active upland landscapes. Laboratory measurements of shear strength use small intact samples of rock and thus generally overestimate rock strength because lab work cannot take into account the localized but critical zones of weakness, such as fractures and bedding surfaces that control the actual strength of slopes. Because of the tremendous range of slope-forming materials and their relationship to topography, methods for characterizing rock strength have been developed based on classification systems that describe the qualities of rock outcrops in the field, including the orientation and density of bedding planes, faults, and fractures.

R. Cox

PHOTOGRAPH 5.3 Madagascar Gully. This large gully near Amparafaravola, Madagascar, is 155 m wide and almost 40 m deep. These features are locally known as lavaka, the Malagasy word for "hole." The gully has a large flat floor and an external debris apron that is partly vegetated. Its headwall has cut back through the ridge crest.

The incision of valleys by flowing water and ice concentrates compressional stresses in valley bottoms and causes extensional stresses along ridgetops and valley walls that help break up intact bedrock. Mechanical unloading at the ground surface commonly results in a zone of fractured rock that extends to some depth beneath the land surface. This creates planes of weakness parallel to the ground surface that promote landslides and accelerate rock weathering by focusing groundwater flow. Chemical weathering by groundwater flow along such fractures further reduces bedrock strength by destroying or weakening the bonds between mineral grains and smoothing asperities (small protrusions) that roughen joint surfaces.

Diffusive Processes

Sediment-transport processes on hillslopes include **diffusion-like processes** (generally called **diffusive processes**) in which the transport rate is proportional to hillslope gradient. Rainsplash, sheetwash, and soil creep are diffusion-like processes that reduce relief and fill in depressions. Diffusive hillslope processes involve sediment movement without entrainment by concentrated flow of water, wind, or ice and are distinguished by basic differences in transport style.

Rainsplash

Raindrops strike bare ground with substantial force during intense rainfall events and the resulting impacts can move a significant amount of sediment by **rainsplash** [**Photograph 5.4**]. Such splash is geomorphically important in regions with scant vegetation cover (such as deserts and disturbed landscapes), because it loosens sediment from soil surfaces, allowing transport by moving water. The thick vegetation cover of humid temperate regions shields many soil surfaces from rainsplash impacts.

On bare sloping ground, rainsplash produces net downslope sediment transport. To understand why this is the case, consider the trajectory of particles ejected from a sloping surface. If the raindrop falls vertically, the same number of particles are directed uphill as downhill; however, those headed uphill run into the slope at a shorter distance than those headed downhill, because the slope, like a ski jump, drops out from beneath the particles transported downslope.

If rainfall rates exceed infiltration rates, as in many desert thunderstorms, then water will begin to flow over the land surface. Raindrops can penetrate shallow overland flow and kick sediment up into suspension. However, as surface water flow deepens, less sediment is entrained because the water shields the soil surface from raindrop impacts. Consequently, the contribution of rainsplash to hillslope sediment transport is largely limited to thinly vegetated areas relatively close to drainage divides, where overland flow is absent or shallow.

USGS

PHOTOGRAPH 5.4 Rainsplash. Rainsplash impacting shallow flow dislodges both water and sediment.

Sheetwash

Overland flow that is not concentrated into discrete channels and spreads across the ground surface is called **sheetwash** [**Photograph 5.5**]. Overland flow is rare on heavily vegetated soil-mantled slopes because the surficial organic layer and porosity caused by root cavities ensure that infiltration capacity generally exceeds rainfall rates and thus that rainwater sinks into the ground. Consequently, sheetwash does not transport much sediment in humid or temperate environments, except where soil has been disturbed or compacted, for example, on construction sites and walking trails (see Photograph 4.3).

K. Nichols

PHOTOGRAPH 5.5 Sheetwash. Sheetwash covers a gentle slope on the coastal plain outside Okambahe, Namibia, during a very heavy thunderstorm. Flow resulted from a high-intensity, short-duration (<1 hour) summer (February) rainfall event. Runoff lasted for about an hour.

Unchannelized sheetwash does, however, transport significant amounts of material down undissected slopes in the arid and semi-arid environments that make up large portions of continental land masses. Sheetwash processes are instrumental in shaping the morphology of hillslopes in these areas, moving mass downslope from slowly eroding hillslopes to depositional basins. Experiments using painted pebbles over the short term and cosmogenic nuclides over the long term showed that sheetwash moved sediment over gently sloping Mojave Desert slopes at rates of at most a few tens of centimeters per year. Sheetwash, while pervasive in these regions, does not change landforms rapidly.

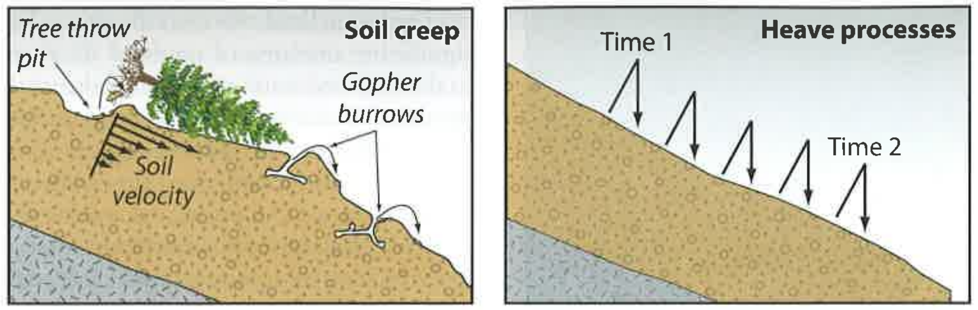
Rainsplash and sheetwash are both diffusion-like processes that fill in topographic depressions and smooth over relief. Where enough overland flow concentrates to incise the ground surface, channels form and the flow then acts to incise and enhance relief at least temporarily. Because sheetwash events can be few and far between in arid regions, other surface processes, such as wind erosion and **bioturbation** (soil-stirring by animals), may erase small channels before the next flow event.

Soil Creep

Soil creep is the incremental downslope movement of soil and sediment. Soil creep processes are too slow to see without instrumental measurement or other indicators of long-term movement. The term incorporates a wide range of processes that include seasonal heave from ice or expanding clays, downslope movement of soil from the burrowing activity of animals, and displacement of material downhill by uprooted trees [**Photograph 5.6**]. In cold regions with permafrost, seasonal thawing of the surficial layers can result in **solifluction,** which involves creep of weak, saturated soils over stronger, impermeable, frozen ground. Seasonally frozen ground also experiences **heave,** cyclic

P. Sanborn

PHOTOGRAPH 5.6 Tree-throw. Tree-throw can move large amounts of material downslope. Here, in the Nahanni National Park in Canada's Northwest Territories, a tipped up rootwad moves the shallow layer of soil downslope. (Lens cap on the top of the rootwad provides scale.)



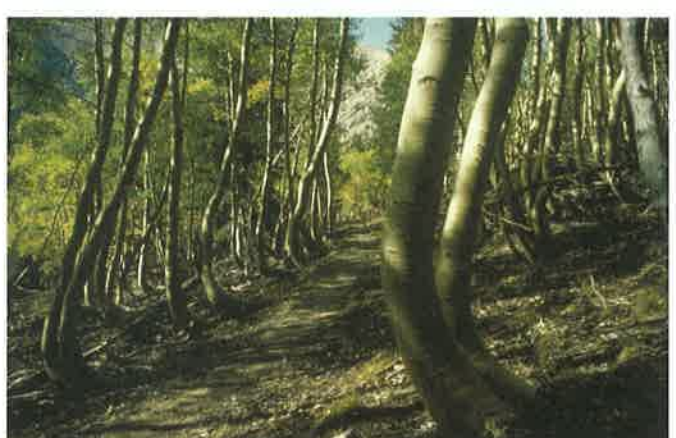
Soil creep describes the suite of processes that move soil and regolith downslope at a velocity proportionate to the to slope angle. Processes contributing to soil creep include tree-throw, animal burrowing, and deformation of fine-grained soil.

Heave contributes to soil creep. Heaving soil rises up perpendicular to the slope through the wetting and expansion of clays or the freezing of interstitial water. When the soil thaws or dries and shrinks, the material drops vertically under the influence of gravity, causing net downslope movement of soil.

FIGURE 5.3 Soil creep. Soil creep is the movement of material downslope at a rate proportional to slope. Heave is one creep process.

expansion and contraction of material that produces net downslope transport because of a differential bias in the direction of movement. On sloping surfaces, soils expand perpendicular to the ground surface, but gravity-induced contraction occurs vertically, resulting in a net downslope movement with each freeze–thaw cycle (or season) [**Figure 5.3**].

Downslope creep rates are variable and depend on slope angle, climate, soil moisture content, and particle size, but they rarely exceed a few millimeters per year. Creep rates typically decrease with depth below the ground surface, and most displacement happens within about a half meter of the surface. Evidence of active soil creep (specifically, the higher creep velocity at the surface than at depth) includes such indicators of net downslope movement as tilted fence posts, curved lower tree trunks (known as pistol-butt trees, in honor of their resemblance to antique dueling pistols), cracked building foundations, and accumulation of soil on the upslope side of fixed obstructions [**Photograph 5.7**]. Slow, gravity-driven deformation of rocky slopes, from repeated freeze–thaw or wetting-and-drying-induced expansion and contraction of near-surface regolith, can lead to downslope bending of fractured and weakened near-surface bedrock [**Photograph 5.8**].



M. Miller

PHOTOGRAPH 5.7 Pistol-Butt Trees. Bent tree trunks resulting from soil creep in Nevada.

B. Bradley

PHOTOGRAPH 5.8 Creep. Creep moves fractured rock downslope near Marathon, Texas. Creep is faster nearer the surface where the rock is less competent.

Mass Movements

The material displaced and the style and rate of deformation distinguish different types of **mass movements** [**Figure 5.4**]. In general, mass movements involve translation of partially to fully saturated soil and/or rock along a well-defined **failure surface**, or **shear plane.** Generally, the failure surface is either approximately parallel to the land surface (**planar landslides**) or extends to some depth as a concave surface along which rotational slippage or slumping occurs (**rotational landslides**). Factors that influence the occurrence of mass movements include ground-shaking, intense or long-duration rainfall, undercutting that removes material buttressing the toe of a slope, and the progressive weathering of hillslope materials.

There are three general types of mass movements: **slides, flows,** and **falls.** These processes can act alone or in combination to form landslide complexes. Mass movements are usually described using a prefix that identifies whether the material involved is rock, coarse soil, and sediment (debris) or fine-grained material (earth). Different types of mass movements can be either wet or dry and slow or fast. Complex mass movements consist of several failure styles.

Shallow mass movements generally involve surficial materials like soil and saprolite. Because there is typically a large strength discontinuity between soil and the underlying bedrock, shallow planar landslides often detach and slide along the soil-bedrock contact. Shallow planar slides are common on steep, soil-mantled slopes, such as the wet, tectonically active Oregon Coast Range.

Deep-seated bedrock failures may involve fresh bedrock as well as weathered surficial material. Bedrock landslides typically involve either failure along a discrete plane of weakness—like a bedding plane or a fault—or slippage along a rotational failure surface. Rockslides, for example, are usually associated with structures like faults, fractures, or joint sets [**Photograph 5.9**]. The orientation of bedding planes can also enhance or decrease slope stability. Bedding that dips back into a slope promotes slope stability; beds that dip in the same direction as the slope promote instability. Hillslopes where underlying strata have an inclination close to parallel with that of the topographic surface are especially prone to sliding (see Photograph 5.2).

Mass movements may be active or dormant, and some can be readily reactivated because the material within landslides and along failure planes generally loses strength after it fails. Recently active landslides can be distinguished from older, inactive landslides based on a variety of criteria. Sharply defined scarps, tilted trees, **sag ponds** (closed depressions that are unconnected to external

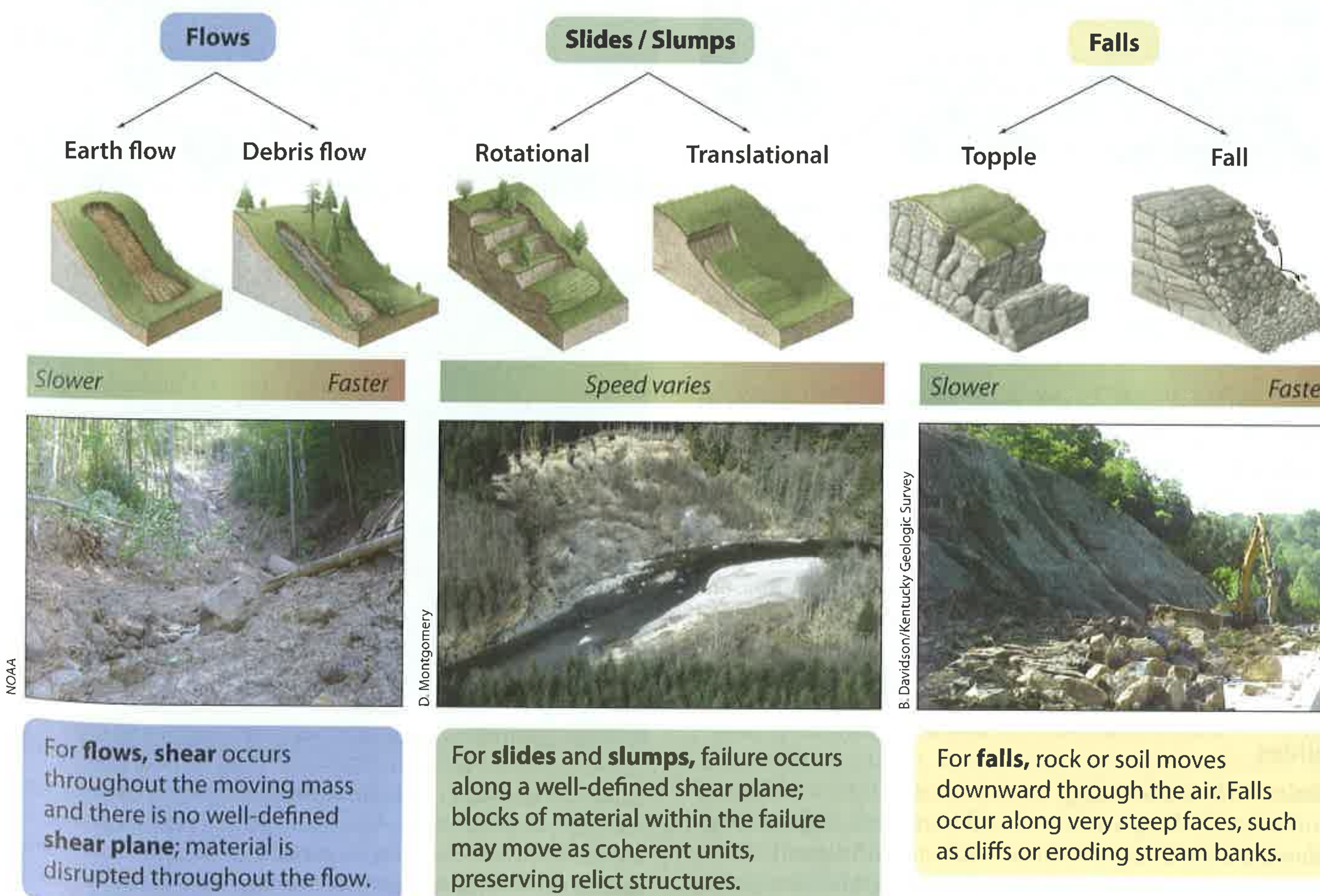


FIGURE 5.4 Taxonomy of Mass Movements. Mass movements are typically classified by both their shape and by the speed at which they move.

(a)

H. Dunn

(b)

D. Thompson

PHOTOGRAPH 5.9 Rockslides. (a) Rockslide on Elephant Rock, Yosemite National Park. The debris spread down the slope, knocking down trees, and the dust cloud covered a wide area. (b) The large boulder failed along the bedding plane in a dip slope rockslide in Death Valley National Park, California.

drainage), and lack of soil development on depositional areas all may indicate recent activity [**Photograph 5.10a**]. Rounded scarps, unaffected trees, well-integrated drainage with few ponds, and soil development on depositional areas generally indicate passage of significant time since landslide activity [**Photograph 5.10b**]. Such features can help determine the relative age of landslides, but the amount of time it takes for a landslide to appear inactive depends on a number of local and regional factors, like lithology and climate, that complicate inferences of landslide age.

Slides

Slides involve downslope movement of cohesive blocks of soil or rock along a relatively thin and well-defined **shear plane,** a zone of intense **shear strain** (deformation). There is little internal shearing within the sliding block or within discrete blocks that move together in the slide. Resistance to movement drops after initial failure, as material weakens during downslope transport and deformation; thus, movement generally continues until the sliding block(s) encounters sufficient resistance to halt further movement, usually because of decreased slope. Slides may have planar or rotational failure surfaces and may be fast or slow moving once initiated. Some slides become flows after initial failure.

Translational slides are typical of many small landslides with shallow planar failure surfaces on which the failed material moves. During shallow, planar slides, coherent slabs of soil move downhill over more solid rock or consolidated soil. Such slides typically occur on planar slopes below ridgelines and in concavities where groundwater flow converges and raises pore pressure, reduces normal stress, and thus lowers the frictional strength of the material. The margins of shallow planar landslides are defined by **scarps,** vertical faces along the top and edges of the detached slide block. Downslope of their **initiation zone,** translational slides can move out over the original ground surface. Translational slides may be soil slides in which blocks of soil detach but retain some internal structure while moving as coherent slabs [**Photograph 5.11a**], or they can be rock slides in which rock slabs detach along planar failure surfaces, such as bedding planes. Translational slides may be rapid or slow, depending on their moisture content and the nature of the failed material. Translational slides can involve whole mountainsides sliding along bedding planes, as occurred in the Frank Slide in Alberta, Canada, and the Love Creek Slide near Santa Cruz, California [**Photograph 5.11b**].

(a)

M. Miller

(b)

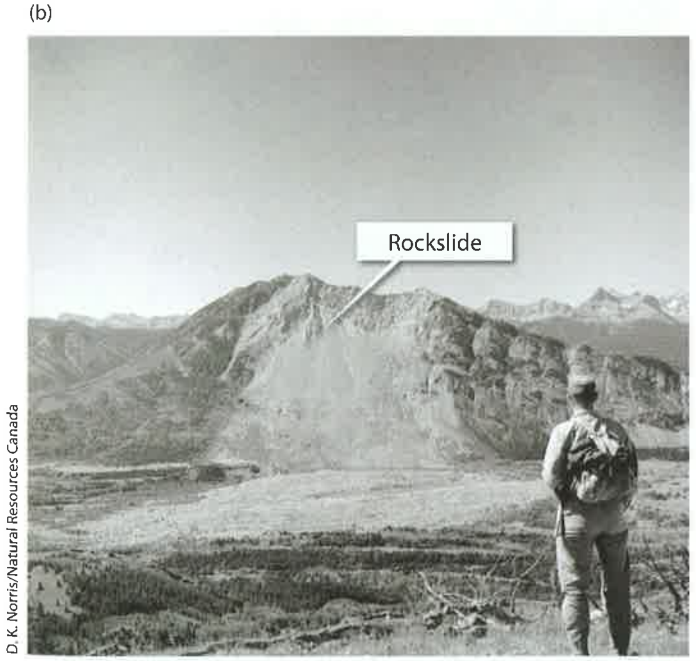
P. Bierman

PHOTOGRAPH 5.10 Active and Inactive Landslides. Recently active and older inactive landslides can easily be distinguished. (a) An active shallow planar landslide, near Florence, Oregon, has recently disrupted a steep forested slope. The slide was likely triggered by wave erosion at the toe of the slide. (b) Inactive, deep-seated bedrock landslide with well-defined scarp in the Bolivian Andes. Landslide is currently stable enough that it is crossed by roads and covered by agricultural fields.

(a)

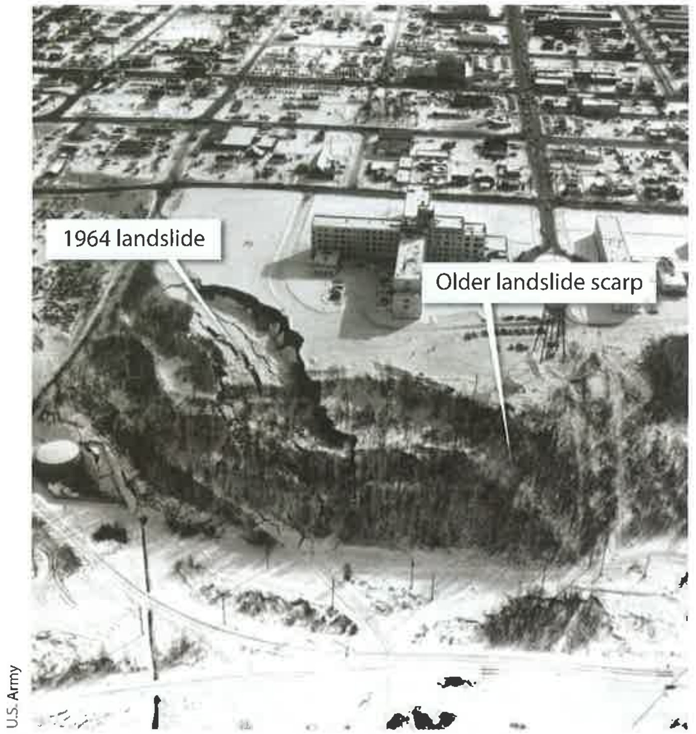
P. Bierman

(b)



D. K. Norris/Natural Resources Canada

PHOTOGRAPH 5.11 Translational Slides. (a) Shallow planar failure in disturbed glacial sediment that occurred during a wet spring, the year after the slope was graded as part of construction of a new highway near Essex, Vermont. (b) The Frank rockslide in Alberta, Canada, failed early on the morning of April 29, 1903, burying almost 100 people beneath about 30 million cubic meters of limestone rock debris.



U.S. Army

PHOTOGRAPH 5.12 Rotational Landslide. A magnitude 9.2 subduction earthquake on March 27, 1964, in Anchorage, Alaska, triggered a rotational landslide. Note the large upper scarp, the back-tilting of blocks, and the bulging toe of the slide. The scar of an older landslide is offset by the recent slide.

Rotational landslides, commonly called **slumps,** involve the movement of soil or rock along a curved, concave failure surface [**Photograph 5.12**]. They typically exhibit a pronounced head scarp, as well as secondary scarps at the heads of back-tilted, rotated blocks within the slide mass and an elevated bulge where material accumulated at the toe of the failure [**Figure 5.5**]. Rotational slides may consist of either a single slump or have multiple nested failure surfaces that create multiple scarps and merge at depth within a single larger complex feature. Rotational slides are common in thick, cohesive, relatively homogenous deposits such as glacial lake clays. They can move rapidly (in seconds) or slowly (over days) and are frequently reactivated, particularly if a stream erodes the toe of the slide.

Spreads are landslides that involve extension of cohesive or hard rock masses due to lateral movement of softer, weaker underlying material. Development of extensional fractures in the overlying rock mass accompanies general subsidence of the fractured cohesive material into the softer underlying material. The dominant mode of deformation in spreads is lateral extension that can break the surficial material into a maze of high-standing **horsts** and downdropped **grabens.** The intricate maze of grabens in Canyonlands National Park in southern Utah represents large-scale lateral spreading of the brittle capping rock (or **caprock**) above underlying, readily deformed, viscous salt that was exposed and remobilized when downcutting of

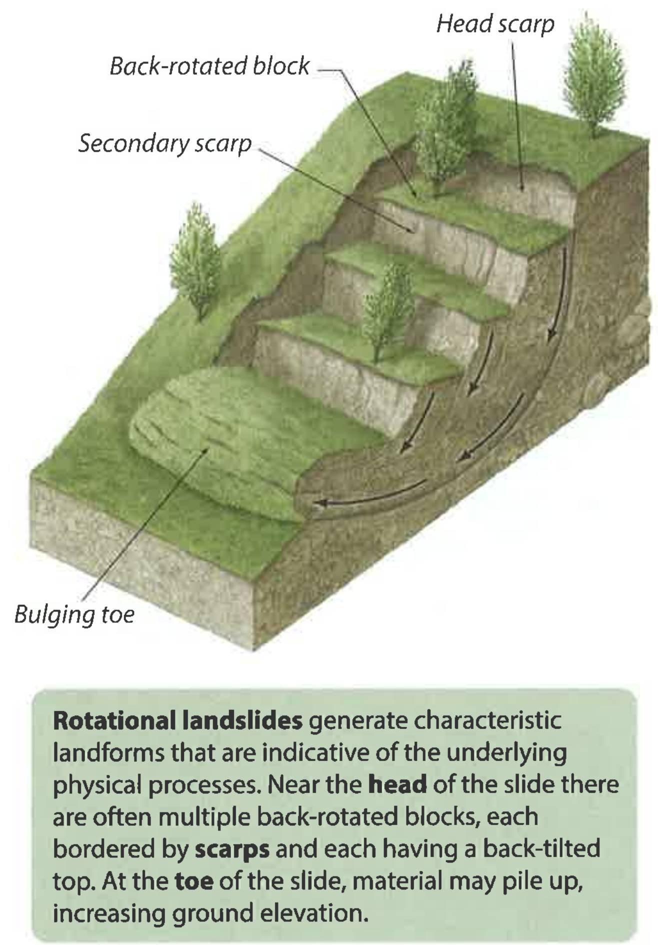


FIGURE 5.5 Rotational Slump. Rotational slumps have specific features that make them easy to identify in the field, including back-rotated blocks, multiple scarps, and bulging toes.

the Colorado River removed the lateral confining support during the past half million years.

Formations of fine-grained sediment deposited in seawater and now elevated above sea level, such as glacial marine clay, are susceptible to spreading for two reasons. After hundreds to thousands of years of subaerial exposure and percolation of freshwater from rain and melting snow, many of the ions that helped to hold the loose clay together have been washed away. If shaken strongly by an earthquake, these already weakened clays collapse, increasing interstitial pore-water pressures and allowing them to flow. A prime example is the spreading that affected the Turnagain Heights neighborhood of Anchorage, Alaska, during the 1964 earthquake.

Flows

Flows move by differential shearing within the sliding material and have no well-defined internal shear planes. Mass flows resemble the flow of viscous fluids in which the maximum shearing occurs at the base and flow velocity decreases with increasing depth from the top of the flow. Most mass flows involve some amount of water, but large rock slides and falls sometimes transform into dry flows that run out long distances at the base of slopes. In such massive, high-energy flows, fluidity can be maintained by grain-to-grain collisions that promote efficient energy transfer. Such dry fluidization is invoked as the explanation for large flows on Mars, where liquid water is currently absent. Flows may move slowly (cm/day) to quite rapidly (m/second) and may involve soil or rock.

A **debris flow** is a rapid movement of saturated material, often channelized, down a steep slope. Debris flows are a slurry of soil, rock, and water that can travel far downslope from the point of failure initiation [**Photograph 5.13**]. Too much water, and the flow will separate into two phases, water and sediment, with the coarser component of the sediment settling to the bed. Too little water, and the flow will be too strong to move downslope. Fine-grained debris flows are often called mudflows. Some geomorphologists use the term **debris avalanche** to describe a very rapid flow of partially or fully saturated debris down a steep slope without confinement in an established channel or valley. These rapidly moving slope failures typically originate on steep bedrock slopes with thin soil cover and are thus common in formerly glaciated, alpine terrain where melting ice left debris on steep slopes.

Initiation of a debris flow generally requires at least a small landslide; thus, abundant groundwater, a steep slope,

PHOTOGRAPH 5.13 Debris Flow. The San Francisco earthquake of April 18, 1906, after heavy winter rains raised soil moisture levels, triggered this debris flow in Marin County, California.

and soil that is susceptible to shear failure (landsliding) are necessary ingredients. Soil properties greatly influence the potential for landslides to mobilize into debris flows because shearing or shaking within loose, sandy soil results in consolidation of the soil matrix that raises the pore pressure of interstitial fluids, triggers fluidization, and enhances mobility. In contrast, deformation of **dilative materials** that expand when sheared, like nonmarine clays, reduces pore pressures by expelling fluid, thereby limiting mobility because such deformation promotes drainage. Debris flows thus tend to form most frequently from landslides that occur in sandy soils with just enough fine material (5 to 20 percent) to retard drainage of interstitial fluids.

As they move downslope, debris flows generally entrain surficial materials like soil, unconsolidated sediment, and saprolite; they are a major sediment-transport process in mountainous landscapes. They typically begin on slopes between 26 and 45 degrees and have flowpaths that consist of a source area defined by the zone of initial slope failure, a scour and transport zone, and a depositional zone [**Figure 5.6**]. Debris flows can grow to more than 100 times the initial failure volume by scouring material from the base and edges of their runout paths—in many cases, steep headwater channels. Debris flows slow down and can then form depositional fans once they reach slopes of about 3 to 6 degrees.

Debris flows typically have a coarse, boulder-rich snout and sometimes deposit substantial levees along their runout path. Flowing material deforms rapidly and then deposits material abruptly when the driving stress (**shear stress**) falls below the yield strength of the flow. This abrupt change in deformation rate with applied stress is referred to as **plastic** behavior. Perfectly plastic materials are considered to have a finite yield strength. At shear stresses below the yield strength, they do not deform. If shear stress exceeds yield strength, plastic materials deform rapidly. A debris flow stops moving once it thins sufficiently or encounters a slope low enough that the yield strength exceeds the driving stress. The main bodies and tails of debris flows are often more fluid slurries than the coarse fronts, and debris flows commonly exhibit pulses because of variations in the fluid content within the flowing mass.

Debris-flow deposits are **matrix-supported;** clasts float in a finer-grain matrix and rock fragments are isolated from each other [**Photograph 5.14**]. A good analogy is chocolate chip cookie dough. Debris-flow deposits are readily distinguishable from those laid down by flowing water because fluvial deposits are **clast-supported;** individual particles rest in contact with one another, as would a pile of beans. Debris flows often come to rest on fans at the bottom of steep slopes where gradients are lower [**Photograph 5.15**].

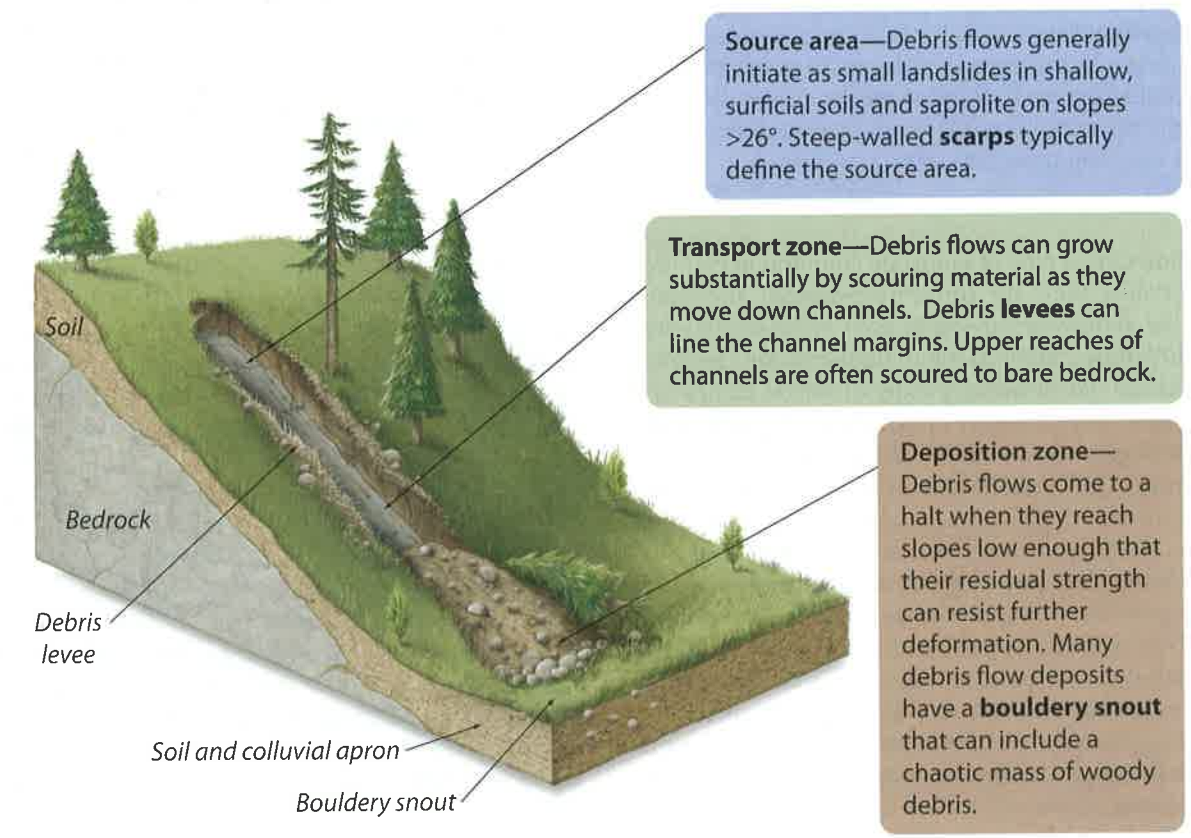
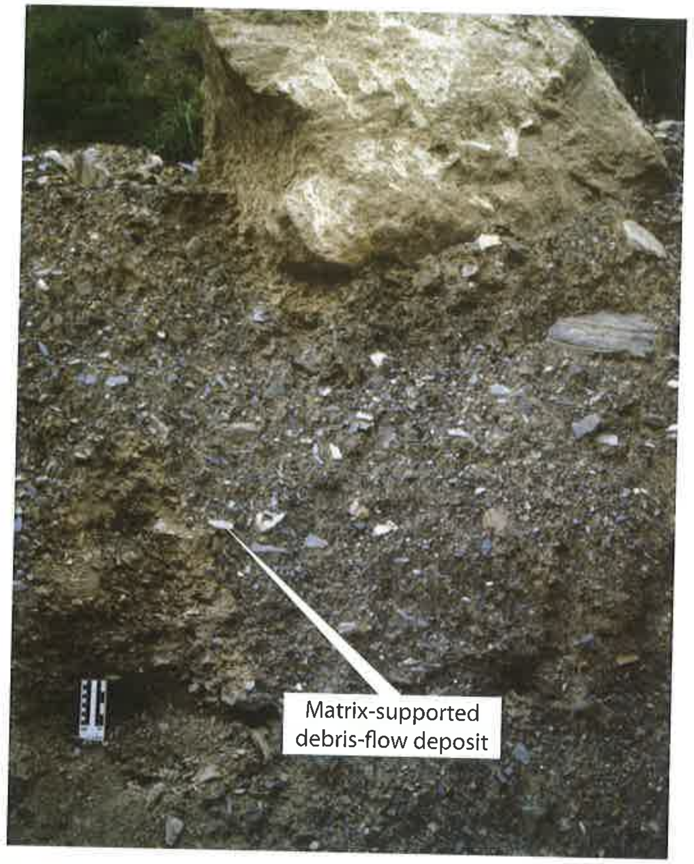


FIGURE 5.6 Debris-Flow Landforms. Debris flows leave characteristic clues on the landscape, including a source area (typically a landslide scar), a transport zone (an incised channel sometimes surrounded by levees), and a deposition zone that is often in the form of a fan.



J. Major/USGS

PHOTOGRAPH 5.14 Debris-Flow Deposit. Bouldery debris flow deposit in the Jiangjia River Valley of China, showing the matrix-supported nature of material deposited by the flow. The large boulder at the top of the photo was carried by the flow.

M. Howe/Sequoia National Forest

PHOTOGRAPH 5.15 Debris Fan. This debris fan of unconsolidated granitic sediment resulted from a debris flow that swept down a narrow canyon in the area denuded by a 2008 fire in Sequoia National Forest, California.

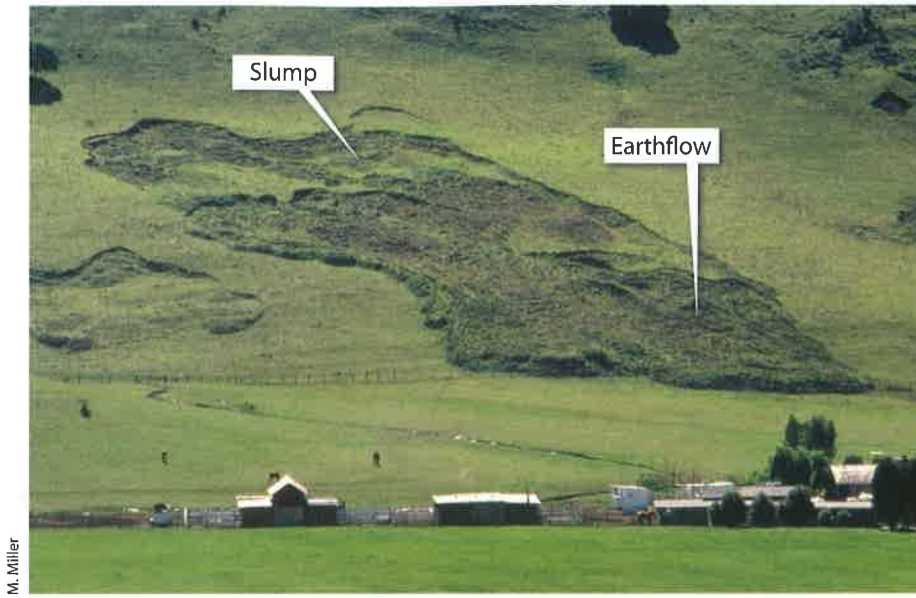
Earthflows are a type of landslide common in clay-rich earth materials; they are often deep-seated, and many originate as slumps involving masses of material that, unlike debris flows, move as rigid blocks within the failed mass. Although earthflows typically occur on slopes of 5 to 25 degrees, they can start on a variety of slope angles and exhibit highly variable mobility and speed, depending on the nature of the material involved (i.e., its unit weight and shear strength), the pore pressure, and geometry of the failure. Earthflows often consist of a complex of multiple flow lobes that may move at different speeds and at different times. They can move over periods of months to years and can episodically show distinct pulses of movement, with individual zones within a larger earthflow reactivating or moving at different paces [**Photograph 5.16**]. Excavation of material from their toes, ground shaking, or excessive rainfall can reactivate earthflows and accelerate their movement.

Rock flows involve deformation distributed among many through-going fractures within a rock mass. They may be extremely slow, deforming over long time spans, or quite rapid, as in the case of **rock avalanches**—massive, extremely rapid flows of fragmented rock that typically originate from a large rock slide or rock fall. Also commonly termed debris avalanches, these slope failures may run out long distances. One of the best-known examples of such a rock avalanche is the Blackhawk landslide in the Mojave Desert of southern California, which dropped more than a kilometer in elevation over its 9-kilometer runout [**Photograph 5.17**].

Falls

Falls begin with the detachment of soil or rock. They involve the downward motion of rock or soil through the air by falling, bouncing, or rolling, with little to no initial interaction with other materials. Falls typically occur on slopes that are steeper than the internal friction angle of the slope-forming material. Thus, falls are the most common mass movements on slopes that range from 45 to 90 degrees; in contrast, slides and flows are most common on slopes gentler than the friction angle (i.e., those < 45 degrees).

Falls tend to be rapid (m/s). Rock and soil falls typically occur when material is dislodged from very steep faces, like cliffs or streambanks. Rock falls are common on steep slopes in arid and alpine landscapes and are often triggered by earthquakes. Soil falls typically occur in places where flowing water or progressive slope failure undercuts cohesive soil or sediment, such as at gully heads or along incised streambanks. Slopes formed of tall blocks are especially prone to toppling, as the forward rotation of a soil or rock mass pivots about a point below its center of mass. **Topples** can range in speed from extremely slow (mm/yr) to extremely rapid (m/s).



M. Miller

PHOTOGRAPH 5.16 Slump and Earthflow. The scarp and bulging toe are clearly visible in this central California slump and earthflow.

K. Nichols

PHOTOGRAPH 5.17 The Blackhawk Landslide. This landslide in a Mojave Desert valley originated as a rock avalanche from the distant ridge and spread pulverized rock over the valley. The road in the foreground provides scale.

Slope Stability

Slope stability is analyzed by considering the balance between shear stress and the strength of earth materials. An examination of the ratio of driving forces to resisting forces per unit area reveals the relationships among the factors that resist and those that promote slope instability.

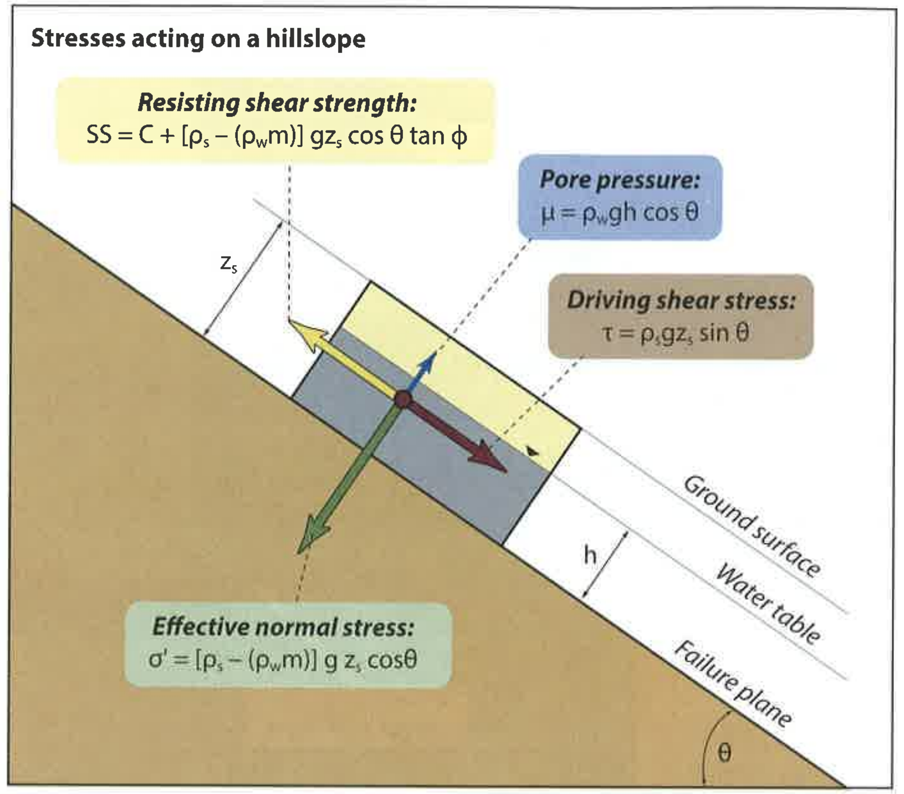
Driving and Resisting Stresses

The shear stress generated by the weight of the soil and rock overlying a unit area of a potential failure plane drives slope instability. Shear stress is defined as the downslope component (force per unit area) of the weight of the slope-forming material. If we consider the stresses acting on a hillslope, the shear stress (τ) acting on a planar surface covered by unconsolidated soil is given by

$$\tau = \rho_s \, g \, z_s \sin\theta \qquad \text{eq. 5.4}$$

where ρ_s is the average density of the soil, g is gravitational acceleration, z_s is the soil thickness measured perpendicular to (into) the slope, and θ is the slope angle [**Figure 5.7**]. In order to simplify the development of the infinite-slope model, we do not explicitly consider the changing density of the soil as it saturates with water during storms.

The shear strength of slope-forming materials (as expressed in eq. 5.1) resists the shear stress acting to move material downhill. Factors that decrease shear strength include weathering processes that weaken earth materials and increase pore pressure (μ), which increases buoyancy and thus acts to reduce the effective normal



Definition of terms:

z_s = slab thickness normal to failure plane
h = water-table height normal to failure plane
ρ_s = soil density
ρ_w = water density
g = gravitational acceleration
C = soil cohesion
m = h/z_s = proportion of soil slab that is saturated

θ = failure plane angle above horizontal
ϕ = angle of internal friction

σ = normal stress
μ = pore pressure
τ = shear stress
SS = shear strength

The **infinite-slope model** is often used to analyze the stability of slopes to shallow, planar landslides. The model is a balance between the **shear strength** of the slope materials and the **shear stress** provided by gravity, due to the downslope-oriented component of the mass of the soil. It considers the **pore pressure** effect on the stress balance as well as the effect of slope.

FIGURE 5.7 Infinite-Slope Model. The stability of planar landslides is commonly estimated using a force-balance approach (the infinite-slope model) to determine the factor of safety, the ratio between the resisting and driving forces per unit area. The analysis considers soil density, soil saturation, slide-block thickness, and the slope of the failure plane.

stress ($\sigma' = \sigma - \mu$). The hydrostatic pore pressure is expressed in terms of the height of the water table above the slide plane (h):

$$\mu = \rho_w \, g \, h \cos\theta \qquad \text{eq. 5.5}$$

It is worth noting that if we cast this equation (and subsequent equations in this section) in terms of vertical soil thickness, we would introduce an additional cos θ term on the right-hand side of eq. 5.4 and 5.5 (and thus on the trigonometric terms in the numerator and denominator of eq. 5.8). Substituting eq. 5.2 and 5.5 into eq. 5.3 yields the effective normal stress for a partially saturated slope:

$$\sigma' = (\rho_s \, g \, z_s \cos\theta) - (\rho_w \, g \, h \cos\theta) \qquad \text{eq. 5.6}$$

After collecting and simplifying terms this reduces to

$$\sigma' = [\rho_s - (\rho_w \, m)] \, g \, z_s \cos\theta \qquad \text{eq. 5.7}$$

where m is the ratio of the saturated soil thickness (h) to the total soil depth (z_s) above the slide plane that is saturated ($m = h/z_s$). Expressing the relative soil saturation as the ratio m allows for consideration of the effect of variations in moisture on the potential for slope failure. If the soil is fully saturated, and the water table is at the land surface, m = 1. If the water table is below the potential failure surface, or if the soil is dry, then m = 0. Slope stability models typically portray relative soil saturation as a single value, but direct measurements of pore-water pressures during landslide-inducing storms show that variations in soil properties, the presence of macropores, and groundwater recharge and discharge (springs) at the base of the soil profile impart significant spatial variability to soil moisture.

Infinite-Slope Model

The simplest physical model of slope stability is the **infinite-slope model.** It considers the balance of forces at a point on a potential failure plane that is assumed to have a constant slope (θ) and extend infinitely at its margins to avoid edge effects. The infinite-slope model effectively reduces the complex three-dimensional problem of slope failure to a two-dimensional problem by ignoring the effects of lateral slide boundaries or along-slope variations in soil depth and slope angle. This formulation greatly simplifies the analysis of the factors that contribute to slope instability, while still allowing for meaningful quantitative assessment of slope properties. In particular, the infinite-slope model allows stability to be assessed for a unit area of the slope as representative of all the unit areas around it.

At the core of the infinite-slope model is the **factor of safety** (FS), which is the ratio of resisting shear strength to the driving shear stress (FS = SS/τ), and it equals 1 at slope failure. Stable slopes have FS > 1, and unstable or failed slopes have (or had) FS ≤ 1. The infinite-slope model is derived by substituting the definitions of the shear strength ($SS = C + \sigma' \tan\phi$: eq. 5.1) and shear stress ($\tau = \rho_s \, g \, z_s \sin\theta$: eq. 5.4) into the definition of the factor of safety (FS = SS/τ). From there, we further substitute the expression for the effective normal stress (eq. 5.7) in the numerator and arrive at:

$$FS = \frac{C + [\rho_s - (\rho_w \, m)] \, g \, z_s \cos\theta \tan\phi}{\rho_s \, g \, z_s \sin\theta} \qquad \text{eq. 5.8}$$

This equation illustrates the basic controls on slope stability and shows all the key factors involved in slope stability: the material properties (ρ_s, ρ_w, C, and ϕ), the geomorphic properties of soil depth and ground slope (z_s and θ), and the time-varying environmental factor, that is, the proportion of the soil thickness that is saturated (m), or the pore-water pressure (μ; see eq. 5.5), and gravitational acceleration (g), which is important in the case of earthquake-induced landslides.

Not only is the infinite-slope model useful for predicting slope failures, it can also be used to understand landscapes. Given the characteristic strength parameters (C and ϕ) for the earth materials underlying a region and assuming groundwater levels, one can predict the maximum stable angle of hillslopes. In the simplest example, the maximum stable slope angle (FS = 1.0) for dry cohesionless material like sand (C = 0 and m = 0) equals the friction angle of the slope-forming material (tan θ = tan ϕ). As cohesion values increase, so will the steepness of slopes the material can support. Conversely, the wetter the slopes (higher water table), the less steep the slopes will be.

Rotational failures can be modeled using a simple adaptation of the infinite-slope approach. In the **method of slices,** the failure surface is defined as a series of chords that collectively approximate the arc of a circle. The analysis then divides the circular arc of the failure surface into a series of slices, each a segment of the arc and each having an average vertical soil thickness. Individual stability analyses for each slice can then be summed to assess the overall stability of the slope based on the geometry and material properties (ρ_s, C, and ϕ) of the slope [Figure 5.8]. For rotational slides, this method improves on the infinite-slope model because it includes the geometry of the toe of the slide, where the rotation surface curves upward and acts to resist stresses generated upslope.

Environmental and Time-Dependent Effects

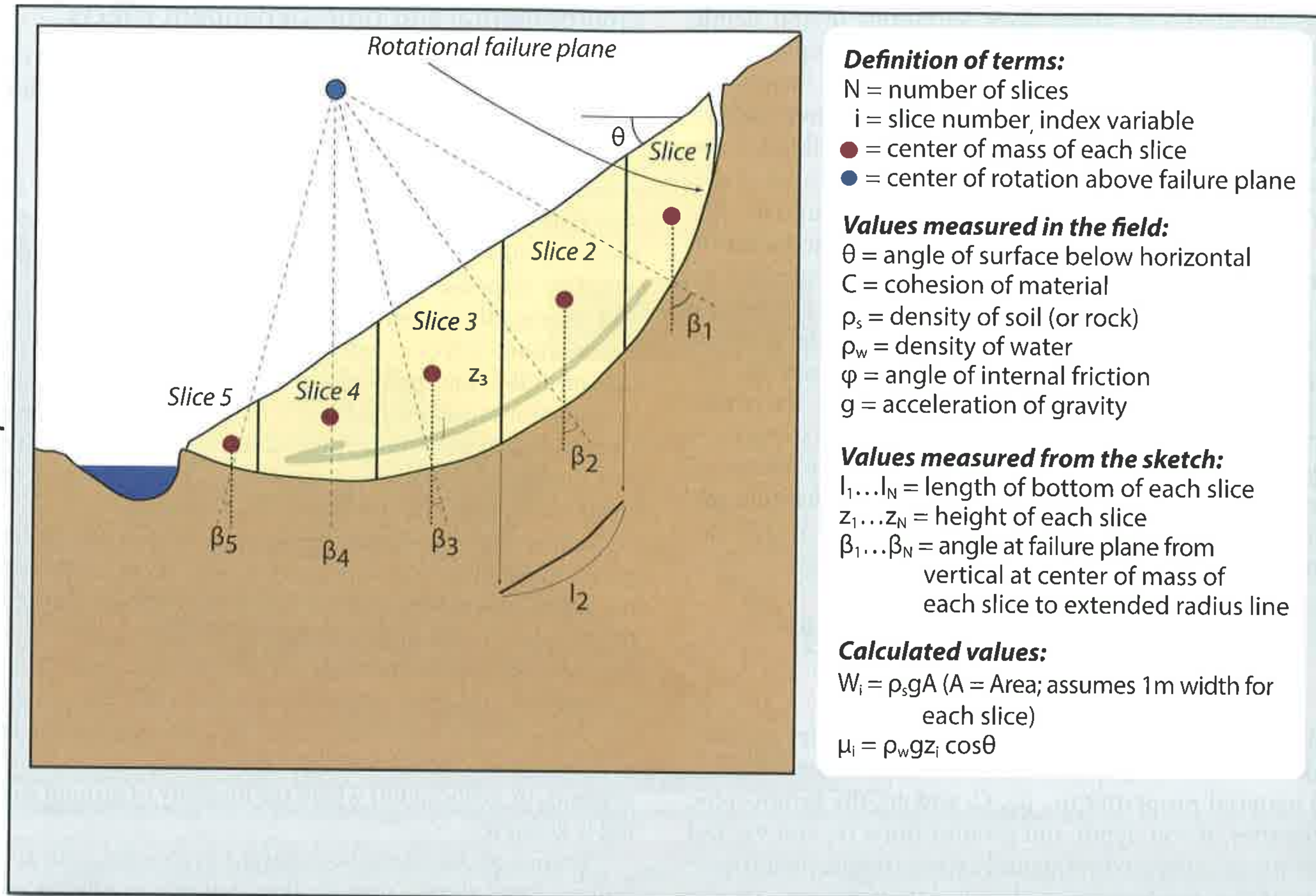
Slope failure is a good example of a geomorphic threshold. When the material on a formerly stable slope slides downhill, it means that something changed to upset the preexisting relationship between driving stress and the resisting material strength. A variety of environmental factors that change over short and long timescales influence slope stability. In the short term, slope failure can be triggered by response to rainfall (change in pore pressure, μ) and seismically driven forcing (change in g, gravitational acceleration). Over decadal time frames, vegetation succession following disturbance (and the associated decline in apparent cohesion from root reinforcement) influences slope stability. Over millennial timescales, the evolution of soil thickness and slope morphology change normal and shear stresses and thus the stability of slopes.

Landslides on potentially unstable slopes can be triggered by earthquake shaking. Ground acceleration during earthquakes destabilizes slope material by changing the effective value of g. Both the intensity of ground shaking and the horizontal and vertical acceleration depend strongly on material properties of the soil. The greatest number of slides typically occurs near the earthquake epicenter and in weak materials (such as alluvium, marsh deposits, or artificial fill) where the intensity of ground shaking is greatest.

Studies of the rainfall conditions associated with slope failures have shown that shallow failures in hillside soils tend to occur once rainfall exceeds a quantifiable intensity-duration threshold that varies by region (see Figure 4.1). During storms, pore pressures increase as soil below the hillside becomes saturated, thereby reducing the effective normal stress and decreasing the shear strength. More intense or longer-duration rainfall translates into a greater saturated thickness and therefore greater pore pressures and less stability. In other words, slides tend to occur once it rains hard enough for long enough to sufficiently saturate hillslopes. Deep-seated landslides may move long after rainfall ceases, or in response to seasonal changes in the water table because of the time required for water to infiltrate to and thus raise the water table.

The apparent cohesion that plant roots contribute to soils differs greatly among species and changes as vegetation grows, matures, and dies [Table 5.2]. Consequently, plant succession or a change in community structure, like conversion of forest to grassland, can cause substantial changes in slope stability. Grassland soils are also stabilized by root reinforcement, and the stability of grass-covered hillslopes has been negatively impacted by the replacement of deep-rooted native grasses with invasive shallow-rooted grasses (as happened both in California and in the Great Plains of North America).

Among trees, conifers generally contribute the most apparent cohesion to soils. Following forest fires or timber harvest, roots die and it takes about a decade for them to rot completely and lose all strength. Thus, there is usually a period of at least several years during which root strength



Model

$$\textit{Factor of safety} = \frac{\textit{Resisting}}{\textit{Driving}} = \frac{\sum_{i=1}^{N} [C_i l_i + (W_i \cos \beta_i - \mu_i l_i) \tan \varphi_i]}{\sum_{i=1}^{N} W_i \sin \beta_i}$$

FS > 1 Stable

FS < 1 Failure

The **method of slices** is a useful way to analyze the stability of a **rotational landslide**. The method approximates the **force balance** of slides with circular failure surfaces by approximating the failure surface as a series of chords and solving the force balance for each of these subsections. Summing the results determines the overall stability of the slide mass.

FIGURE 5.8 Method of Slices. The stability of rotational failures is modeled using a permutation of the infinite-slope model, referred to as the method of slices. This approach considers the slide surface as a series of slices.

remains low as root networks of newly planted or established trees regrow. During this window of low root strength, potentially unstable slopes are particularly vulnerable to slope failure (see Figure DD5.4). If slopes are kept clear, as they were in New England during most of the 1800s, they remain susceptible to failure during soaking rain and spring snowmelt. During periods of low root strength, more landslides occur in response to rainfall events than would have prior to vegetation clearing, or under mature forest.

In addition to altering or clearing vegetation, human land use affects slope stability by changing hydrologic pathways, such as when urbanization or road construction concentrates and delivers runoff to steep slopes [Photograph 5.18]. The effect is even more dramatic when rainwater runoff is routed to topographic concavities, known as **hollows**, where colluvial soil accumulation and both surface and shallow subsurface runoff are focused (see Photograph 7.9). Shallow, planar failures often start in hillslope hollows

TABLE 5.2

Typical values of apparent cohesion associated with vegetation

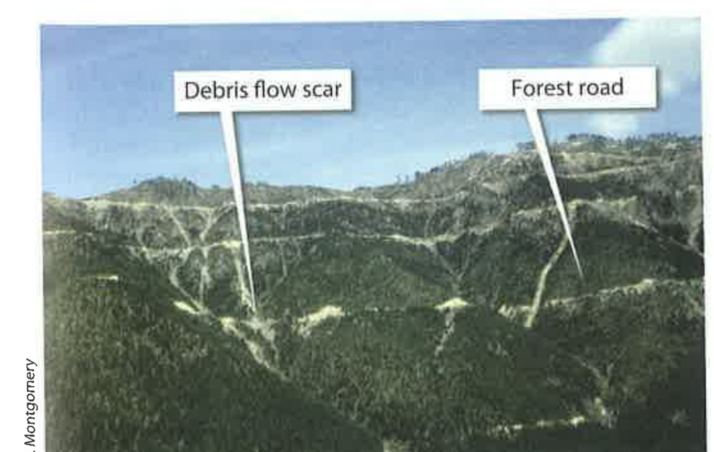
Vegetation	Apparent cohesion (kPa*)
Grass	0–1
Stumps	0–2
Chaparral	0–3
Hardwoods	2–13
Conifers	3–20

*kPa = kg/ms^2

because these parts of the landscape favor both the accumulation of thicker soils and greater soil saturation.

Although landslides tend to be triggered by destabilizing events, the potential for slope failure changes gradually over centuries to millennia as the hydrology and strength properties of a slope change. Once-stable slopes can become unstable and slopes near the stability limit can become more stable. Over time, soil thicknesses (z_s) and slope angles (θ) change. For example, when processes like tectonic tilting and stream incision cause a slope to steepen, the slope's factor of safety decreases. Soil depth increases over time as colluvium fills in topographic hollows, resulting first in decreased stability as soil depth exceeds tree-root depth; then, strength grows as clay content in older soils increases, and the thickness of fill prevents the hollow from saturating completely during storms. Consequently, the slope grows stronger and landslides are less likely.

Simple slope-stability models like eq. 5.8 can generally predict failure of steep slopes with wet, thick soils, but the spatial variability in cohesion (C), soil thickness (z_s), and soil saturation (m) complicates predictions of landslide initiation in particular locations in response to specific



PHOTOGRAPH 5.18 Debris Flows from Forest Roads. Debris flows originate on a series of forest roads and carry sediment downslope to the Tolt River in the Cascade Mountains, Washington State.

storm events. Regional landslide hazard assessments and inventories typically find that only a fraction of potentially unstable slopes fail during any single storm. The number of slope failures generally increases with storm size and duration, but accurately identifying individual slopes that will fail in a given storm is not yet possible. Model-based assessments of slope stability are thus best viewed as assessing the generalized probability of slope failure.

Slope Morphology

Slope morphologies reflect slope-shaping processes. In profile, hillslopes may be convex, straight, or concave [Figure 5.9]. In humid and temperate regions, soil-mantled slopes generally steepen downslope from convex ridgetops at drainage divides to a planar midslope segment with a constant angle and a concave, less steep basal zone at the toe of the slope. Straight portions of slope profiles are typically more pronounced in steep terrain where frequent landslides plane off topography in the midslope zone. In contrast, arid slopes in areas of high relief often have a vertical cliff face downslope of the ridgetop and exhibit slope variations that are controlled by the relative strength of the bedrock, with harder rocks holding up steeper slopes and weaker rocks supporting gentler slopes. Such bedrock differences are apparent when the rock is not mantled by soil or regolith.

In map view, slopes are described as convergent, divergent, or planar. In general, ridgelines are divergent, and valleys are convergent. Planar slopes are neither convergent nor divergent. On convergent slopes, flow lines converge downslope. On divergent slopes, flow lines diverge downslope. Two adjacent balls rolling down a convergent slope would be at risk of colliding. They would move farther apart rolling down a divergent slope. Likewise, sediment moving down slopes tends to accumulate in convergent valleys over time.

Slope processes generate divergent, planar, or convergent topography, depending on the predominant erosional and depositional mechanisms. The diffusion-like action of creep and sheetwash creates zones of hilltop convexity and divergent noses on soil-covered hillsides. Divergent slopes are farthest from streams, and broad, rolling hills generally reflect slow uplift and erosion rates. In contrast, steep slopes and deeply incised valleys suggest more rapid uplift and erosion. Long, planar slope segments with only short hilltop convexities are common in rapidly eroding terrain and reflect the influence of landslides where the relief and hillslope gradient are high.

Weathering-Limited (Bedrock) Slopes

Weathering and sediment transport both affect slope morphology. At one extreme, **weathering-limited slopes** (also called production-limited slopes) have net rates of soil transport that are determined by the rate at which weathering

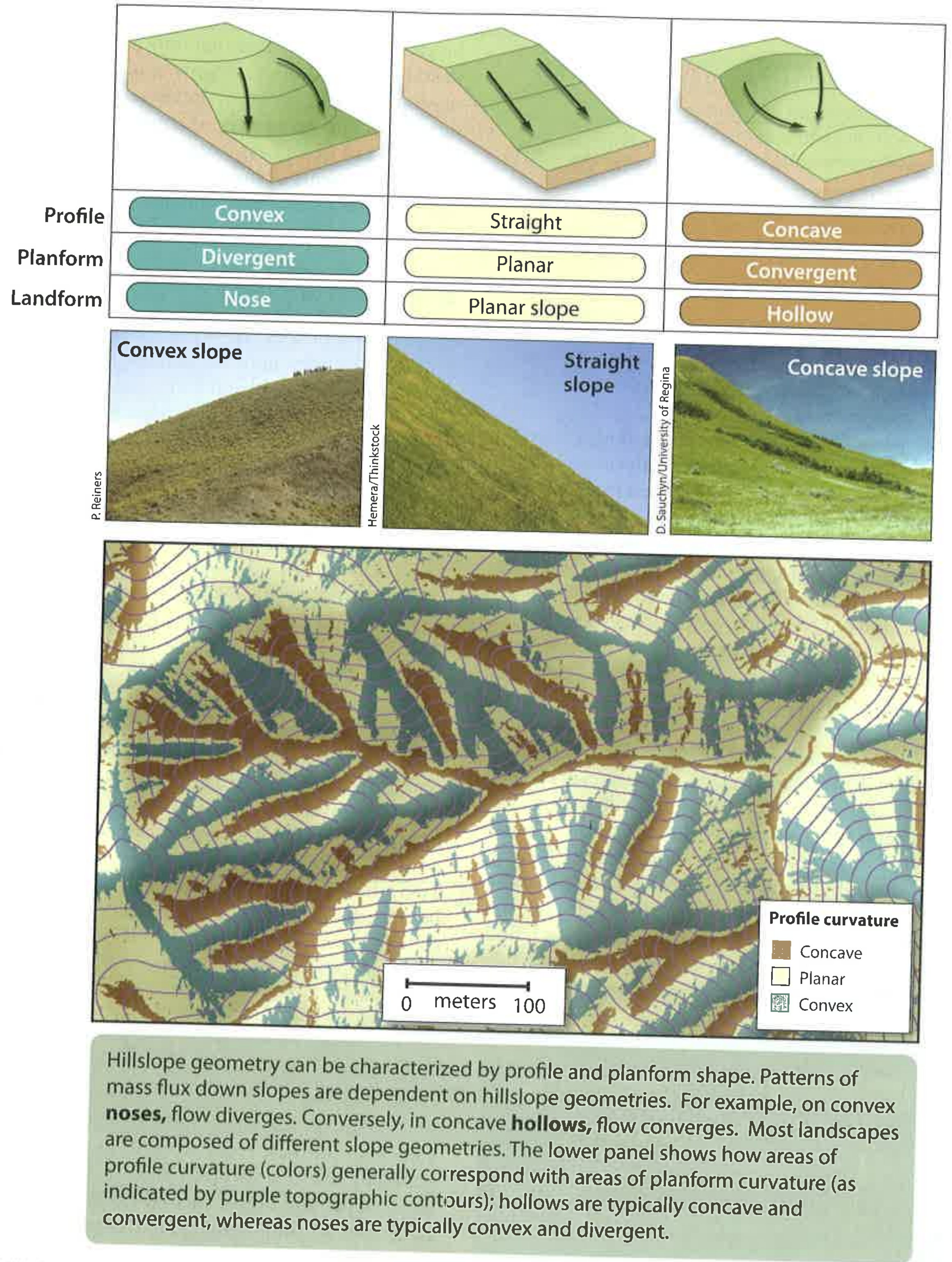
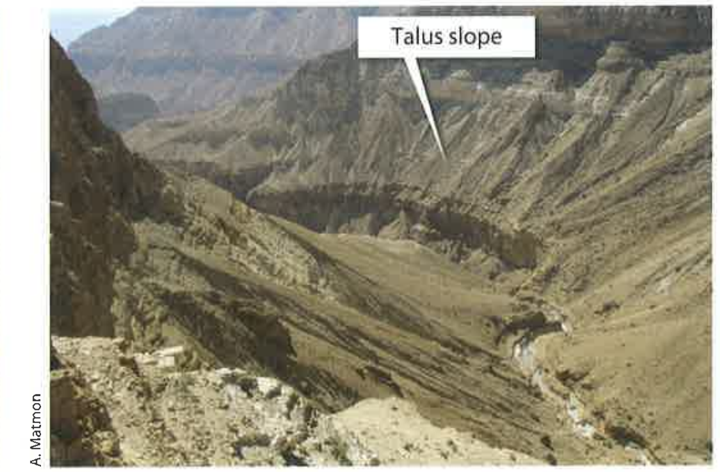


FIGURE 5.9 Hillslope Geometries. Hillslopes can be grouped into three distinct geometries. Convex slopes are divergent and found on noses or ridgelines. Concave slopes are convergent and make up valleys. Planar slopes are straight and often reflect slopes steep enough to fail by mass movements.

provides new material [Photograph 5.19]. On these slopes, rates of mass removal by erosion generally match rates of regolith production, so weathered material is transported downslope about as rapidly as it is produced. These slopes tend to be bedrock with thin, discontinuous soils. The steepness of weathering-limited slopes is generally controlled by rock mass strength, and slope morphology is often closely related to the underlying rock type. Weathering-limited slopes are commonly found in arid regions where chemical weathering and rock detachment rates are

PHOTOGRAPH 5.19 Bedrock Hillslope. Bare rock hillslope (weathering limited) at the Nahal Yael experimental watershed in the Negev Desert, Israel. Here, runoff from the bare granite hillslope is funneled into a weir so that it can be measured.



PHOTOGRAPH 5.20 Talus Slopes. These talus slopes, which originate from the cliffs above, line a canyon in the Judean Desert, Israel.

slow, as well as in arctic and high alpine areas where strong, bare rock slopes were steepened and stripped of previously weathered rock by glaciers.

The properties of bedrock slopes with little to no soil are strongly influenced by the material and strength properties of the bedrock itself. On many bedrock hillslopes, the angle of the slope depends on the resistance of the underlying bedrock to collapse or disintegration. Harder rocks form steeper slopes; weaker rocks form gentler slopes. Where slopes are formed by interbedded strong and weak rocks (sandstone and shale, for example), variability in erosion resistance commonly produces the stair-stepped morphology typical of arid and semi-arid landscapes. The walls of the Grand Canyon provide a classic illustration of this morphology (see Photograph 1.3). Rock slopes are typical of arid and semi-arid landscapes but generally are found only in the steeper parts of temperate and humid landscapes because weathering maintains a soil mantle on most lower-gradient slopes.

In places where the sediment supply from a bedrock cliff face exceeds the rate of removal by streams, angular fragments of rock debris, called **scree**, accumulate and pile up at the base of the cliff. Eventually, such material builds up a **talus slope** or ramp that progressively buries the cliff under its own debris [Photograph 5.20]. Because scree falls from the cliff face above, the size of individual blocks primarily reflects the structural characteristics and fracture density of the parent rock.

Transport-Limited (Soil-Mantled) Slopes

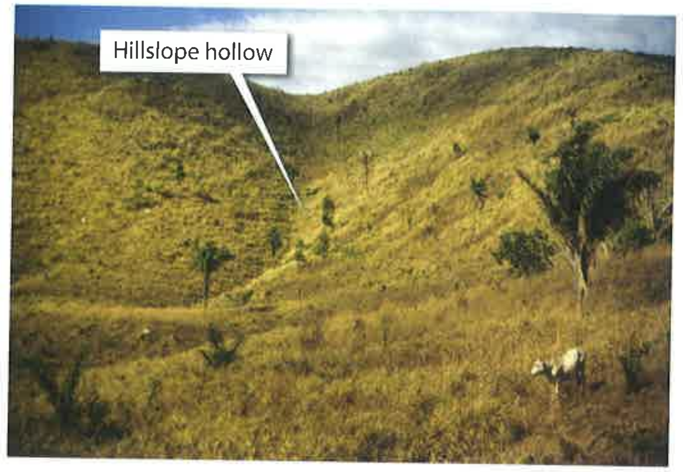
Transport-limited slopes have a supply of readily transportable material at the surface and generally have soil production rates that equal or exceed rates of downslope soil transport, a condition that results in the development of a persistent, continuous soil mantle. There is enough transportable material available on these slopes that the transport capacity of hillslope processes governs the rate at which soil leaves the slope. The morphology of transport-limited slopes is strongly controlled by soil properties. The smoothly convex to planar form of soil-mantled slopes typically masks variations in the underlying rock type as well as in bedrock structures like folds, joint patterns, and fault scarps.

Soil-mantled slopes generally steepen downhill, with a greater sediment flux corresponding to the increasing slope. On soil-mantled, transport-limited slopes, rates of soil movement are generally considered to be a simple linear function of slope steepness, θ, so the volumetric flux rate, q_s, increases with increasing slope, θ, as governed by a hillslope erosion (or diffusivity) rate constant, k, the value of which reflects soil type, vegetation, and climate through

$$q_s = k \tan\theta \qquad \text{eq. 5.9}$$

The flux of sediment moving downslope grows as one moves from the ridge crest to the valley bottom. Increasing soil flux downslope is the inevitable outcome of mass conservation, as each bit of hillslope above the valley bottom contributes mass that moves downslope under the influence of gravity. Hence, unless the shape of the slope is changing with time (from aggradation), slope angle must increase downslope in order to accommodate the downslope increase of soil flux by diffusive processes (as θ increases in eq. 5.9, so does q_s). Increasing slope with distance downslope by definition leads to hillslope convexity [Photograph 5.21].

If rock-uplift rates are uniform over time, soil-mantled slopes can reach an equilibrium profile that reflects a balance between soil-formation rates and soil-transport rates.



D. Montgomery

PHOTOGRAPH 5.21 Convex Hillslopes. Grass-covered convex hillslopes in Brazil with a cow in foreground and hillslope hollow (unchanneled valley) in distance.

I. Larsen

PHOTOGRAPH 5.22 Bedrock Landslide. Large bedrock landslide on a steep, threshold slope in the Tien Shan mountains, Kyrgyzstan.

In such a case, slopes will maintain a convex profile. Indeed, the convexity of drainage divides can be used to define the zone of diffusion-dominated erosion and sediment transport near those divides. For the case of an equilibrium hillslope with a uniform soil thickness eroding at the same rate, the convex hillslope form can be derived analytically [Box 5.1].

Threshold Slopes

On slopes that have steepened to the point where landslides are the dominant erosional mechanism, river incision, rather than further slope steepening, triggers additional landsliding. Consequently, once slopes reach an upper, limiting angle between about 30 and 40 degrees (depending on soil and rock strength), they become **threshold slopes** on which landslide frequency, instead of slope angle, controls slope erosion rates [**Photograph 5.22**]. At slope angles above about 26 degrees, the relationship between slope angle and hillslope erosion rate thus changes from the linear relationship common on low-gradient slopes to a nonlinear, asymptotic relationship [**Figure 5.10**].

Slopes with gradients less than about 26 degrees, a lower limit for the initiation of most debris flows and shallow soil landslides, tend to erode and evolve through slope-dependent sediment transport in which the pace of soil erosion increases linearly with slope. Development of threshold slopes as hillslopes approach a **critical angle,** the angle at which landsliding is both frequent and the dominant means of mass transport, indicates that in steep terrain, erosion rates can increase greatly with little change in slope. Thus, it becomes increasingly difficult to infer erosion rates from slope angles as slope steepness approaches the upper limiting angle set by the soil or rock strength. The development of threshold slopes at angles close to the upper limiting angle is a major reason why mountainous topography looks remarkably similar across a wide range of climates and tectonic settings worldwide.

Hillslope Evolution

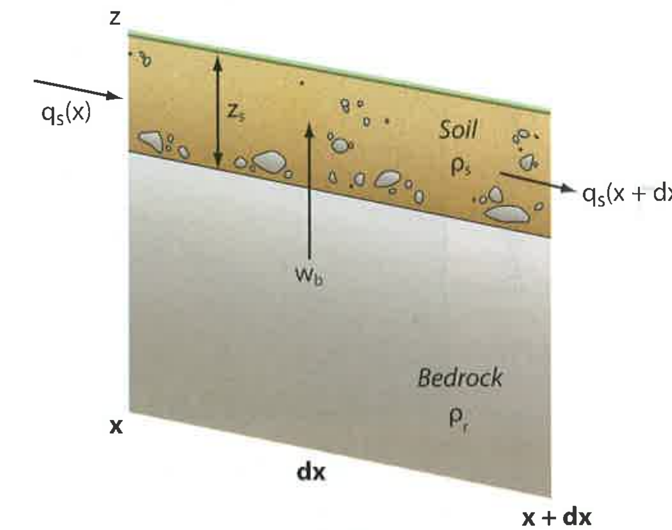
Geomorphologists have developed three general models of slope-profile evolution to explain how slope morphologies change over time: slope replacement, parallel retreat, and slope decline [**Figure 5.11**]. In **slope replacement,** which occurs in all climates, the gentler profile of the lower slope segment extends uphill as the steeper upper angle (or cliff face) retreats and a talus apron accumulates. This scenario occurs when material shed from a cliff face is not removed from its base. During **parallel retreat,** a slope maintains a constant angle as erosional retreat of the ground surface cuts back into the bedrock. **Slope decline** occurs as steep slopes gradually become shallower, flattening the overall slope profile. The end result of slope decline is a profile with a convex upper slope segment above a concave lower segment that develops where material is deposited near the base of the slope. All three modes of slope retreat occur in nature. Slope decline generally characterizes profile evolution in temperate and humid regions with soil-mantled slopes, and slope replacement and parallel retreat are more common in arid and semi-arid regions with bedrock slopes.

Drainage Density

Drainage density quantitatively describes the degree to which a landscape is dissected by stream channels and their valleys [**Photograph 5.23**]. The degree of topographic dissection of a landscape is related to the distance that channels begin downslope from drainage divides. Drainage density (DD) equals the total channel length (ΣL) within a given drainage area (A):

$$DD = \Sigma L/A \qquad \text{eq. 5.10}$$

BOX 5.1 Derivation of the Form of Convex Hillslope Profiles



Combining the concepts of conservation of mass and a simple, slope-dependent soil transport rate, we can derive the expected form of a soil-mantled hillslope. Based on the assumption of conservation of mass, the relationship between the change in soil thickness z_s over time (expressed as the derivative of soil thickness with respect to time, dz_s/dt) and the rate of bedrock weathering (W_b), the density of rock and soil (ρ_r and ρ_s, respectively), and the downslope change in the transport of soil (q_s) may be expressed as a function of the distance downslope (x):

$$dz_s/dt = (\rho_r/\rho_s)\,W_b - (1/\rho_s)\,(dq_s/dx) \qquad \text{eq. 5.A}$$

In other words, the rate at which the soil thickness changes (dz_s/dt) will be equal to the difference between the rate of bedrock weathering (adjusted for the difference in bulk density between rock and soil; the first term on the right-hand side of the equation) and the downslope change in the rate of soil transport (dq_s/dx, the spatial derivative of the sediment flux; the second term on the right-hand side of the equation).

For the case of steady-state soil thickness for which the soil thickness does not change over time (i.e., $dz_s/dt = 0$), eq. 5.A reduces to

$$\rho_r\, W_b = dq_s/dx \qquad \text{eq. 5.B}$$

which indicates that the local soil production ($\rho_r W_b$) is balanced by the downslope change in soil transport (dq_s/dx). Integrating eq. 5.B with respect to x results in an expression for the soil flux as a function of position on the hillslope:

$$q_s = \rho_r\, W_b\, x \qquad \text{eq. 5.C}$$

which states that the soil flux increases linearly with distance downslope.

Recasting the equation for slope-dependent hillslope transport (eq. 5.9) with a negative sign to reflect the convention of using the ridgetop elevation as a benchmark yields

$$q_s = -k\; dz/dx \qquad \text{eq. 5.D}$$

where q_s is the soil transport (soil flux), k scales the transport efficiency, and dz/dx is the ground slope, the change in elevation with distance downslope. Substituting eq. 5.D into eq. 5.C reveals that the slope (dz/dx) increases with distance (x), resulting in a convex slope profile.

An expression for the form of the slope profile may be derived by substituting the expression for q_s in eq. 5.D into eq. 5.A, which yields a form of the standard one-dimensional diffusion equation:

$$dz_s/dt = (\rho_r/\rho_s)W_b + (k/\rho_s)\,(d^2z/dx^2) \qquad \text{eq. 5.E}$$

where the ratio k/ρ_s is the landscape diffusivity, which may be expressed as $K_d = k/\rho_s$. For the case of a spatially uniform hillslope lowering rate and constant soil thickness (i.e., a steady state topographic form where $dz_s/dt = 0$), eq. 5.E reduces to

$$(d^2z/dx^2) = -(\rho_r\, W_b/K_d\, \rho_s) \qquad \text{eq. 5.F}$$

This expression shows that the hillslope curvature, the second derivative of elevation as a function of distance, is a constant set by the rate of soil production (W_b) and the landscape diffusivity (K_d). Integrating this expression once yields the topographic slope (d_z/d_x) as a function of position along the slope (x):

$$(d_z/d_x) = -(\rho_r\, W_b/K_d\, \rho_s)x + c_1 \qquad \text{eq. 5.G}$$

where c_1 is a constant of integration, which equals zero because the slope (d_z/d_x) is zero at the hilltop ($x = 0$). Integrating this expression once again yields the shape of the hillslope profile expressed through the relationship between hillslope elevation (z) and distance downslope (x):

$$z = -(\rho_r\, W_b/2\, K_d\, \rho_s)x^2 + c_2 \qquad \text{eq. 5.H}$$

in which c_2 (another constant of integration) is equal to the hillslope elevation at the ridgecrest, z_{max} (the value of z at $x = 0$) from where elevation decreases as a function of the square of the distance downslope. This yields the functional form of an inverted (convex-up) parabola:

$$z = z_{max} - (\rho_r\, W_b/2\, K_d\, \rho_s)x^2 \qquad \text{eq. 5.I}$$

While the functional form of this equation closely fits the convex profile of diffusion-dominated hillslopes, the hillslope curvature and the width of the zone of convexity vary with differences in soil production (and thus the bedrock erosion) rate and landscape diffusivity.

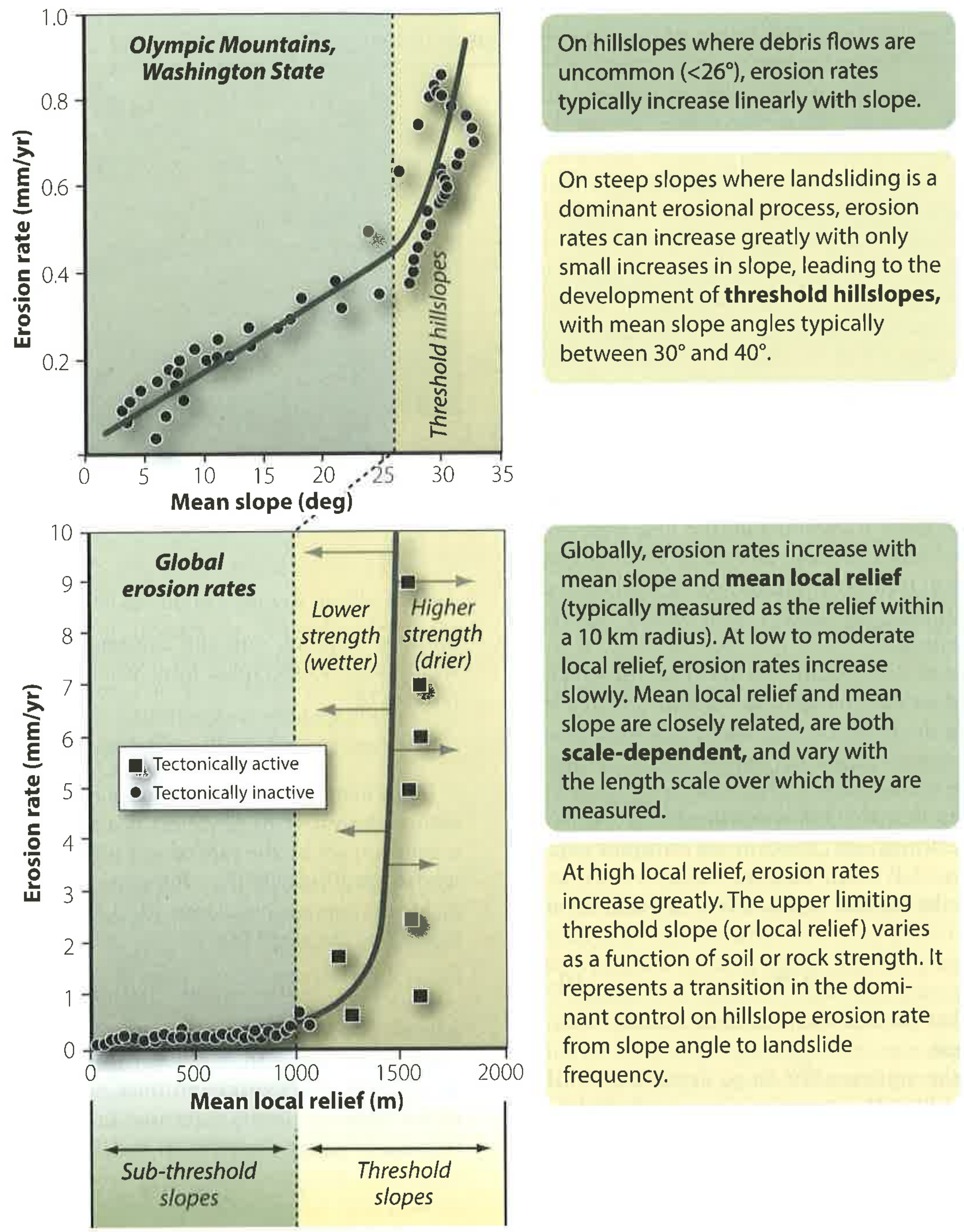


FIGURE 5.10 Threshold Hillslopes. Threshold slopes are steep and transport of mass from them is dominated by landsliding. The slope angle of the transition to threshold slopes reflects the strength of the underlying rock. [Adapted from Montgomery and Brandon (2002).]

Terrestrial drainage densities typically range from 1 to 100 km/km^2, and the length of unchanneled slopes upslope of well-defined stream channels ranges from less than a meter on some intensively rilled badlands to more than a kilometer on broad, low-gradient hillsides in semi-arid landscapes developed on cratons (such as in Kenya, eastern Wyoming, and the Amazon Basin). In disturbed areas and sparsely vegetated arid terrain, channel heads may extend as rills above unchanneled valleys onto the surrounding side slopes.

The drainage density within a watershed depends on the underlying lithology and the length of time over which dissection has occurred. In highly dissected watersheds, typical of shales and other easily erodible bedrock, drainage density can be quite high. This degree of dissection has important hydrologic manifestations, as it is one determinant

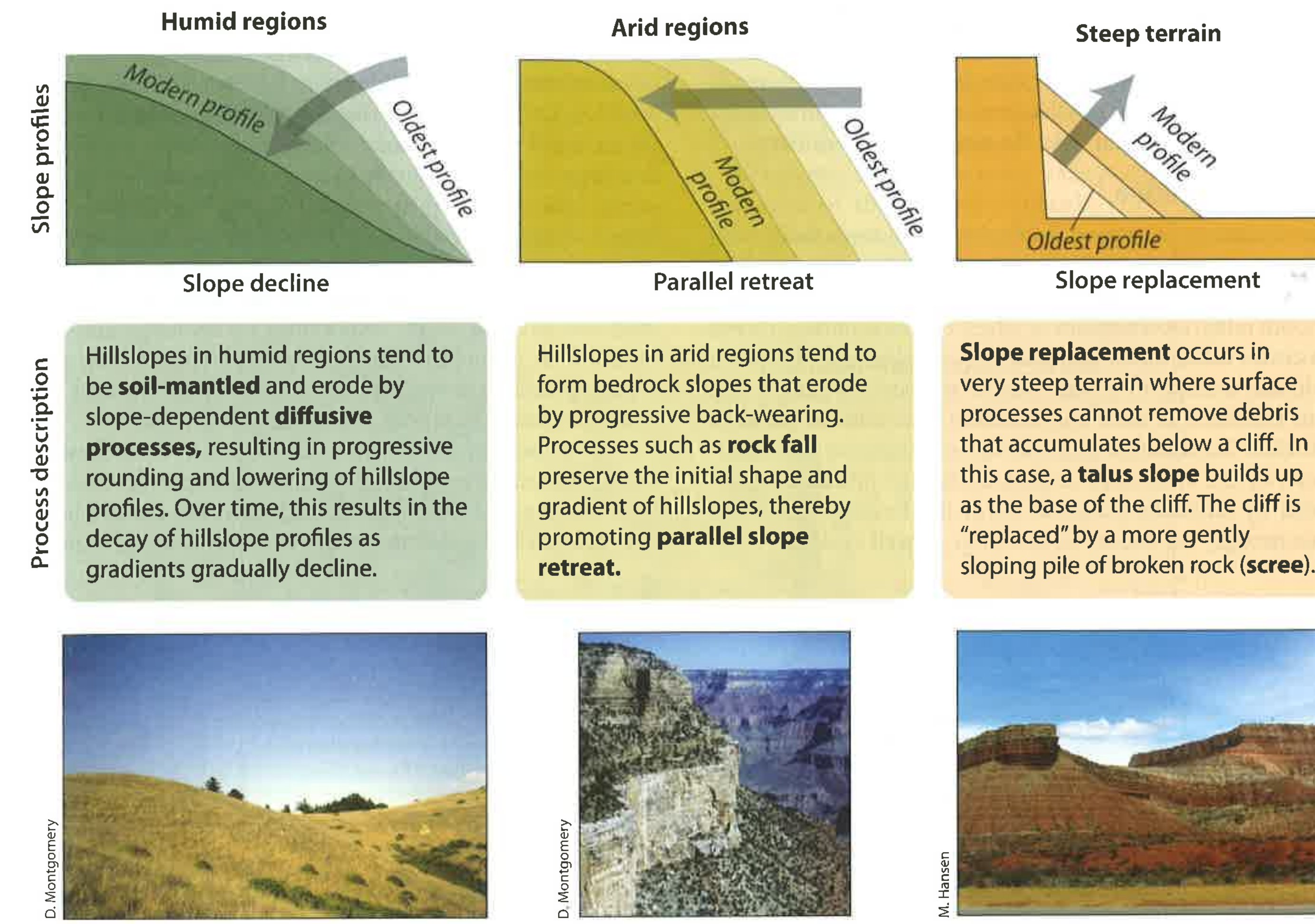
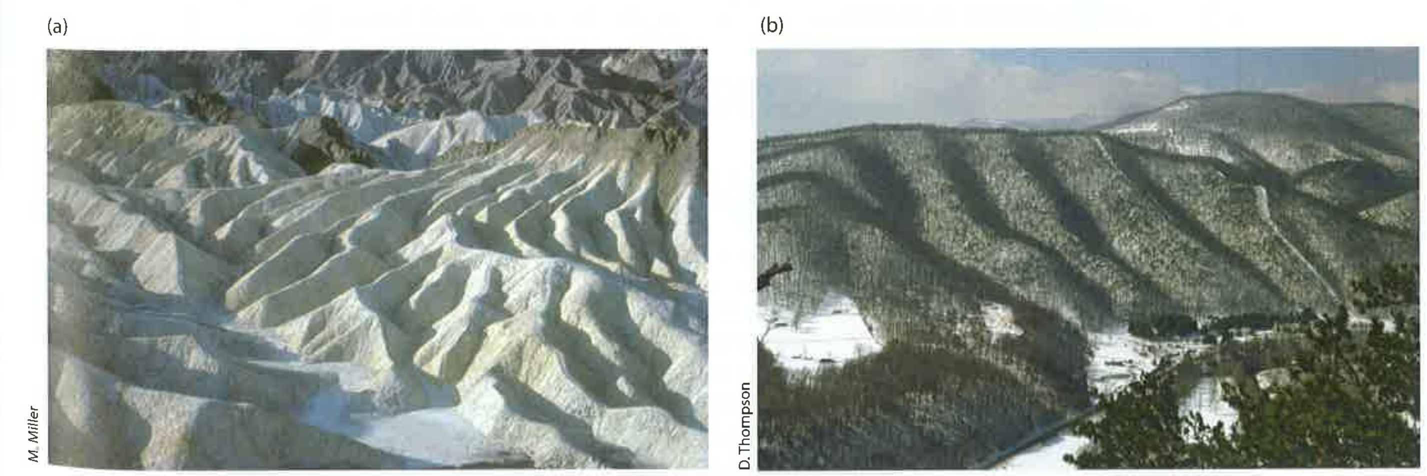


FIGURE 5.11 Slope Evolution. Slopes can evolve over time by parallel retreat, common in arid regions, or by declining steepness, more common in humid regions.



PHOTOGRAPH 5.23 Drainage Density. (a) High drainage density of the weak, nonvegetated badlands landscape eroded into poorly consolidated lakebed deposits, at Zabriskie Point, Death Valley, California. Spacing between the finest valleys shown is about 10 m. (b) Lower drainage density shown in the tree-covered, humid-temperate landscape of the southern Appalachian Mountains in West Virginia, an area underlain by strong, metamorphic rocks.

of the shape of the hydrograph. High drainage densities efficiently move water off slopes and into channels; therefore, where drainage density is high, the hydrograph tends to have a higher peak discharge and shorter lag times relative to watersheds that have lower drainage densities.

Channel Initiation

The initiation of stream channels on hillslopes marks an important process transition. Above the site of **channel initiation**, hillslope processes dominate sediment transport and smooth relief. Downstream of where channels initiate, fluvial processes cause incision, increase local relief, and dominate sediment transport through channel networks [**Figure 5.12**]. This transition is often characterized as a contrast between diffusion-like hillslope processes driven by gravity and dominated by the local slope angle and channel processes dominated by advective transport involving flowing water and thus strongly dependent on discharge as well as slope.

Soil strength and vegetation act to resist channel initiation. The cohesive strength of soil helps to retard entrainment of material from the ground surface by overland flow and the detachment of material by seepage erosion in humid terrain; vegetation shields the ground surface and provides roughness that helps reduce the velocity of moving water, thereby reducing its erosive capacity. Roots also help bind soil together and retard channel incision. As discharge increases downslope from a drainage divide and overland flow deepens, the basal shear stress that flow imparts on the slope surface also increases. Because of the importance of discharge and soil moisture in channel initiation, channels typically begin in topographically convergent source areas at the head of the channel network in humid, soil-mantled landscapes.

The **channel head** is the most upslope part of the channel and is defined by evidence of sediment transport by flowing water confined between identifiable banks. The channel head may be a gradual or abrupt change in shape. Gradual channel

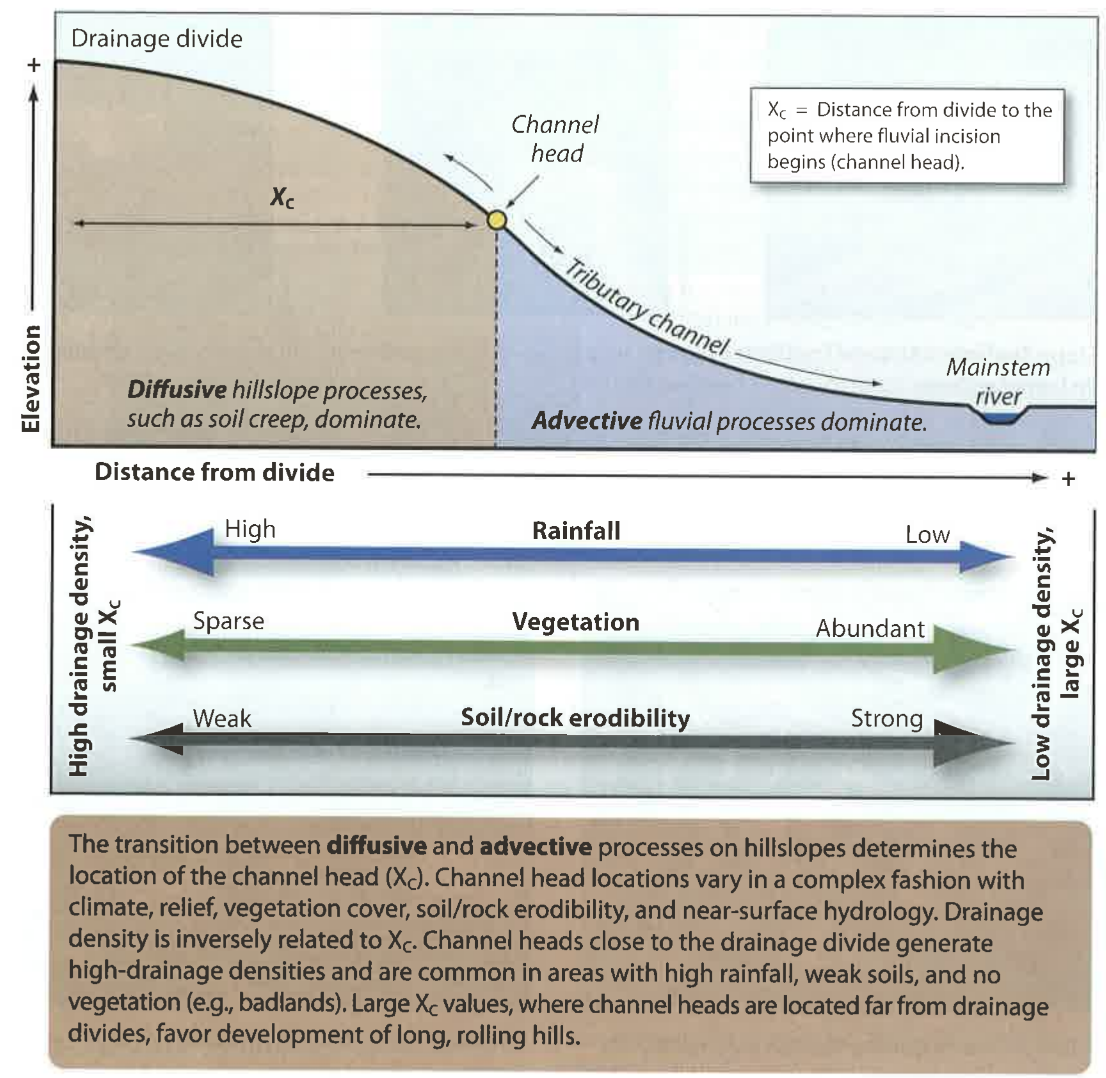


FIGURE 5.12 Channel Head Location. The channel head defines the location where advective processes moving mass out of the drainage basin become more effective than diffusive processes delivering sediment from the hillslopes to the channel. Drainage density is inversely related to the distance of the channel head from the drainage divide.

heads tend to form where sufficient overland flow accumulates to overcome the erosion resistance of the ground surface (see Photograph 4.8). On gentler slopes, abrupt channel heads often occur at the upslope end of incised gullies where subsurface water converges into unchanneled valleys. Landsliding is a primary channel initiation process in steep terrain.

Channels typically initiate where drainage areas become large enough to support a discharge capable of generating flow with sufficient shear stress to carve a channel. In many landscapes, the drainage area upslope of channel heads is inversely related to the slope angle, giving rise to smaller source areas, and thus shorter hillslopes, on steeper slopes and larger source areas, and broader hillslopes in low-gradient terrain [**Figure 5.13**].

Hollows are unchanneled valleys that typically lie at the head of the channel network. Hollows concentrate subsurface flow and gradually fill with colluvium derived from adjacent slopes. Over time, channel head extension

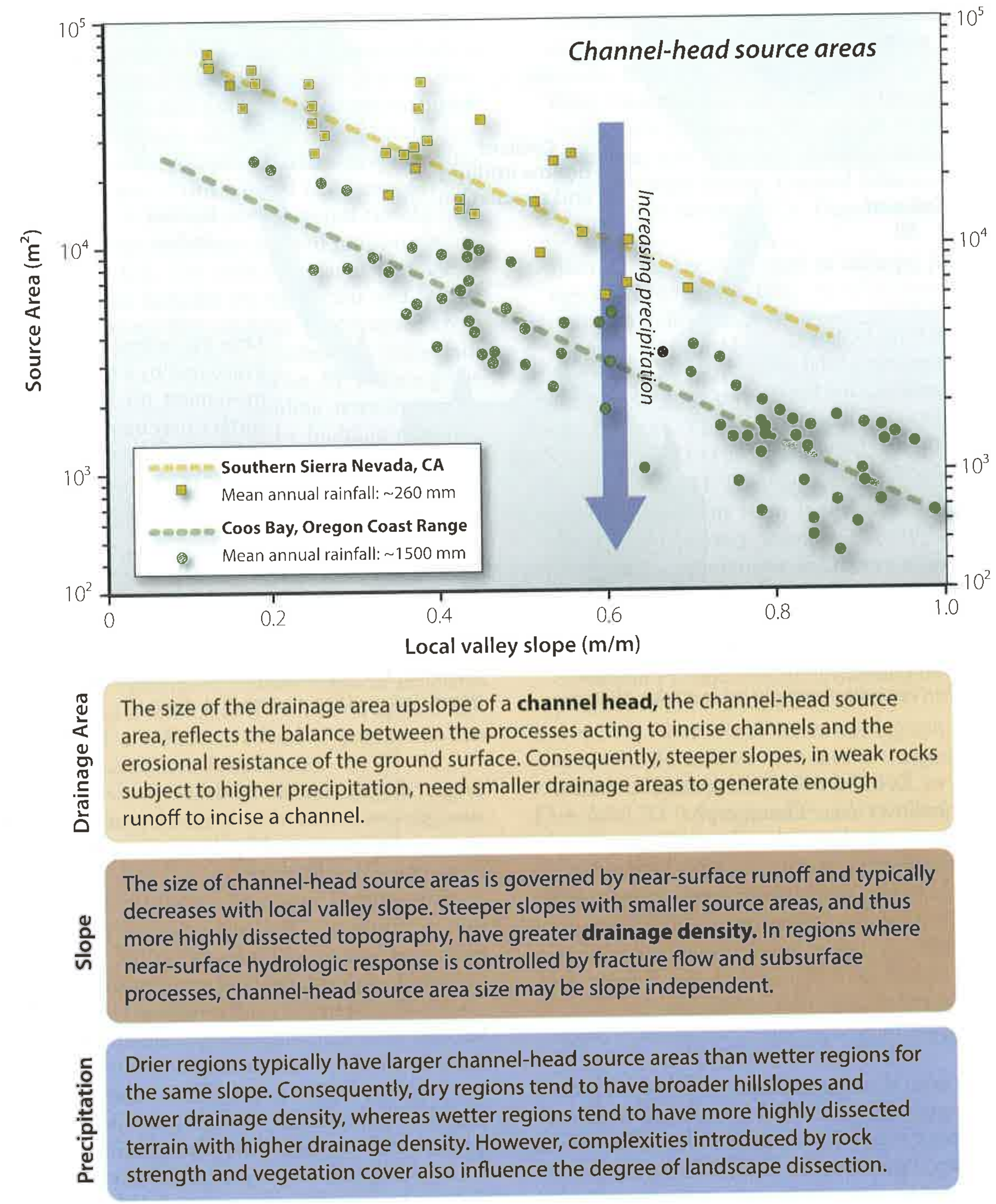


FIGURE 5.13 Drainage Area, Slope, and Climate Control on Channel Initiation. Channels begin at the downstream end of source areas, the size of which is a function of both slope and climate. In humid climates, where water is plentiful, smaller source areas are required to initiate channels than in arid climates. [Adapted from Montgomery and Dietrich (1988).]

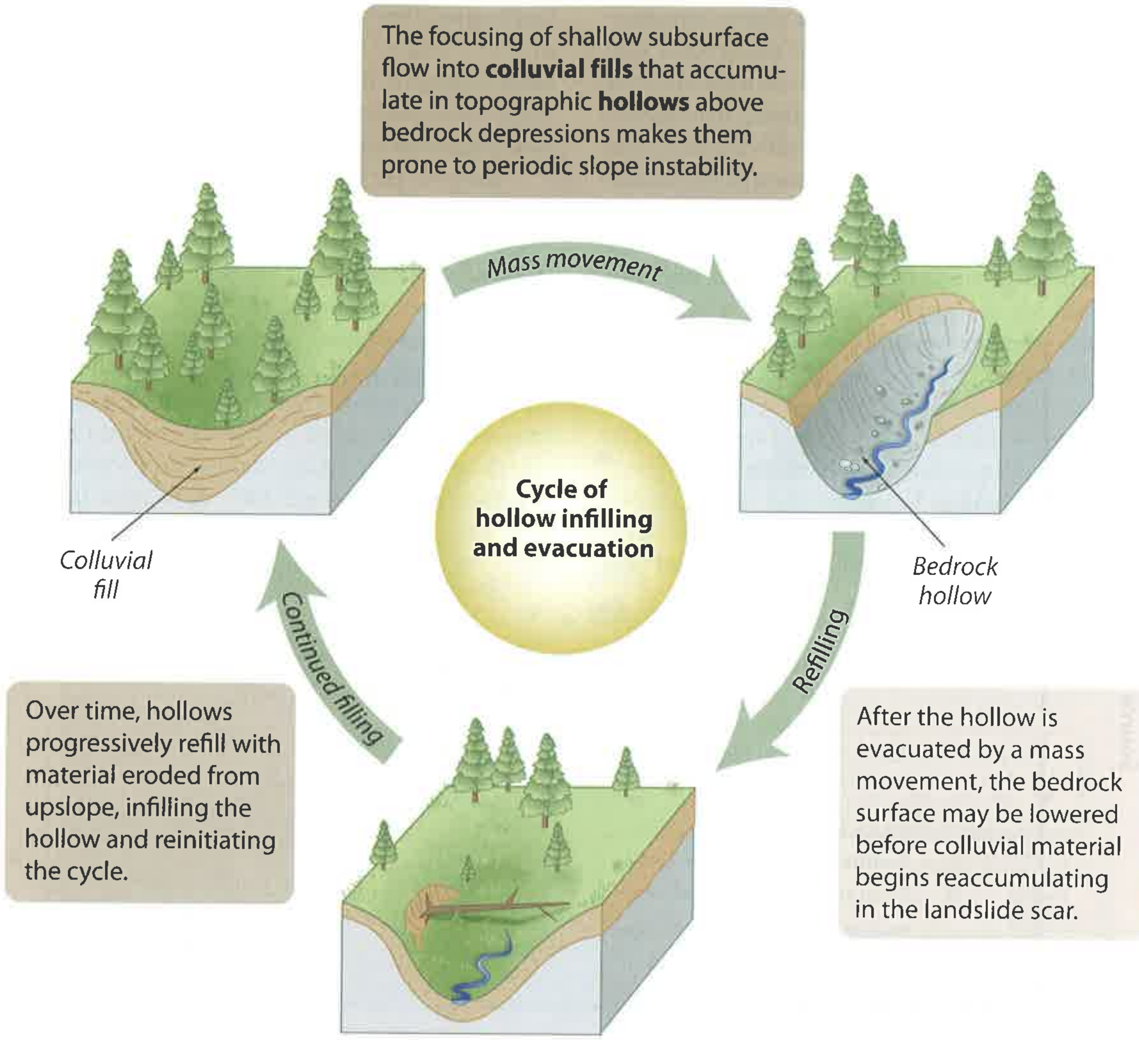


FIGURE 5.14 Bedrock Hollows. Hollows, depressions along bedrock ridgelines, cyclically fill with colluvium from bordering slopes and then episodically empty in landslides. Sliding is usually triggered by heavy rains and/or the removal of tree roots that otherwise stabilize the weak, granular colluvium on steep slopes. [Adapted from Dietrich et al. (1982).]

by landsliding or fluvial erosion is required to excavate and maintain hollows (see Photographs 4.7 and 4.8). Radiocarbon dating of charcoal collected from colluvial soils in hollows indicates gradual infilling between excavation events. In steep, landslide-prone terrain, hollows undergo a cycle of infilling, excavation, and refilling over the course of hundreds to thousands of years [**Figure 5.14**].

Applications

An understanding of slope processes and dynamics is important for engineering, hazard assessment, and land-management applications. Recognition of landslide hazard zones and ancient landslides is essential for upland development and provides important context for geotechnical engineering analyses. Building a house on an ancient slow-moving earthflow or in the path of future debris flows can invite disaster if such potential hazards are not recognized and accounted for during siting of infrastructure and in project design. Ancient landslides sometimes have residual strength lower than their original, prefailure strength and thus can be reactivated by changes in slope hydrology that accompany development.

For example, increased soil moisture from lawn irrigation in northern California communities has reactivated enormous ancient landslides. By reading terrain well enough to identify an upslope head scarp or downslope landslide toe, one can better identify ancient landslides and assess hazards to communities or new developments. Landslide hazard mapping of potentially unstable slopes—those that are prone to failure but that have not yet failed—can be based on slope stability calculations like eq. 5.8, extrapolation based on statistical characterization of observed failure locations, or the propensity for landslides to initiate in steep, convergent topography.

The increasing accessibility of LiDAR topographic data over the past decade has made slope morphology much easier to portray and analyze. Such high-resolution topographic data provide new ways to assess topography in order to identify the locations of past landslides that pose a risk to development on their surfaces or locations downslope. Similarly, advances in GIS, radar-based precipitation estimates, and landslide modeling have made it possible to produce real-time landslide hazard assessments based on digital topography at high spatial resolution over even large areas.

Human actions influence slope processes by changing the properties of hillslope materials, altering slope configuration, changing the hydrology, or adding and removing vegetation. "Follow the water" is the first rule of forensic landslide investigations in rural and urban environments where many slope failures can be traced to concentration of road runoff, leaky underground pipes, or places where homeowners and developers have directed storm runoff from roof drains and gutters onto steep slopes. Excavating the toe of a slope to make a road cut, to construct a building pad, or to mine gravel can remove basal support and trigger slope failure. Routing street or gutter runoff onto steep, slide-prone slopes can increase soil moisture levels and cause slopes to fail. The risks associated with clear-cutting steep, forested hillsides are significant and can be assessed using slope-stability analysis to delineate areas that are especially susceptible to postharvest slope failures. Slope stability is increased either by reducing the driving stress, for example, by building terraces, or by increasing the stability of the slope by building retaining walls or replanting denuded areas.

Changes to the erosion resistance of low-gradient slopes associated with agricultural practices like conventional tillage or overgrazing cause gully development, soil loss, and other types of damage to farm lands, as well as sedimentation problems for locations downslope and downstream. Effective erosion control practices begin with an understanding of the nature of the slope processes. Generally, it is far less expensive and more effective to address sediment problems in streams by reducing erosion at its source on upland hillslopes than it is to deal with an overload of sediment and its environmental consequences downstream.

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DIGGING DEEPER How Much Do Roots Contribute to Slope Stability?

Keen observers have long recognized that trees help stabilize soils on steep mountain slopes. Lyell (1853) and Marsh (1864) interpreted associations between forest cutting and mass wasting as evidence that forest clearing accelerated erosion in mountainous terrain. Since Lyell's day, the influence of root reinforcement on shallow landsliding has been well established by studies of landslide erosion under mature forest and in harvested plots, mechanistic studies of root reinforcement, and theoretical analyses based on the infinite-slope stability equation (eq. 5.8), where root strength is considered as part of the cohesion term (Sidle et al., 1985). Although roots contribute to soil strength by providing apparent cohesion and holding the soil mass together, they have a negligible effect on frictional strength. Studies from the western United States, Japan, and New Zealand all indicate that the stability of the soil mantle on steep, soil-mantled slopes depends in part on reinforcement by tree roots and that after the loss of forest cover (either by timber harvest or fire), the decay of tree roots increases the potential for slope instability, especially when soils are partly or completely saturated (Sidle et al., 1985; Bierman et al., 2005).

Root reinforcement may occur through the base of a potential landslide as roots grow into the underlying bedrock or more stable surface materials. Dense, interwoven root networks both reinforce soil and provide lateral reinforcement across potential failure scarps. Burroughs and Thomas (1977) demonstrated a rapid decline in the tensile strength of Douglas-fir roots following timber harvest in western Oregon and central Idaho and indicated the increased potential for landslides when trees were removed. Building on the Burroughs and Thomas approach, Sidle (1992) developed a quantitative model of root-strength reinforcement that combined the decay of roots after timber harvest with the regrowth of new roots [**Figure DD5.1**]. Although the decay and regrowth times vary for different tree species, a period of low root strength occurs some time between 3 and 20 years following timber harvest or fire. If a big storm occurs in

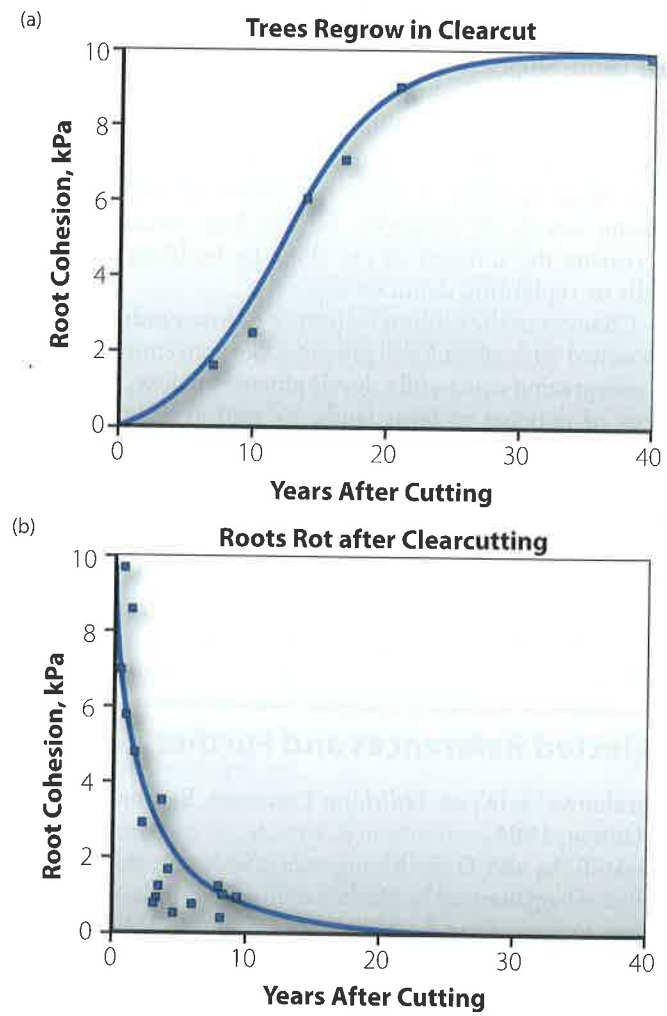

FIGURE DD5.1 Root strength changes over time as (a) trees grow in clearcuts and (b) as roots decay after trees are clear-cut. It takes about a decade after cutting for the dead roots of coastal Douglas fir trees to lose all of their strength and about 20 years for new trees to take root and develop full root strength. Planting seedlings right after harvest is a land-management strategy that reduces the chance of landsliding because new roots are growing as the old ones are decaying. [From Sidle (1992).]

this window and saturates the soil, landslides will likely follow.

Studies comparing the rate of landsliding on forested versus clear-cut slopes have reported a range of effects, from no detectable increase in landslide frequency to more than a ten-fold increase following timber harvest (Sidle et al., 1985). In a study that both analyzed a regional data set of >3200 landslides and intensively monitored a study area, Montgomery et al. (2000) found that storms with 24-hour rainfall recurrence intervals of less than 4 years (common storms) triggered landslides in the decade after timber harvesting in the Oregon Coast Range [**Figure DD5.2**]. Comparison of these postharvest rates of landsliding with the estimated background rate implied that clear-cutting of slopes increased landsliding rates by 3 to 9 times over the natural background. This increase reflected reduced root strength as the dead roots of the cut trees rotted and weakened. Without strong roots, less soil saturation was required to induce slope failure, and thus smaller storms could trigger landslides.

Schmidt et al. (2001) measured root cohesion in soil pits and scarps of landslides triggered during large storms in February and November of 1996 in the Oregon Coast Range. They found a preponderance of broken roots in the margins of recent landslide scarps, indicating that root tensile strength contributed to stabilizing the soil (until the roots snapped) in most locations. They also found that root density, root penetration depth, and the tensile strength varied among species; the tensile strength increased nonlinearly with root diameter. The median lateral cohesion provided by roots in mature natural forest ranged from 26 to 94 kPa. It was much lower in planted, industrial forest stands, ranging from 7 to 23 kPa. In clear-cuts, the lateral root reinforcement was uniformly low, under 10 kPa [**Figure DD5.3**].

Similar to Montgomery et al. (2000), Schmidt et al. (2001) found that a persistent reduction in root strength

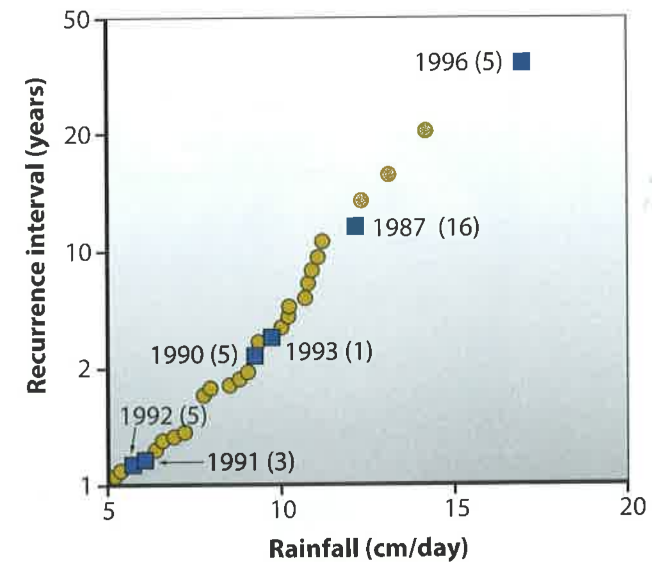

FIGURE DD5.2 Plot of recurrence intervals for 24-hour rainfall events from 1931 through 1996 (yellow circles) in a steep 0.43 km^2 study area that was clear-cut in the 1980s. Storms that occurred after clear-cutting and are known to have generated landslides are shown as blue squares. Numbers in parentheses after years indicate how many landslides occurred. Note that eight landslides occurred in this area during storms having less than 2-year recurrence intervals, all after clear-cutting. Vertical axis is logarithmic. [From Montgomery et al. (2000).]

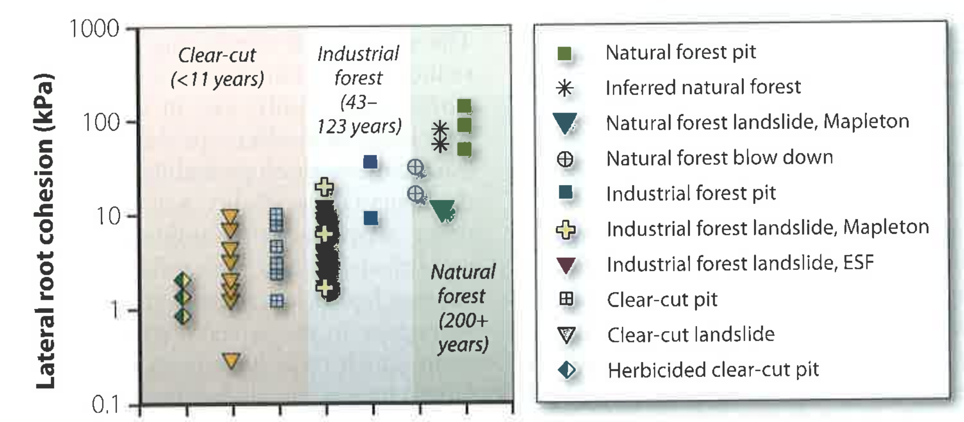

FIGURE DD5.3 In the Oregon Coast Range, not all roots provide the same amount of lateral root cohesion. Roots in clear-cuts do little to stabilize slopes. Industrial forests, those planted and managed for wood products, have roots that provide some stabilization, but the highest apparent root-cohesion values are found in mature, natural forests. [From Schmidt et al. (2001).]

DIGGING DEEPER How Much Do Roots Contribute to Slope Stability? *(continued)*

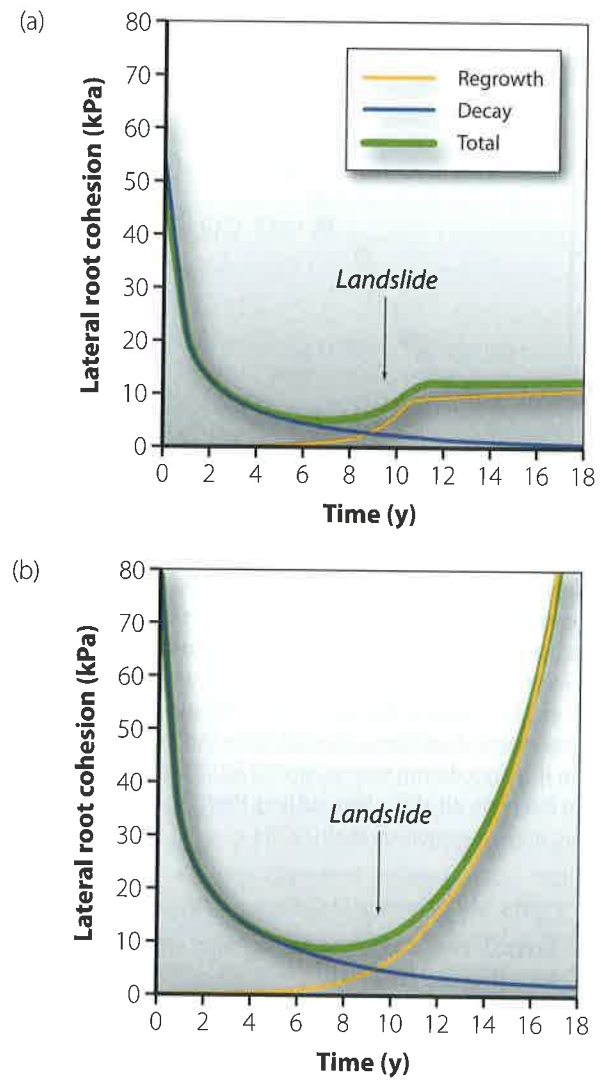

FIGURE DD5.4 Predicted total lateral root cohesion considering contributions from tree regrowth and decay of old roots for two sites that were clear-cut in 1986 and yielded landslides in 1996. Figure (a) represents a site where understory regrowth dominates vegetation. Figure (b) is a site where growth consists of abundant conifers and deciduous trees. [From Schmidt et al. (2001).]

resulting from timber harvest significantly reduced the soil moisture (m in eq. 5.8) required to trigger slope instability. They modeled root decay and regrowth for two sites that were clear-cut in 1986, and then slid in 1996. Both failures occurred close to the predicted root-strength minima, about 10 years after clear-cutting [Figure DD5.4].

Root strength varies spatially in a forest, complicating slope-stability modeling. Roering et al. (2003) documented the distribution and characteristics of trees adjacent to 32 shallow landslides in the Oregon Coast Range. Not surprisingly, bigger trees had larger root systems. The diameter of the tree crown and the root network was a function of the tree diameter (and thus tree age), and Roering et al. (2003) quantified root strength in landslide scarps by pulling on roots and measuring the tensile strength at which they broke. Summing the total root strength in each landslide perimeter, they found that root strength correlated with the size, species, condition, and spacing of trees around the landslide scarps; bigger, healthier trees spaced more closely together gave greater root strength. They also found that landslides tended to occur in areas of low root strength and thus that the potential for shallow slope instability was a function of the diversity and distribution of vegetation on potentially unstable slopes. Well-vegetated slopes were more stable.

Root strength can also vary with topographic position. Hales et al. (2009) investigated the spatial variability of root network density and strength in the southern Appalachian Mountains in North Carolina by measuring the distribution and tensile strength of roots from soil pits on topographic noses and hollows. They found that roots from trees on noses had greater tensile strength than those found in hollows, a pattern suggesting that not only does vegetation help stabilize topography but that topography affects vegetation, specifically, root strength (presumably due to differences in soil moisture). Trees on noses provided more effective root cohesion than those in hollows, a pattern that would increase further the propensity for landslides to occur in hollows.

The variability of root reinforcement with tree species, root diameter, tree diameter, topographic position, and time after timber harvest complicates quantitatively predicting the effect of root reinforcement on slope stability. The evidence is convincing that taking trees off slopes reduces root reinforcement and allows soils to fail on slopes more easily, i.e., in smaller precipitation events; however, this effect is difficult to incorporate into landscape-scale slope stability models due to the tremendous spatial variability not only in root strength but in other properties that influence slope stability, such as regolith depth and hydraulic conductivity, and the influence of bedrock fractures on soil saturation. There is no ambiguity in the science indicating that clear-cut slopes, from which trees have been removed, are more likely to fail than similar slopes under mature forest. However, managing timber-harvest-related slope instability is difficult because it is impossible to identify with certainty which potentially unstable slopes will actually fail in a particular storm. [Figure DD5.5].

FIGURE DD5.5 Debris flows off a steep, clear-cut slope, Stillman Creek, Washington. The timber company's application to the State Department of Natural Resources before harvest reported that the site had been inspected and was found to have no potentially unstable slopes. [Photograph by S. Ringman, from *Seattle Times*.]

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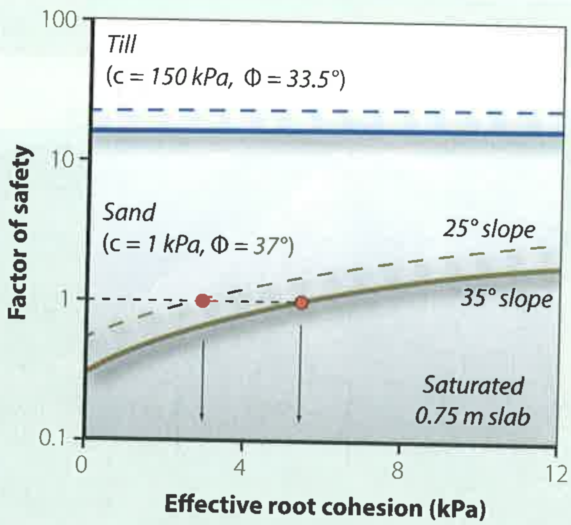
WORKED PROBLEM

Question: Using the infinite-slope model, what is the maximum stable angle for both dry and saturated sand with no cohesion and a friction angle of 37 degrees? How does this stable angle compare to that of more cohesive material such as till or clay?

Answer: For dry cohesionless materials, the maximum stable angle is the friction angle, ϕ, in this case, 37 degrees. For the failure of a fully saturated, cohesionless soil like coarse sand (FS = 1.0, C = 0, and m = 1.0), eq. 5.8 reduces to $\tan\theta = [[(\rho_s - \rho_w)/\rho_s] \tan\phi]$, which may be approximated by $\tan\theta = 1/2 \tan\phi$ (since for most soils $\rho_s \approx 2\rho_w$). This indicates that sandy slopes steeper than about half the friction angle tend to fail if saturated. Thus, when saturated, cohesionless sand with a friction angle of 37 degrees will fail when the slope is about 23.5 degrees. At higher slopes where $\theta \geq \phi$, cohesionless soils tend to slide even when dry; the soil mantle rarely stays on such steep slopes unless there is significant root reinforcement. Soils with even modest amounts of cohesion can stand at much steeper angles over length scales shorter than typical hillslope lengths. For example, excavations in clay (and other cohesive materials like glacial till) can hold vertical faces of up to several meters in height, as can riverbanks, especially if reinforced by roots that provide apparent cohesion.



Infinite-slope models are useful for evaluating slope stability and determining the effect of both cohesion and friction angle on slope stability. Here, we plot factors of safety for saturated earth materials—till with high cohesion and virtually cohesionless sand—as a function of saturation, root cohesion, and slope. The dashed lines are model results for a 25-degree slope. The solid lines are model results for a 35-degree slope. The red circles indicate the amount of root cohesion needed to prevent slope failure (factor of safety = 1). Clearly, small amounts of root cohesion can stabilize saturated sandy hillslopes. For glacial till, the effect of roots is unimportant as the material already has large amounts of cohesive strength. [From Bierman et al. (2005).]

KNOWLEDGE ASSESSMENT Chapter 5

1. List three factors that control rock strength.
2. Explain how saprolite and colluvium differ.
3. Explain how the angle of internal friction relates to soil strength.
4. What is cohesion and why is it important geomorphically?
5. List three properties that control the cohesion of soils.
6. Predict the effect of a heavy rainstorm on slope stability and explain the reasoning behind your prediction.
7. Define shear stress and explain why and how it varies with slope angle.
8. Explain two biologic effects on soil strength and erosion.
9. Contrast the form of arid and humid region slopes and explain why they differ.
10. There are straight, convex, and concave slopes. Where is each type most likely to be found?
11. How does weathering affect hillslope form and behavior?
12. Explain how weathering-limited slopes differ from transport-limited slopes and where you might typically find each.
13. Explain how talus and stair-step topography form and why they are linked.
14. Where would you be most likely to find threshold slopes?
15. List and explain the three general models of slope profile evolution.
16. Explain the importance of raindrops in shaping hillslopes.
17. Where is sheetwash likely to occur and why?
18. List the three main categories of mass movements.
19. List four specific physical processes that cause soil creep.
20. What specific observations would allow you to distinguish between active and inactive landslides?
21. Describe the physical characteristics of debris flows and the deposits they leave.
22. Predict the behavior of low-yield strength debris flows.
23. Describe three distinguishing characteristics of rotational landslides.
24. Explain the primary controls on rockslides.
25. Describe how slope stability is modeled for planar and rotational slides.
26. Explain why landslides are random in time or clustered.
27. How and where are channels initiated on slopes?
28. Explain what a hollow is and where it might be found.
29. Provide the equations for resisting and driving stresses on slopes.
30. What is the effective normal stress and how does it relate to slope stability and climate?
31. How does climate affect the shape and behavior of hillslopes?
32. Explain why the strength of rock samples measured in hand samples differs from the strength of the same rock at the scale of hillslopes.
33. Give three examples of diffusive hillslope processes and explain how the rate of such processes relates to slope steepness and slope length.
34. Describe the infinite-slope model and explain why it is useful.
35. What characteristics control drainage density and how does drainage density affect landscape response to storms?