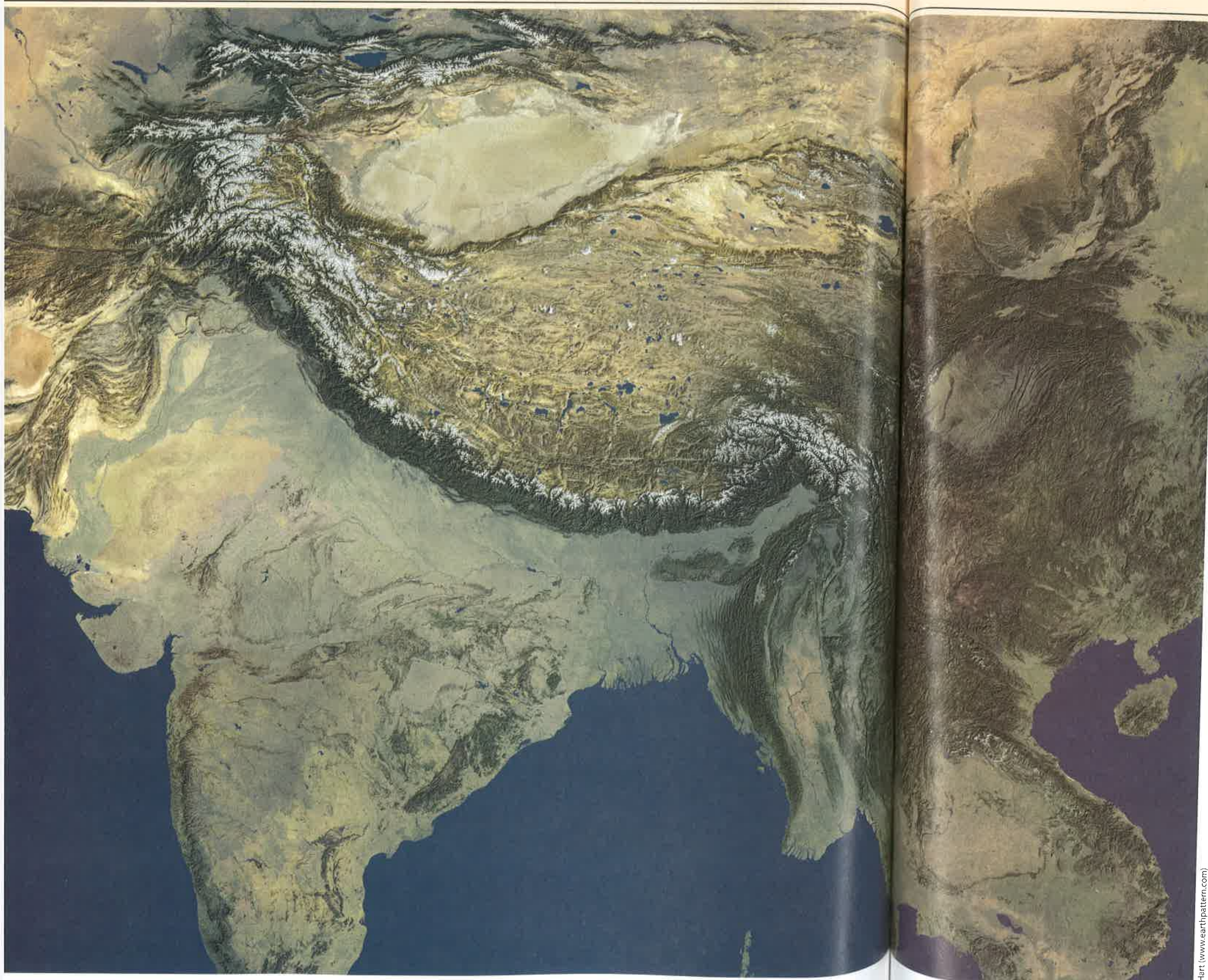
PART IV



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The Bigger Picture

(Chapters 12 through 14)

Tectonics and climate govern landscape process, evolution, and form by changing important boundary conditions including base level, uplift rate, temperature, and precipitation. Together, these characteristics set the pace of weathering, erosion, and sediment transport. Chapter 12 considers how tectonic processes shape landscapes and how tectonic influences can overwhelm those of erosional processes and earth materials to produce distinctive features. In Chapter 13, we examine the influence of climate on the rate and distribution of surface processes and consider the distribution of landforms indicative of specific climatic conditions. Chapter 14 brings together ideas presented earlier in the book to understand how landscapes change and evolve over time. After reading these three chapters, you should understand how tectonics affects landscapes, how landscapes both respond to and influence climate, and how landscapes evolve over time in response to changes in the processes shaping Earth's dynamic surface. Our goal for these chapters is to integrate the more specific material presented earlier in the book into a broad treatment through the big-picture lenses of tectonics, climate, and landscape evolution.

Structural deformation resulting from India's collision with Asia is apparent in the folded terrain to either side of the Himalaya, which defines the southern edge of the Tibetan Plateau. Deposition of material eroded from the elevated areas formed the extensive alluvial plains along the Indus and Ganges rivers. Image is a composite of Moderate Resolution Imaging Spectroradiometer (MODIS) natural color imagery superimposed on a digital elevation model of India, Tibet, and southeast Asia from the Shuttle Radar Topography Mission data. The daily MODIS data from August 2002 were stacked together and the least cloudy sample was selected to produce a nearly cloud-free image.

Tectonic Geomorphology

12

USGS

Alaska's Denali Fault Zone last ruptured on November 3, 2002, in a 7.9 magnitude earthquake. The fault scarp just west of the Delta River had an offset of about 5 m. Ground cracks, like the one pictured, were up to 3 m deep.

Introduction

The tectonic processes that move rocks shape the global distribution of continents, mountains, and ocean basins that, in turn, control the distribution and intensity of surface processes and the resulting landforms. Tectonic processes elevate rocks above sea level, where weathering prepares rock materials for sculpting into landscapes through erosion by wind, rain, rivers, and glaciers.

Earth's high-elevation surface features coincide with the boundaries of tectonic plates because rock uplift and deformation are focused at such boundaries. Tectonically active regions give rise to topography with spectacular vertical relief, whereas tectonically quiescent areas host broad lowlands and older, lower-relief uplands. Geomorphically intriguing landscapes include those that are tectonically quiescent at present but topographically significant (such as the southern Appalachian Mountains). Equally impressive are tectonically active areas raised to high elevation but lacking significant local relief (such as the high plateaus of Tibet and the Altiplano in South America). At the edge of these flat-lying highlands, spectacular landscapes develop, the result of erosional processes acting on the steep margins of these uplifted plateaus.

The imprint of tectonics on geomorphology is apparent not only in the size, extent, and location of mountain ranges, but in the localized steepness of river profiles, the character of mountain slopes, and in the form of river networks that develop along regional joint patterns. Tectonics sets the stage for, and sometimes directs, the

IN THIS CHAPTER

work of erosion. Tectonic processes also form particular small-scale landforms that provide clues as to the style and pace of tectonic deformation.

Tectonics influences geomorphology both actively and passively through the effects of rock uplift and subsidence on relief, spatial patterns of slope, and the juxtaposition of bedrock of different mechanical properties. **Active tectonic controls** involve landscape response to ongoing deformation of the land surface, such as the response of a river's slope to fault offset and regional uplift. **Structural controls** (sometimes referred to as passive tectonic controls) are those that influence landforms and landscape dynamics indirectly through spatial patterns in the erosional resistance of rocks. In tectonically active regions, geologic structure and lithology develop only minor geomorphic expression during active mountain building, as uniformly steep slopes and rapid erosion mask the underlying geologic structure and make structure difficult to infer from hillslope morphology. Once active tectonic forcing ends, lithology and structure emerge as the major controls on landforms, sometimes allowing one to map or infer geologic structure from topography.

This chapter examines landscape response to tectonics, beginning with how different tectonic settings give rise to broad landscape assemblages with characteristic landforms that reflect specific tectonic processes. We discuss landforms indicative of deformation and explore geologic controls on landforms that can persist long after active tectonic forcing ceases. We also consider how high rates of erosion in tectonically active areas influence the growth and development of geologic structures by spatially focusing and sustaining both erosion and **isostatic rebound** (see Chapter 1) over geologic time.

Tectonic Processes

The motion of Earth's tectonic plates drives the lateral and vertical movement that results in extension, compression, uplift, and subsidence—all of which are important in creating characteristic landcapes at the largest scale. Maintaining the mean elevation of a landscape requires that the tectonic processes driving rock uplift counteract the mass removed by erosion above sea level. Landscapes continually erode and, given time, eventually come to reflect the balance between the forces driving rock uplift and the erosional potential arising from the steepness of slopes and rivers draining them—the form of the land itself. Although uplift of the land surface is generally the result of the movement of tectonic plates, buoyancy due to thermal and density contrasts also elevate areas above neighboring regions. In addition, the isostatic response to erosion raises fresh rock to replace eroded material even as the average elevation of the land surface lowers (see Chapter 1).

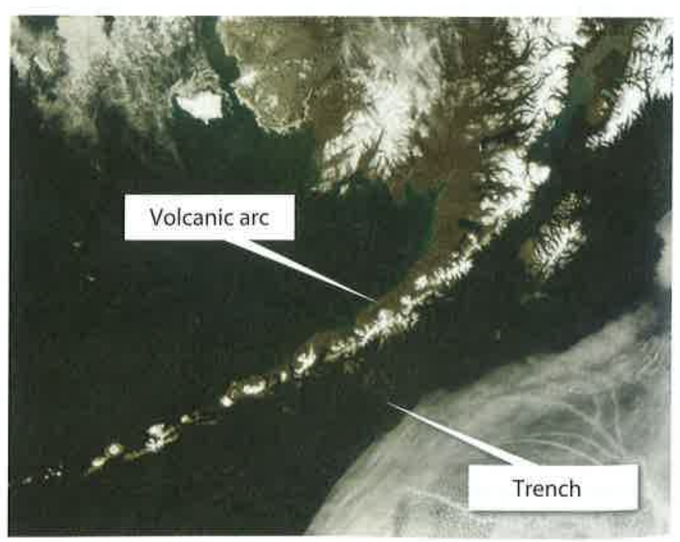
Each of the primary mechanisms driving tectonic geomorphology—plate tectonics, **dynamically supported topography** (i.e., topography supported by heat and density contrasts in the crust and mantle), and isostatic response—results in different amounts and styles of rock uplift.

Dynamically supported topography is a key component of mean landscape elevation and is not independent of plate tectonics. Both plate tectonics and dynamically supported topography are consequences of a convecting thermally inhomogeneous mantle.

The common element among these tectonic processes is that rock uplift, and spatial gradients in rock uplift, set the stage for erosion to sculpt the land. In general, the relatively small areas of spectacular mountainous topography that fascinate tectonic geomorphologists occur in extensional rift zones, zones of collisional tectonics along continental margins, and areas that were formerly in such tectonic settings (see Chapter 1, Figure 1.4).

The interaction of tectonic plates elevates rock masses above sea level, where they are subject to subaerial erosion. Spreading centers along mid-ocean ridges and continental rifts make new crust and create ocean-bottom topography far from where plates collide, crumple, and deform to make mountains. Where oceanic and continental plates collide, cold and dense oceanic lithosphere (the outer solid part of Earth, including the crust and uppermost mantle) sinks into the mantle beneath less-dense continental material. This creates a subduction zone along which oceanic sediments can be scraped off the downgoing slab to form a thickened wedge of sedimentary rock that forms a coastal mountain range in front of a plate-margin-parallel range of active volcanoes. Examples of such a wedge include the Olympic Mountains and Oregon Coast Range that rise seaward of the volcanic arc of the Cascade Mountains in western North America.

Volcanic arcs like Alaska's Aleutian Islands [**Photograph 12.1**], or the northern and southern Andes in South America, lie above the point where subducted material descends deep enough (about 100 km) to drive volatiles (like water) from the subducting and metamorphosing rock and sediment and up into the overlying mantle. These volatiles depress the melting point of rock enough that it partially melts, sending lower-density magma rising toward the surface. Thus, near oceanic subduction zones, chains of volcanic islands parallel to the subduction zone characterize the topography above the subducting plate. The horizontal distance from the subduction trench to the arc of volcanoes, and thus the width of the depositional forearc seaward of the arc, varies between many tens and a few hundred kilometers, depending on the angle at which the slab subducts. Steep subduction reduces the distance because the slab more quickly reaches a depth (and thus temperature) where partial melting and magma generation occur.



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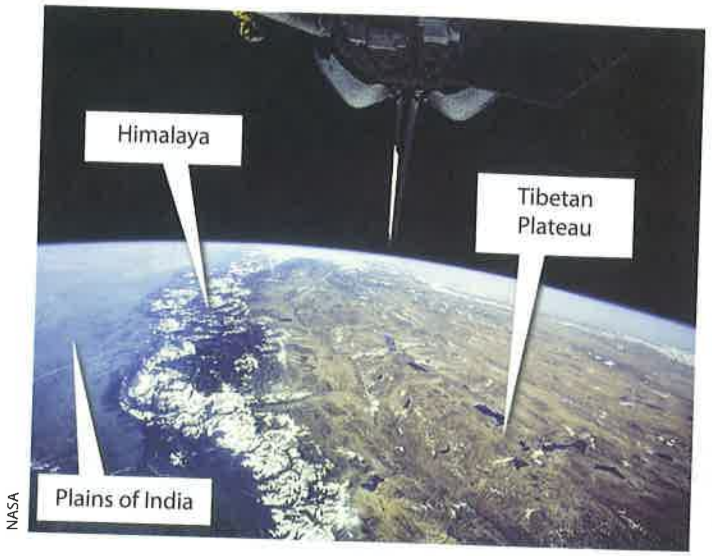
PHOTOGRAPH 12.1 Aleutian Volcanic Arc. Satellite image of the Aleutian Islands, showing the volcanic arc that parallels the adjacent trench. Clouds obscure lower right of image.



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PHOTOGRAPH 12.2 Himalayan Front. View from the space shuttle looking west at the Himalayan front, with the foreland basin and plains of India to the left, and the Tibetan Plateau to the right.

C. Hormann

PHOTOGRAPH 12.3 East African Rift System. Synthetic image based on digital topography of East African Rift Valley, looking north (top of image). Several large lakes occupy the rift. On the uplifted western (left-hand side) rift flank, the Ruwenzori Mountains (Uganda) are about 5000 m high.

The nature of plate boundaries governs tectonic influences on landscapes. The mean elevation of Earth's surface generally reflects the thickness of continental crust; thicker continental crust leads to higher average elevation because continental crust is less dense than oceanic crust. When two lithospheric plates carrying continental crust collide, neither is dense enough to sink into the mantle. Instead, they crumple together in a collision that creates high mountains, like the Himalaya, through crustal thickening that builds up a low-density root; isostatic compensation of this root elevates the range well above sea level [**Photograph 12.2**].

Where two plates diverge in a rift zone, they create a spreading center associated with crustal thinning and mantle upwelling. Thermal buoyancy of the underlying, rising mantle and flexural uplift of the unloaded rift flank elevate rift-valley flanks above surrounding areas [**Photograph 12.3**]. At rifts, no crustal root is created, and this uplift is thermal and dynamically driven. Without a crustal root to drive a long-term isostatice response, uplift at rifts is relatively short-lived and ends when the lithosphere cools.

Where two plates slide past each other, the result is a series of long, linear landforms [**Photograph 12.4**] with differential uplift only if there are bends in the fault that cause local extension or compression.

Within each of these settings, tectonics influences geomorphology across a wide range of spatial scales, from the regional physiography to local fault interactions that raise some areas and cause others to subside. At each of these scales, understanding the influence of tectonics on geomorphology centers on landform analysis and knowledge of the underlying tectonic and geomorphic processes.

Uplift and Isostasy

Several distinct types of uplift are defined based on their reference frame: surface uplift, uplift of rock material, and

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PHOTOGRAPH 12.4 Fault Trace. The San Andreas Fault viewed from the air looking along the Carrizo Plain in southern California. Two streams are offset (right laterally) in this 150 m length of the fault.

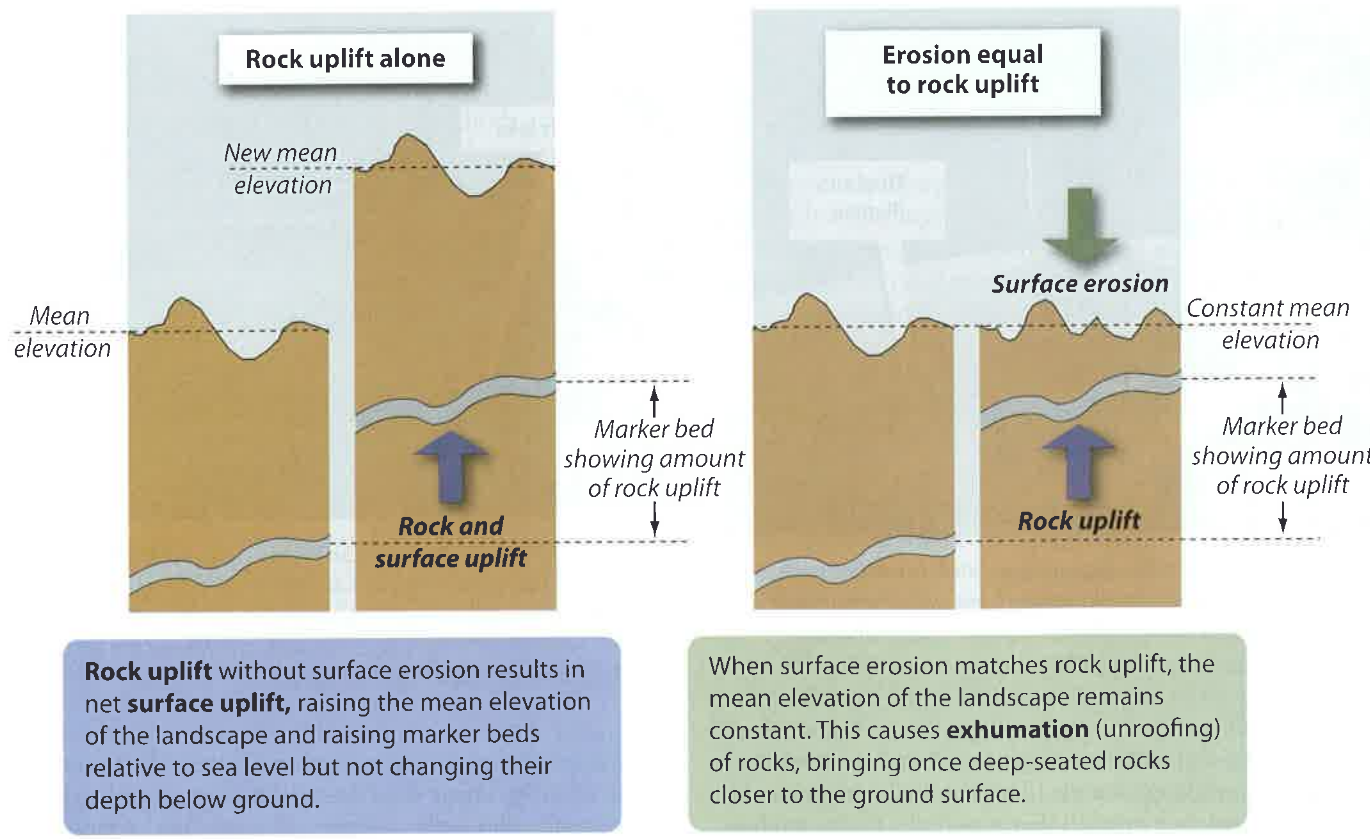


FIGURE 12.1 Surface Uplift Versus Rock Uplift and Exhumation. Relationship between surface uplift and the uplift and exhumation of rocks, illustrated through the end-member cases of no erosion and of erosion equal to rock uplift.

exhumation [Figure 12.1]. **Rock uplift** refers to changes in a rock's vertical position relative to a fixed datum such as sea level. **Surface uplift** refers to the change in the elevation of the land surface (again referenced to a fixed datum) rather than the uplift of a specific rock layer. Although the uplift of rocks and surfaces are tied to the same datum (usually sea level), rock uplift equals surface uplift only if no erosion occurs. If there is erosion, the surface uplift must be less than the rock uplift. **Exhumation** of rock describes the uplift of rocks relative to the ground surface through erosion that brings subsurface rock closer to the ground surface by stripping off the overlying material (lowering the ground surface). Unless countered by erosion, tectonically driven rock uplift will increase the land surface elevation. Surface uplift (U_S), uplift of rock (U_R), and erosion (or exhumation, E) are all interrelated:

$$U_S = U_R - E \qquad \text{eq. 12.1}$$

where rock uplift (U_R) is the sum of tectonically driven rock uplift (U_T) and isostatically driven rock uplift (U_I)

$$U_R = U_T + U_I \qquad \text{eq. 12.2}$$

Erosion and rock uplift are not independent because when erosion occurs it triggers an isostatic response that raises rock toward the ground surface. Consequently, the resulting net surface lowering from erosion is much less than the depth of rock removed—due to the uplift of rock by **isostatic compensation** (see Chapter 1). The magnitude of such compensation is given by $U_I = (\rho_c/\rho_m)E$, where ρ_c and ρ_m are the density of the crust and mantle, respectively.

Hence, the resulting net change in surface elevation due to erosion will be equal to:

$$U_S = E\,[(\rho_c/\rho_m) - 1] \qquad \text{eq. 12.3}$$

For each meter of rock stripped off the landscape, isostatic response lifts the underlying rock back up by about 82 cm due to the density contrast between the less dense eroded crust ($\rho_c \approx 2.7$ g/cm^3) and the more-dense underlying mantle ($\rho_m \approx 3.3$ g/cm^3). When tectonic uplift ceases (i.e., $U_T = 0$), the isostatic response to erosion (U_I) becomes the dominant form of rock uplift. Consequently, in tectonically quiescent regions, erosion itself plays a prominent role in rock uplift as the isostatic response to erosion continues to exhume rock, bringing once-deeper rock closer to the ground surface in response to removal of overlying mass.

Tectonic convergence over the lifetime of a mountain range builds up a thick crustal root. This root isostatically sustains elevated topography long after tectonic activity ceases because erosion must remove an amount of rock equivalent to many times the total relief before the root is reduced enough that it no longer supports elevated topography. Because postorogenic erosion rates are relatively slow, it can take surface processes a very long time to erase a mountain range supported by a deep crustal root (see Worked Problem).

Over geologically short timescales, isostatic compensation is influenced by the wavelength (size) of the load and by the **flexural rigidity** (stiffness) of Earth's lithosphere, which distributes and partially supports topographic loads [Figure 12.2].

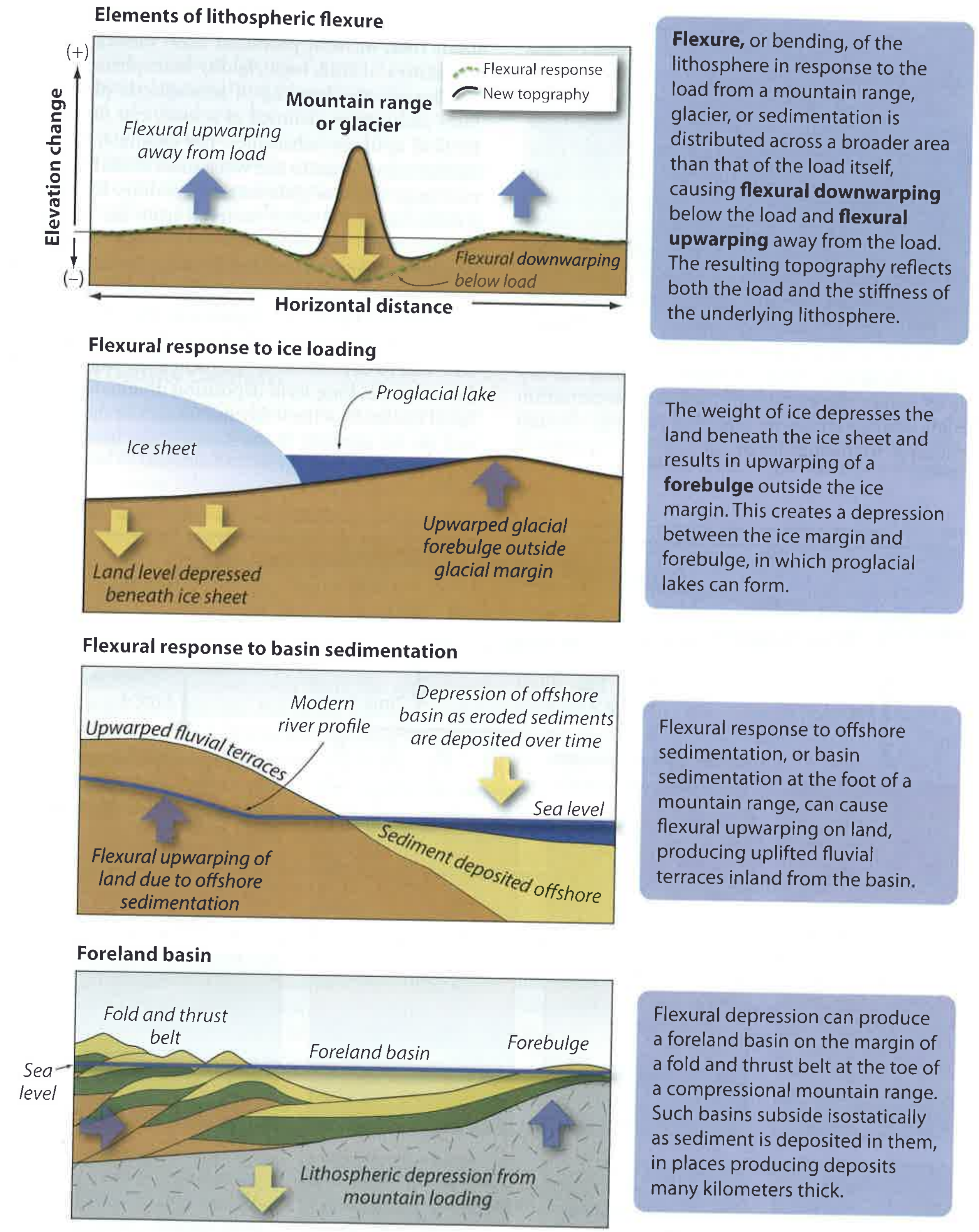


FIGURE 12.2 Plate Flexure. Plates have flexural strength that distributes the loading or unloading of material, resulting in far-field depressions, bulges, and tilting.

The flexural rigidity of the lithosphere means that isostatic rebound or subsidence are distributed across a wide area, in contrast to how local erosion can vary greatly across the landscape. The flexural rigidity of the crust varies regionally and controls the wavelength and amplitude of isostatic response to either the addition of a topographic load (such as from a rising mountain range produced by lateral convergence of rock material) or the removal of a load (whether from erosion of rock or melting of glacial ice).

The wavelength of isostatic response is on the order of tens to hundreds of kilometers, exceeding the width of individual valleys. Thus, it is possible that mountain peaks can rise in response to the excavation of valleys if mass is preferentially removed from valley bottoms [**Figure 12.3**], triggering isostatic uplift across a broader region. As much as a quarter of a mountain's height can result from the incision of nearby valleys. It is no coincidence that many high peaks in the Himalaya rise above adjacent, deeply incised river valleys.

The magnitude of isostatic compensation for erosion or glacial melting and the spatial scale over which it occurs are related to the flexural rigidity. Hot, thin lithosphere is weak and has low flexural rigidity, resulting in greater isostatic compensation. The greater strength of cold, thick lithosphere produces higher flexural rigidity that acts to reduce the amount of isostatic compensation and spreads it across a broader area. The ongoing rebound of Scandinavia following deglaciation is an example of how the removal of a load (the ice sheet) triggered rock uplift that, in turn, produced slow surface uplift over a broad area of cold, high-rigidity lithosphere.

Flexure, the bending of lithospheric plates from the local addition or removal of a load, can indirectly cause far-field uplift or subsidence. For example, flexure causes upland source areas to rise when mass eroded from a mountain range fills a neighboring sedimentary basin. Just such a thing happened when material from the Sierra Nevada was deposited in California's Great Valley and to a lesser degree when material eroded from the Appalachian Mountains was deposited on the Atlantic shelf.

Foreland basins are depressions that develop adjacent and parallel to a mountain belt from flexural isostatic subsidence due to deposition of sediment shed from the uplands. For example, long-term deposition dominates the great alluvial forelands of the Indus and Ganges-Brahmaputra rivers, and on the east side of the Andes in the headwaters of the Amazon. Depositional foreland basins drive their own subsidence as the weight of deposited sediment (with a typical bulk density of about 1.6 to 1.7 g cm^{-3}) compresses previously deposited material and isostatic response causes roughly 50 cm of subsidence for each meter of material deposited. Piles of sediment >10-km-thick have accumulated in broad forelands adjacent to steep mountain fronts of the Himalaya and the Andes due to the isostatic depression of the lithosphere during millions of years of sediment deposition. Not all of the sediment shed off the Himalaya and the Andes makes it to the sea. Much is trapped in extensive, low-relief depositional zones (subsiding basins) along the Ganges and the Amazon rivers.

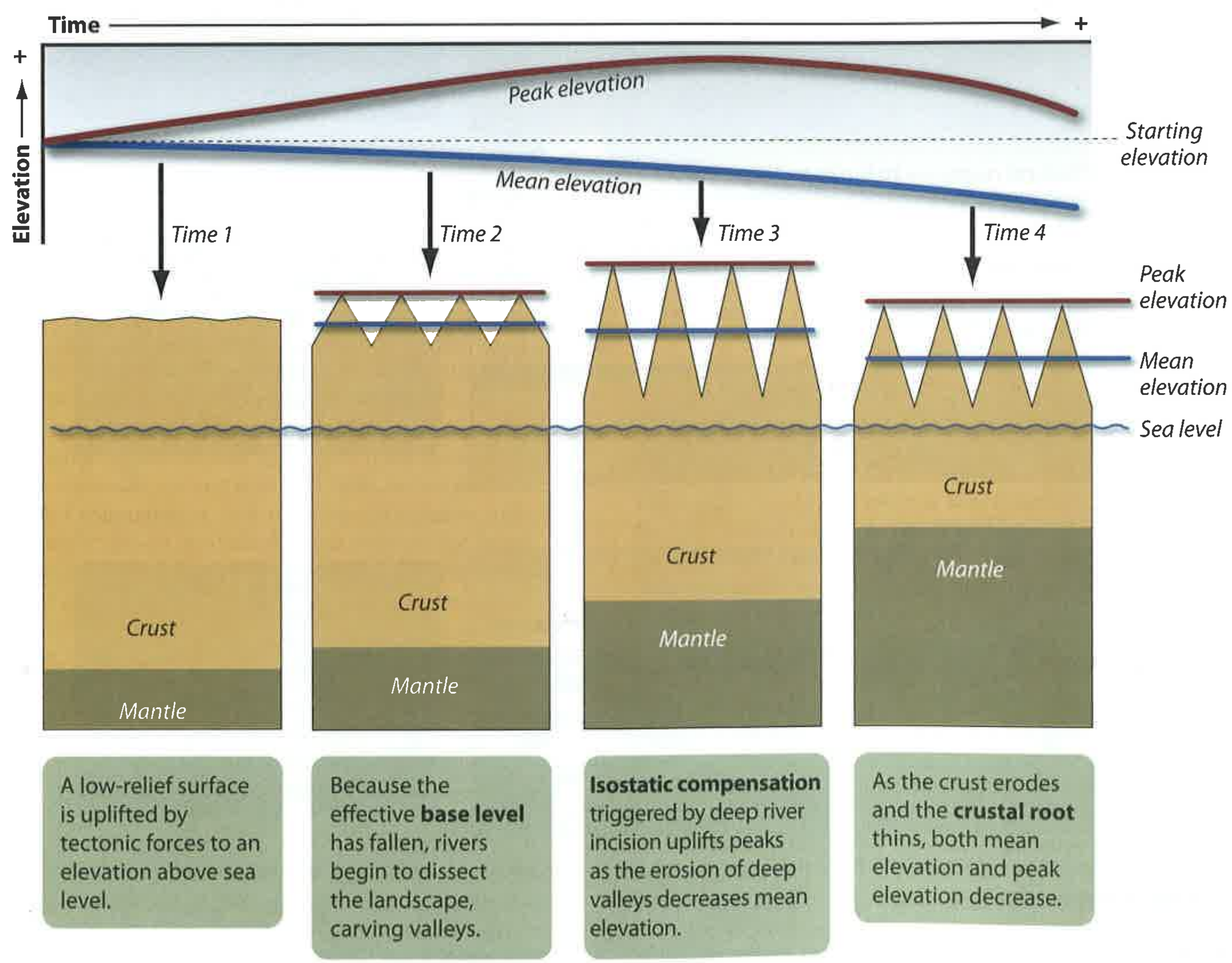


FIGURE 12.3 Isostatic Uplift of Mountain Peaks. Incision of deep valleys removes mass from mountains, resulting in isostatic compensation at broader length scales. Hence, even as mean elevation of the range lowers, peak elevation can be driven higher by rock uplift. [Adapted from Burbank and Anderson (2001).]

Thermal and Density Contrasts

Tectonic uplift and the resulting landscape response are not limited to collisional settings along plate margins. Thermal contrasts related to tectonic forcing can elevate topography wherever heating from subsurface sources produces lower-density, and thus more buoyant, lithosphere. Lithospheric thinning that accompanies upwelling of hot rock in extensional settings, such as rift zones, can lead to **thermal uplift** of landscapes as buoyancy contrasts raise warm, less-dense lithosphere above adjacent cooler, denser lithosphere. The resulting physiographic upwarping decays over time as rocks cool with distance from the thermal anomaly. Thus, the elevation of deep-sea ridges decreases away from the spreading center at the ridge crest because the newly extruded rock cools and becomes denser as it moves laterally away from the rift over time (see Chapter 8). Large areas of thermal buoyancy, associated with rising mantle plumes, can result in dynamically supported topography, high-standing regions that owe their elevation to thermal forcing. In such settings, the lack of a thick crustal root means that topography rapidly becomes subdued as the lithosphere begins to cool. Dynamic support from mantle upwelling, which can result in higher elevations than would otherwise be supported by the crustal root (as is argued by some for southern Africa), further illustrates the connections between surface topography and deep-Earth processes. Low-standing topography, such as deep ocean trenches near subduction zones, can also be dynamically maintained; in this case, Earth's surface is depressed by sinking lithosphere.

Not only are thermal contrasts central to igneous processes that lead to volcanoes and volcanic landforms, but thermally driven uplift associated with mantle plumes also influences continental drainage patterns and volcanism in continental interiors (e.g., Yellowstone, Montana) or on plate margins (e.g., Iceland). The hot spot responsible for forming the Hawaiian Islands left a trail of progressively older (and now lower) islands to the west from the big island of Hawaii, where thermally driven uplift is buoying up the oceanic lithosphere. Along the island chain, older islands have subsided more because the older lithosphere below them has progressively cooled; their age is also reflected in their landscapes. The degree of dissection, rock weathering, and soil development all increase as the islands age (see Chapter 11).

PHOTOGRAPH 12.5 Salt Glaciers. Dark-colored salt flowing from breached anticlines in the Zagros Mountains in southern Iran, a zone of collision and uplift. The salt is less dense than the surrounding rock and moves upward as diapirs. When these diapirs reach the surface, the salt flows ductilely under gravity-induced stress.

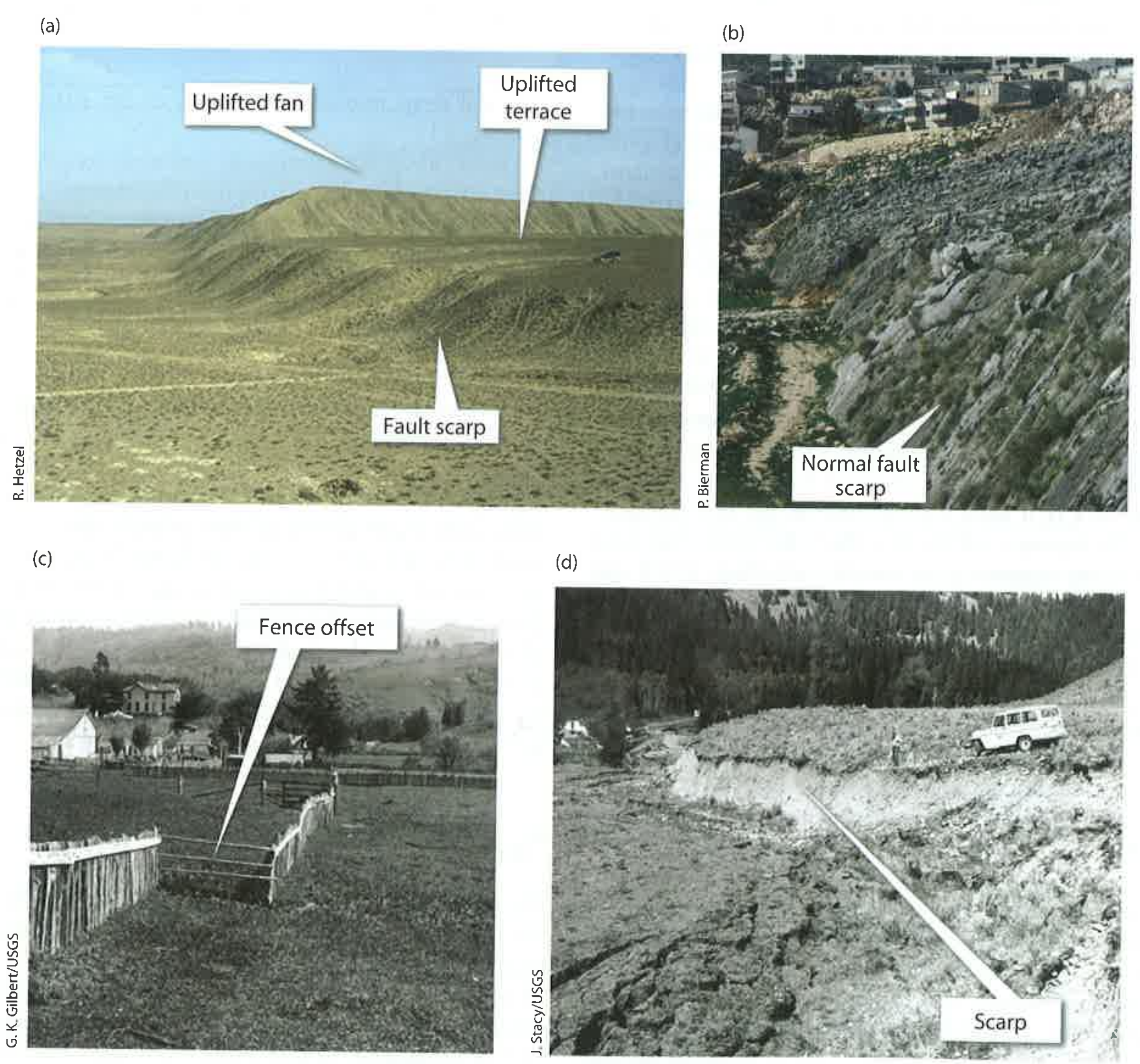
In addition to tectonically induced density contrasts, there are stratigraphic situations in which low-density materials, like salt, underlie higher density rock. This creates gravitational instabilities through which buoyancy drives uplift and structural deformation. When a body of salt, such as a deposit of marine evaporites, is buried beneath denser overburden, the salt may rise up through the overlying material to form a **diapir,** a bulbous intrusion that can take odd shapes and form a salt dome where it displaces and deforms overlying strata. Rising salt bodies that reach the surface can flow out over the ground as salt glaciers [**Photograph 12.5**], which move much like conventional glaciers, only much slower, at rates on the order of meters per year. Although they are rare on Earth, being concentrated in extremely arid regions such as the Zagros Mountains in Iran, salt glaciers also appear to occur on Mars (although there they might be made of sulfates and other salts rather than sodium chloride). Layers of salts interstratified with other types of rock provide zones of weakness along which substantial deformation may become focused.

Tectonic Settings

Tectonic setting is the primary control on the global pattern of regional physiography and landscape character. The regional tectonic settings of active plate margins, passive continental margins, and continental interiors strongly influence landforms through styles of tectonic deformation and uplift, differences in dominant lithologies, and changes in the degree of fracturing (which affects erosion resistance). Distinct topography characterizes plate margins with different types of lithosphere (oceanic or continental). Because tectonic plates generally consist of both oceanic and continental lithosphere, it is useful to consider separately the dominant controls on the topography of extensional zones (including passive continental margins), compressional margins, transform margins, and continental interiors.

Tectonic setting and structural geology influence landforms through the direct action of faulting and isostasy and through the indirect influences of spatial variability in erosion resistance generated by folding, faulting, and offset of rocks of differing lithology. **Faults** are discontinuities in Earth's lithosphere that record deformation and movement and that are now boundaries between rocks that did not originally form next to one another. Springs preferentially occur along fault traces due to interception and blockage of subsurface water flow across the fault and/or enhanced discharge through a permeable zone of crushed rock along the fault. Faults can influence topography either through surface uplift and offset or through fault-influenced patterns of differential erosion, which can create erosional relief where highly erodible rocks are juxtaposed against erosion-resistant rocks. **Fault scarps** are steep linear slopes that reflect the direct topographic expression of fault offset [**Photograph 12.6**]. Offset across fault scarps in seismic events is highly variable but can be as great as 10 m for a single earthquake. The height of a fault scarp often reflects multiple discrete episodes of offset.

Below, we consider the types of landforms that typically develop in association with fault systems along **normal faults** with extensional offset, **reverse faults** with compressional offset, and **transverse faults** with lateral (strike-slip) offset. The topographic influences of tectonics and structural geology depend on the rate and type of crustal deformation, the three-dimensional geometry of the resulting geological structures, and differential erosion of rock involved in the deformation.



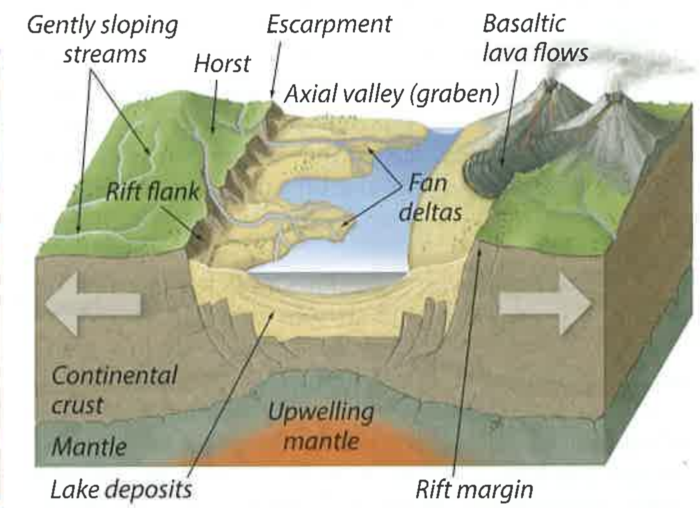
PHOTOGRAPH 12.6 Fault Landforms. Some landforms provide surface evidence of faulting. (a) In the distance, a thrust fault scarp in Tibet offsets a fan and a river terrace on which the vehicle is parked. (b) Normal fault scarp in limestone in the northern Galilee of Israel strikes directly through a village. (c) Fence offset by the 1906 San Francisco earthquake along the strike-slip San Andreas Fault in Marin County, California. (d) Normal fault in alluvium generated by the 1959 earthquake at Hebgen Lake, Montana.

Extensional Margins and Landforms

Zones of divergence (extension) are places where continents split apart, producing geomorphologically distinctive landscapes that include escarpments, rift valleys, volcanoes, and alluvial and debris fans. At extensional plate boundaries, tectonic forces pull plates apart and form rift zones that can, over time, mature into a spreading center and mid-ocean ridge. Such tension (extension) produces fracturing and normal faulting that facilitate the eruption of lava and the erosion of rock masses. Although mid-ocean ridges constitute most of the world's extensional plate boundaries, divergence within continents forms major active **rift valley systems** like those of the East African Rift system and the Rio Grande Rift in New Mexico and Texas [**Figure 12.4**]. Rifting is the first step in the formation of passive margins like the east coast of South America and the west coast of southern Africa.

Zones of continental extension range in scale from individual **pull-apart basins** within transform fault systems like the Dead Sea Rift of the Jordan Valley to regional extension along high-angle normal faults that extend through the shallow crust in the Basin and Range Province of western North America. The East African Rift system, long thought to be the cradle of human evolution, is still in the early stages of breaking up the heart of Africa. The separation of the Arabian Peninsula from Africa across the Red Sea, to which the East African Rift is connected, represents a more advanced stage of rifting and the birth

Red Sea Rift Zone



Where tectonic spreading centers occur on land, they form extensional **rift zones.** Where they occur in ocean basins, they form **mid-ocean ridges.** Both are characterized by an axial valley and upwarped rift margins. As a continental rift zone develops and widens, it can progress from upland topography like the East African Rift Zone to form an incipient ocean like the Red Sea. With time, the Red Sea may continue to widen and eventually become an ocean like the Atlantic.

Extensional **rift zones** often have a down-dropped **axial valley** filled with accumulations of lake sediments and alluvium shed from steep rift flanks that drain into the rift. Gentler slopes drain away from the rift flanks outboard of the upwarped rift margin. Hot mantle material upwelling beneath the thinned crust of the axial valley drives **thermal uplift** and **dynamic support** of the overlying topography. Lakes can develop in down-dropped grabens between high-standing horsts.

Extensional zones contain diagnostic **normal faults** that delineate range fronts, can offset alluvial fans, and form triangular bedrock facets that represent the exhumed fault surface. Hourglass-shaped valleys extend into the range front between the facets.

FIGURE 12.4 Rift Zones—Extensional Settings. Rift zones characterize extensional tectonic settings both on land and in the ocean, and they produce landforms characteristic of normal fault zones and extension.

of a new ocean basin. Spreading at the Mid-Atlantic Ridge created the Atlantic Ocean basin and progressively separated Africa and South America while creating the largest and longest mountain chain in the world. Ancient rifting created continental margins and the world's great escarpments (see Chapter 14).

In most rift zones, the central axial valley is part of a system of interconnected local depressions atop a great ridge. Typically, several strands of rift valleys together define an integrated rift system. The topography within rift zones can be broken up into extensive blocks separated by substantial escarpments. Large normal fault scarps define the steep inner shoulder of a rift leading down into closed, internally drained depressions that form deep sedimentary basins and elongated lakes parallel to the rift axis [**Photograph 12.7**].

Rift zones are geomorphically distinct. Rockslides typically contribute to high erosion rates on the steep slopes facing into the rift zone where deep canyons feeding numerous fans are found. Short, steep channels drain down the rift-margin escarpment and into a down-dropped chain of basins along the bottom of the rift, which can host either lakes (such as Lake Tanganyika in East Africa) or longitudinal rivers (such as the Rio Grande in New Mexico).

Regional drainage typically flows away from the uplifted rifted margins. Rivers and streams draining away from rift shoulders are characterized by relatively low gradients and minimal sediment supply. Warping due to thermal uplift of rift shoulders can lead to drainage reversals or captures along an actively developing rift system, with smaller channels more likely to be diverted due to their lower stream power and limited ability to incise bedrock. For example, small rivers originating near the Atlantic coast of South America first flow west down the upwarped rift flank away from the coast before turning to the east and emptying into the Atlantic Ocean—the result of capture by trunk streams breaching the uplifted rift shoulders.

Large rivers can incise rapidly enough to keep pace with the rising rift shoulders. For example, the major rivers of Africa and South America (with the exception of the Nile) flow toward the passive (trailing-edge) Atlantic margin over what was once the rift shoulder. Many of these rivers flow over great knickzones along the middle of their courses. While some of these knickzones are lithologically controlled, others reflect knickzone retreat from the original rift margin since the breakup of the supercontinent Pangaea more than 100 million years ago.

Passive continental margins unaffected by active tectonic processes develop after rifting has advanced sufficiently that oceanic crust is created at the spreading center and the two continents involved are drifting apart. Together, thermochronologic and cosmogenic nuclide data suggest that rift shoulders rapidly eroded back and then stabilized to form steep, long-lived escarpments characteristic of passive margins (such as those along the southern African or eastern Australia coasts). A slowly eroding, low-relief coastal plain typically lies on the ocean side of the escarpment. Above the escarpment are low-relief highlands that also erode slowly. The escarpment itself, however, is eroding somewhat faster and thus slowly retreats. Over time, through-going rivers dissect escarpments creating large topographic embayments and isolated plateau surfaces.

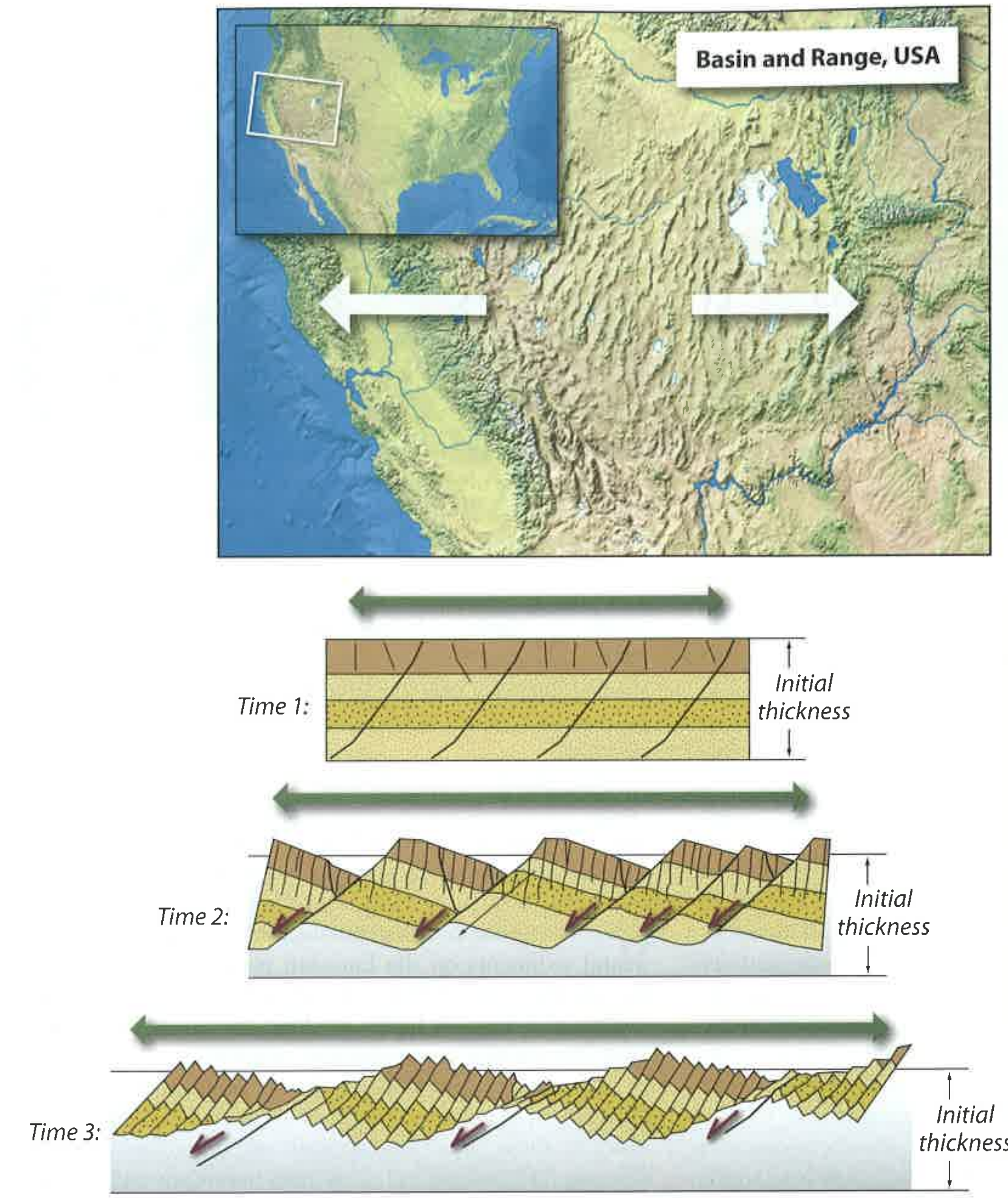
Lithology and structure greatly influence the geomorphology of passive margins that evolve from rifts. For example, in southern Africa and along the east coast of North America the passive margin runs parallel to the location of ancient mountain ranges. Today, rocks that were pervasively deformed and metamorphosed at depths of many kilometers are exposed on the surface, their coarse-scale structures (such as anticlines and synclines) clearly exposed by differential erosion.

Regional extension in continental settings results in distinctive large-scale topography even if it does not lead to rifting. For example, the linear north-south trending mountain ranges that collectively define the Basin and Range province of western North America are due to normal faulting resulting from east-west tectonic extension of the region. These fault-block mountains are thought to have formed from the stretching of a regionally extensive plateau [**Figure 12.5**]. In the Basin and Range, the valleys are generally tectonically down-dropped along normal faults (**grabens**), and the mountain ranges are uplifted as **horsts**, representing the toppled ends of gigantic tilted and back-rotated crustal blocks. Fault-bounded grabens typically form broad, flat-floored valleys that may create closed, internally drained depressions. Some, such as Death Valley in California, are below sea level.

Normal faults often form distinct fault-bounded range fronts. Where rivers or streams incise V-shaped valleys across such a range front, the intervening slopes are characterized by distinctive triangular facets [**Photograph 12.8**]. Along many extensional range fronts, active faulting visibly offsets

P. Bierman

PHOTOGRAPH 12.7 Dead Sea Rift. The internally drained Dead Sea and the adjacent steep, faulted rift flank in Israel. This is the steep, western (inboard) side of the rift. The view is to the north.



The **Basin and Range** physiographic province extends from the Rocky Mountains to the Sierra Nevada. The distinctive array of linear north–south basins (**grabens**) and ranges (**horsts**) reflects substantial east–west extension, with the crust being stretched to as much as twice its original width. The extension involves the uppermost 20–30 km of the continental crust.

Extension of an initial high-elevation plateau along sets of nested normal faults has resulted in a fallen-down block structure in the Basin and Range. Alternating series of down-dropped basins and uplifted ranges reflect rotational deformation along a complex set of deep-seated regional **normal faults** and result in a landscape of fault-controlled, alternating valleys and mountain ranges.

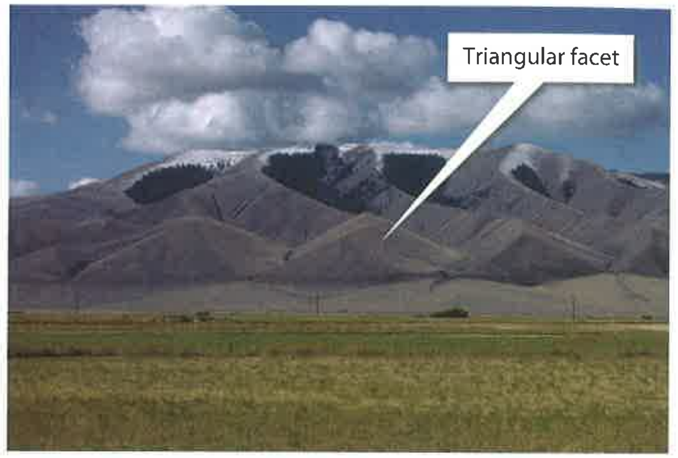
FIGURE 12.5 Tectonic Extension—Basin and Range Province. The Basin and Range physiographic province of western North America is the product of east-west extension that produced numerous north-south trending ranges and valleys.

moraines or fans developed at the topographic break in slope that separates bedrock uplands from alluvial lowlands [**Photograph 12.9**]. Dating such offset features provides a means of calculating rates of fault movement and patterns of fault offset.

Compressional Margins and Landforms

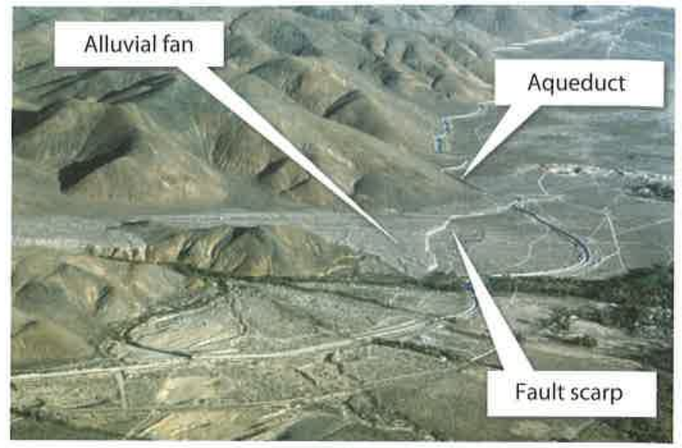
Compressional **orogens**, mountain ranges composed of linear belts of highly deformed rock, form along collisional plate boundaries where tectonic convergence leads to thickening of the crust that, in turn, elevates mountains. Where lithospheric plates converge, tectonically active mountain ranges are generally arrayed as linear belts of high elevation and significant topographic relief. Examples include the western margins of North and South America and continental collision zones like the Himalaya. Mountains at active compressional orogens generally consist of deformed sedimentary rock intruded or interlayered with igneous rock.

Compressional orogens typically have high rates of deformation and uplift. This uplift drives river incision that, in turn, leads to steep slopes and frequent landsliding. These mass movements deliver large sediment loads to steep, energetic rivers with high flows capable of rapidly



M. Miller

PHOTOGRAPH 12.8 Triangular Facets. Triangular facets on a range front from normal faulting at the base of the Tendoy Mountains in southwest Montana.



P. Bierman

PHOTOGRAPH 12.9 Faulted Fan. Lone Pine strand of the Owens Valley fault zone offsets late Pleistocene alluvial fan. The white fault scarp (center, at the foot of the hills) cuts across the Los Angeles aqueduct, which runs around the toe of the fan.

transporting all the sediment delivered to them. In active orogens, deeply incised rivers commonly flow through narrow bedrock valleys or deep gorges cut into rock. Different types of convergent orogens have different topographic characteristics that we describe below [Figure 12.6].

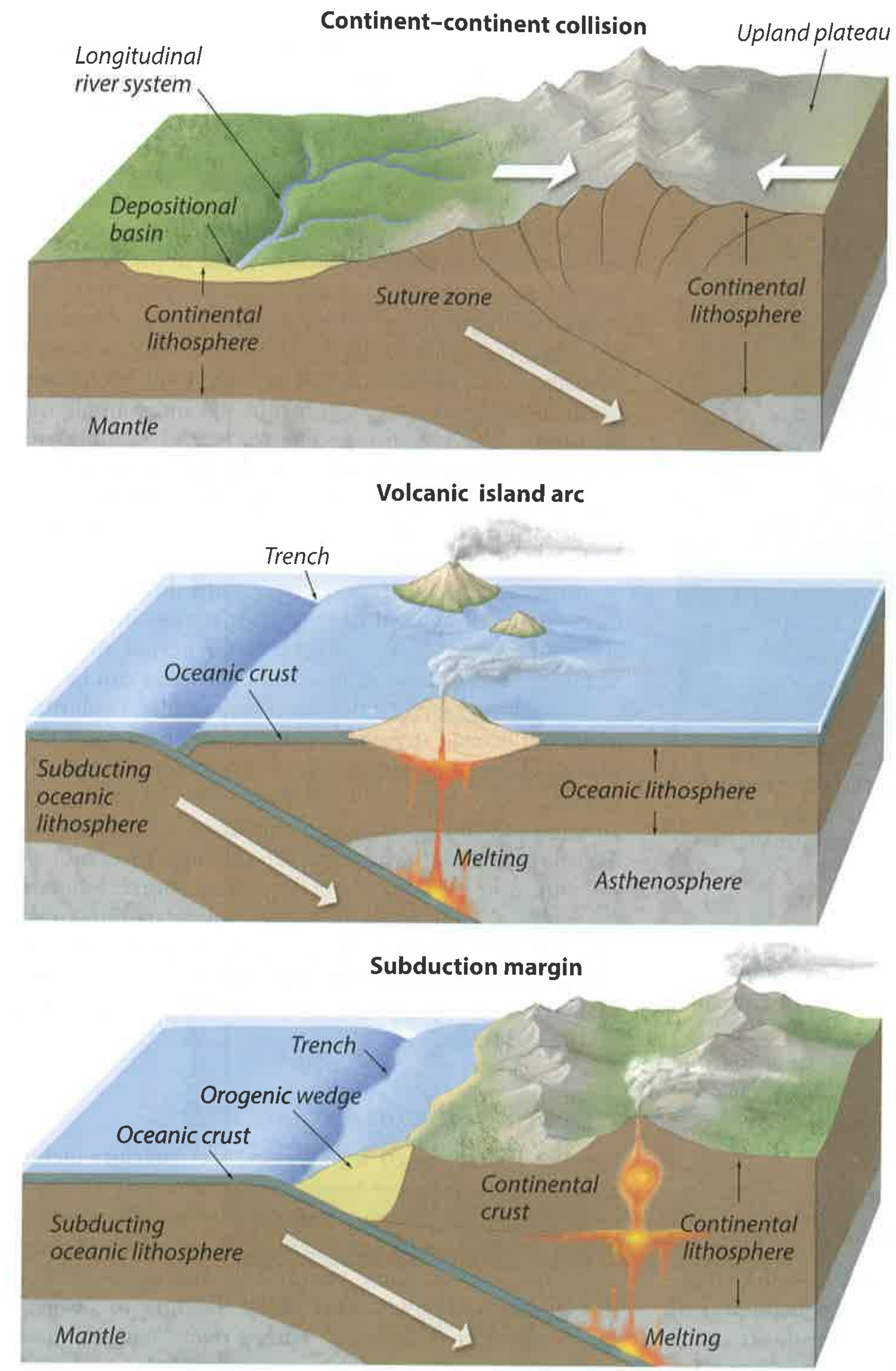
When two plates carrying continental crust converge, neither is as dense as the underlying mantle and thus neither will readily sink. Consequently, where continents collide, material piles up and the crust thickens, leading to dramatic high-standing mountain ranges like the Himalaya. Continental collisions lack significant volcanic activity and have a zone of thrust faulting that typically separates the upland drainage system of bedrock channels from the alluvial channels flowing across the sedimentary basin that develops in the depositional foreland where rivers leave the uplands. Longitudinal drainage paralleling the strike of a mountain system can develop because of orogen-parallel flexural subsidence of the foreland; in contrast, short, steep channels typically drain the mountain front. Large rivers that cross major mountain ranges (like the Indus and Tsangpo/Brahmaputra that begin on the Tibetan Plateau and flow across the Himalaya) are generally thought to be **antecedent rivers** older than the mountains (see Chapter 7). These rivers were able to maintain their courses by vigorous erosion across the rising mountain range. Landforms typical of compressional orogens include upland bedrock channels and hillslopes with thin soils and little if any saprolite (weathered rock) due to rapid erosion and therefore short residence times of materials on slopes.

In contrast, volcanic arcs form along subduction zones where an oceanic plate subducts beneath either a continental or oceanic plate. Volcanic arcs may be either continental, in which an oceanic plate subducts beneath a continental plate, or an **island arc,** in which an older, colder, and denser oceanic plate subducts beneath another plate that is younger, warmer, and thus more buoyant. In either case, a wedge of sedimentary rock scraped off the downgoing, subducting plate can rise seaward of the mountains of the volcanic arc that parallels and is fed by the subduction zone.

The island of Taiwan is an example of a **sedimentary wedge** formed in an ocean-continent collision. Mountain building and erosion in Taiwan are particularly rapid because of the combination of a wet tropical climate, weak rocks, and a huge tectonic influx of material into the orogenic wedge due to the great thickness of incoming continental sediments on the Eurasian plate. In regions where the incoming plate has a lot of sediment on top, thick wedges of deformed, highly erodible, fractured sedimentary rock characterize compressional orogens.

The rocks exposed along the spines of active volcanic arcs consist of massive lava flows in oceanic island arcs, and more erodible pyroclastic deposits in continental arcs (see Chapter 11). Some deeply exhumed mountain ranges, such as the British Columbia Coast Range of Canada or the Sierra Nevada of California, expose hard, relatively massive undeformed rocks (like granite) that represent the exhumed roots of ancient volcanic arcs. The Sierra Nevada, for example, consists of the eroded roots of an ancient continental volcanic arc (like the modern Cascade Mountains of Oregon and Washington State) that was shut off from its magma source more than 65 million years ago.

Reverse (or thrust) faulting is the dominant form of tectonic displacement in compressional orogens, and typically has indirect topographic expression. **Thrust faults** are shallow reverse faults that place older rocks over younger rocks [Photograph 12.10]. Thrust faults can extend over hundreds of kilometers, but typically do not produce distinct fault scarps. Many reverse faults either terminate in **blind thrusts** (that do not produce scarps) before reaching the surface or splay out into multiple fault traces near the surface.



Continent–continent collisional margins, such as the Himalaya, exhibit a complex **suture zone.** If crustal thickening proceeds far enough, these margins develop an extensive upland **plateau** surface. Longitudinal river systems form in mountain-front depositional basins that parallel the range front.

Subduction margins where an oceanic plate dives beneath another oceanic plate create **volcanic island arcs,** such as Japan and many South Pacific islands. Adjacent and parallel to the arc is a deep trench.

Where oceanic lithosphere sinks beneath continental lithosphere, there are terrestrial volcanic arcs, such as the Cascade Range in western North America. A sedimentary **orogenic wedge** can form a subsidiary mountain range between the trench and volcanic arc, such as the Olympic Mountains and the Oregon Coast Range.

FIGURE 12.6 Collisional Tectonic Settings. Collisional tectonic settings include continent–continent collisions with a major mountain range bordered by a foreland sedimentary basin, ocean–ocean collisions forming oceanic island arcs, and continental-oceanic subduction zones.

Reverse faults can have substantial topographic expression at large scales, such as the topographic front of the Himalaya, which is an impressive escarpment generated by long-term offset along thrust faults. In regions where rates of deformation are higher than rates of erosion, compressional tectonic deformation and folding may directly affect topography. In such cases, structural synclines (troughs) form valleys, and the crest of structural anticlines (arches) form topographic ridges, the best known of which are in the Zagros Mountains in Iran and the Ventura anticline in

PHOTOGRAPH 12.10 Thrust Fault. Reverse (thrust) fault in the northern Gulf of Corinth between Athens and Delphi in Greece. The unit below is clastic sedimentary rock. The unit above is a deformed fault mélange of originally sedimentary material. The fault plane is reddish brown.

southern California. Where a river breaches an anticlinal ridge, surveying and dating sets of river terraces that have been bowed upward and deformed across the rising fold (anticline) reveal the pattern of uplift centered on the crest or hinge of the anticline and allow calculation of the rate of surface uplift [**Figure 12.7**].

Transform Margins and Landforms

Plates slide past one another at transform margins, where strike-slip faults dip steeply creating a distinctive set of landforms [**Figure 12.8**]. Transform margins can be simple with one distinct fault trace or complex with activity distributed along a series of faults that define a broad shear zone. Such shear zones produce distinctive landforms and topography resulting from lateral movement.

A regional component of extension or compression across a transform margin produces distinct extensional or compressional features. **Transtensional margins,** such as the Gulf of California, have a string of down-dropped basins, or grabens, along the shear zone. In contrast, compressional transform margins (**transpressional margins**) typically have fault-zone parallel mountains due to compression distributed across the shear zone. A good example of transpressional transform margin topography are the central California Coast Ranges that extend to either side of the main San Andreas Fault, reflecting compression across the margin and complex interactions between fault strands.

On a more local scale, as material moves laterally through bends along a strike-slip fault, the geometry of the fault system creates zones of local uplift or subsidence where compression or extension occurs. This results in the development of compressional pop-up structures in **restraining bends** and extensional pull-apart basins in **releasing bends** [**Figure 12.9**]. The Santa Cruz Mountains, south of San Francisco, California, are an example of a mountain block uplifted due to lateral compression through a restraining bend along the San Andreas Fault. Conversely, San Francisco Bay lies in a subsiding area (transtensional) between the San Andreas and Hayward faults along the same fault system.

Strike-slip faults are associated with distinctive landforms that can be used to map fault traces. Streams that cross strike-slip faults may be offset across the fault, resulting in distinctive right-angle bends that can be used to determine offset directions and magnitudes. Fault offset may even behead streams, separating channels from their sources. Trenching and dating of truncated or offset alluvial deposits along such **offset streams** can reveal the amount and timing of movement across a fault. Distinctive **shutter ridges** form where lateral fault offset moves a ridge in front of a stream, deflecting its course. Similarly, **sag ponds** develop where drainage courses are impounded or small depressions form along the trace of strike-slip faults [**Photograph 12.11**]. Transform faults also occur in marine settings where they cut and laterally offset mid-ocean ridges.

Continental Interiors

Landforms associated with continental interiors include deeply weathered low-relief terrain, areas of extensive loess deposition, and large alluvial rivers flowing across broad floodplains (where there is an extensive orogenic sediment source). Some continental interiors have both low slopes and low overall relief, leading to development of internal drainage where rivers end in closed depressions.

Continental interiors in humid, tropical regions develop thick weathering profiles, whereas those in arid regions generally have bare rock slopes. Consequently, overland flow in arid landscapes and diffusive processes in humid landscapes (see Chapter 5) dominate hillslope transport as there typically is only limited, localized landsliding due to the predominance of gentle slopes. Weathering processes greatly influence landforms in such low-gradient environments, and isolated high-standing rock outcrops, known as **inselbergs** (see Chapter 3), tend to be best developed in continental interiors [**Photograph 12.12**].

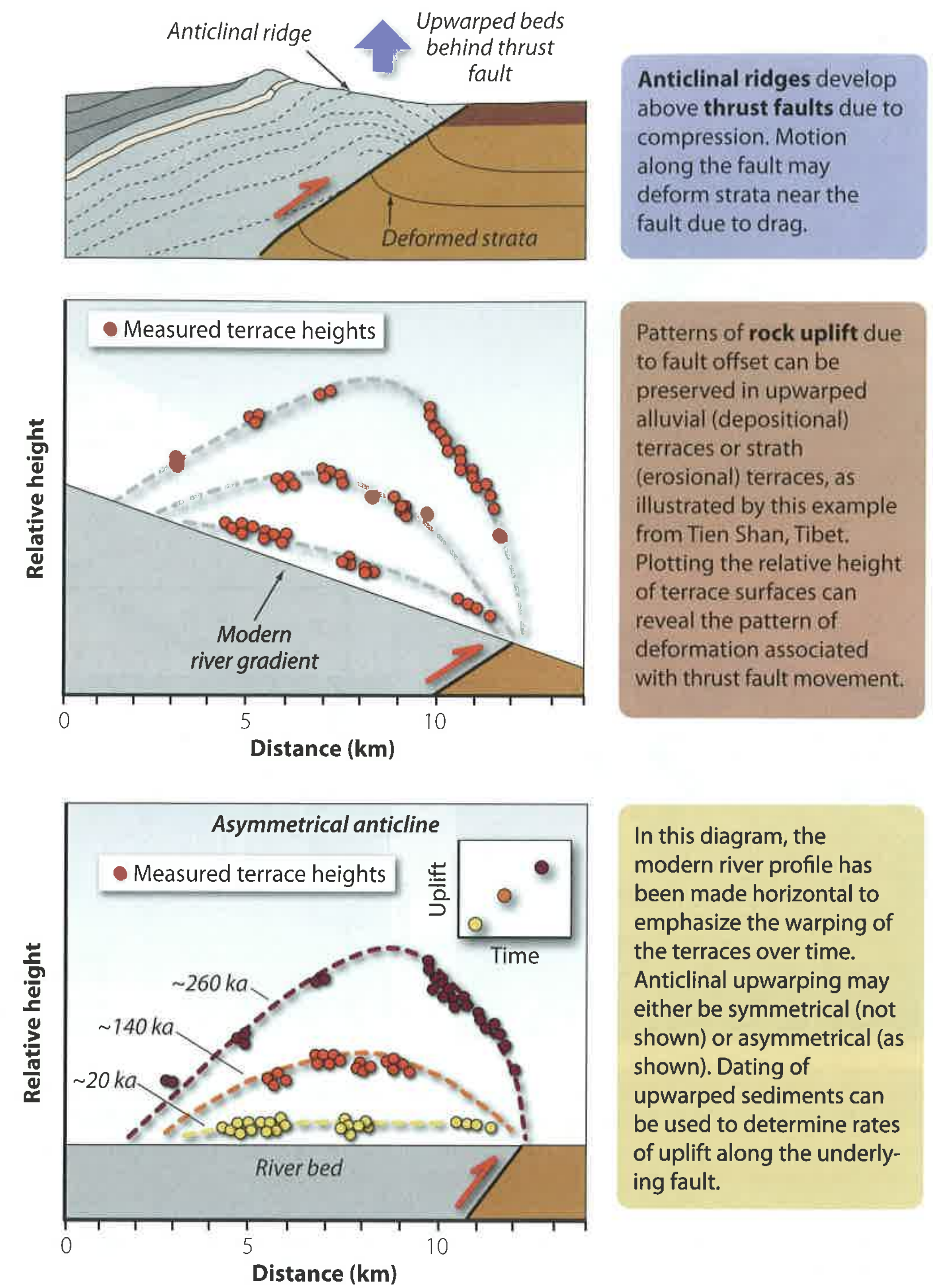
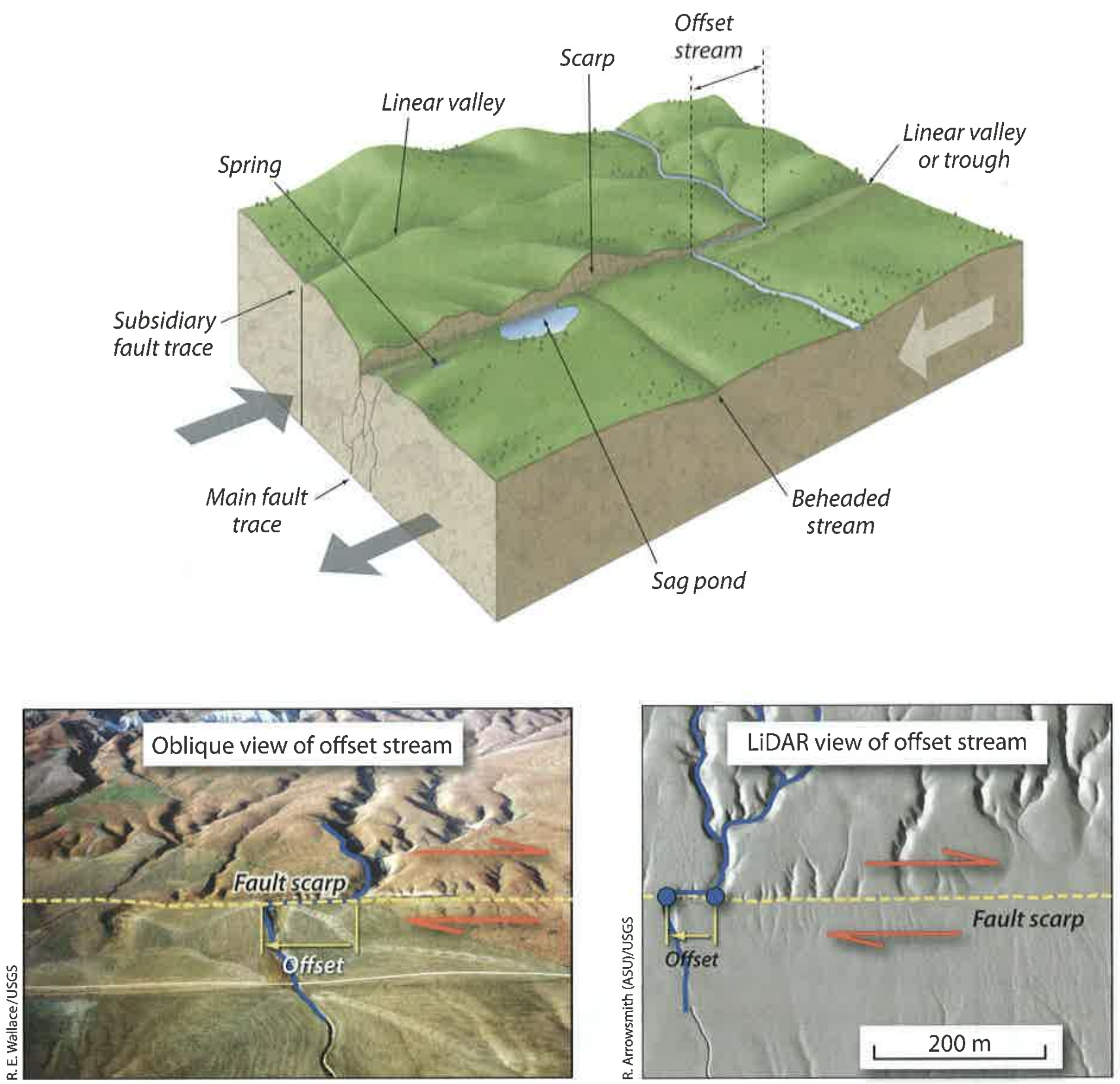


FIGURE 12.7 Anticlinal Ridges. Uplift of anticlinal ridges associated with thrust faults may deform stream terraces. The age and pattern of deformation of these terraces may be used to investigate the history of uplift along the fault. [Adapted from Molnar et al. (1994) and Burbank and Anderson (2001).]

The vast interiors of continents primarily consist of relatively flat areas known as **cratons**—tectonically stable regions of relatively low relief, typically rising no more than a few hundred meters above sea level. Cratons are underlain by ancient continental crust with the low-elevation and low-relief terrain characteristic of regions of prolonged tectonic stability. Cratons consist of **shields,** areas composed of complexly deformed crystalline rocks, and **platforms,** areas where younger, undeformed sedimentary rocks lay on top of far older crystalline basement rock.

Plateaus are high-elevation, low-relief surfaces separated from surrounding land by steeper slopes. The world's high elevation plateaus—Asia's Tibetan Plateau and South America's Altiplano—are tectonically constructed landforms



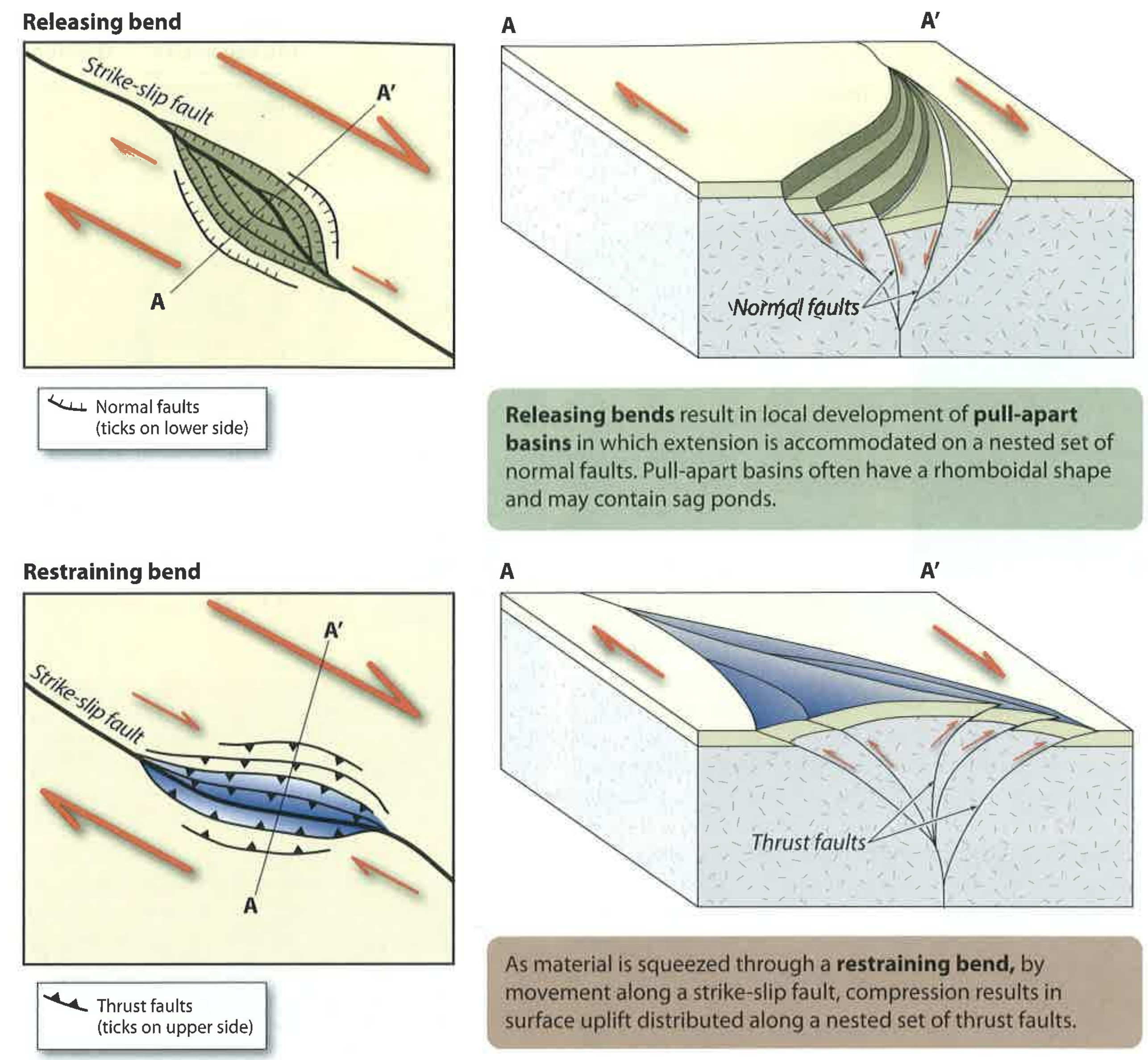
Strike-slip fault zones produce a characteristic suite of landforms that reflect both linear offset (**beheaded streams, offset channels**) and weakening of rock or tectonic extension near the fault (**linear valleys, sag ponds**). LiDAR and aerial photographs can reveal the location and amount of recent offset on strike-slip faults.

FIGURE 12.8 Strike-Slip Fault Zones. Characteristic landforms of strike-slip fault zones include sag ponds and offset stream channels, as well as evidence of changes in hydrology such as springs.

that formed when the lithosphere thickened enough that its strength limits further increases in elevation [Figure 12.10]. When continental crust reaches a height of 4 to 5 km, geothermal heating weakens the base of the lithosphere enough to make it susceptible to lateral extrusion upon further thickening. This effectively limits the mean height of topography, as further thickening leads to lateral flow at depth that controls plateau elevation. Once a plateau reaches this critical height, continued tectonic convergence leads to widening of the plateau.

Some continental interiors have substantial topography reflecting in large part the location of former orogens, now tectonically inactive. The combination of isostasy and a thick root ensures that tectonically inactive mountain ranges persist, lasting far longer than one would calculate by considering erosion rates alone (thanks to isostatic rock uplift offsetting ~82 percent of erosion).

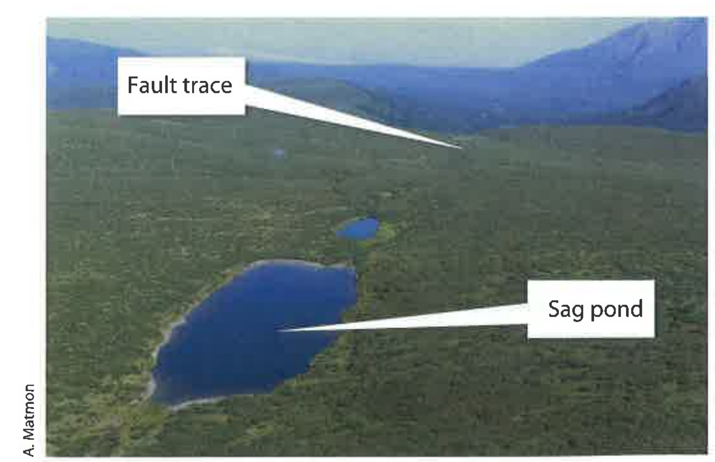
The interiors of continents are for the most part tectonically quiescent. Earthquakes and fault scarps are few and far between and many are associated with old structures



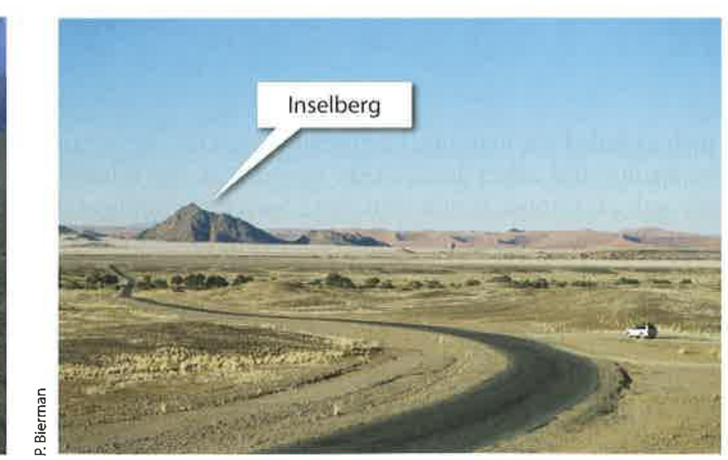
Releasing bends result in local development of **pull-apart basins** in which extension is accommodated on a nested set of normal faults. Pull-apart basins often have a rhomboidal shape and may contain sag ponds.

As material is squeezed through a **restraining bend,** by movement along a strike-slip fault, compression results in surface uplift distributed along a nested set of thrust faults.

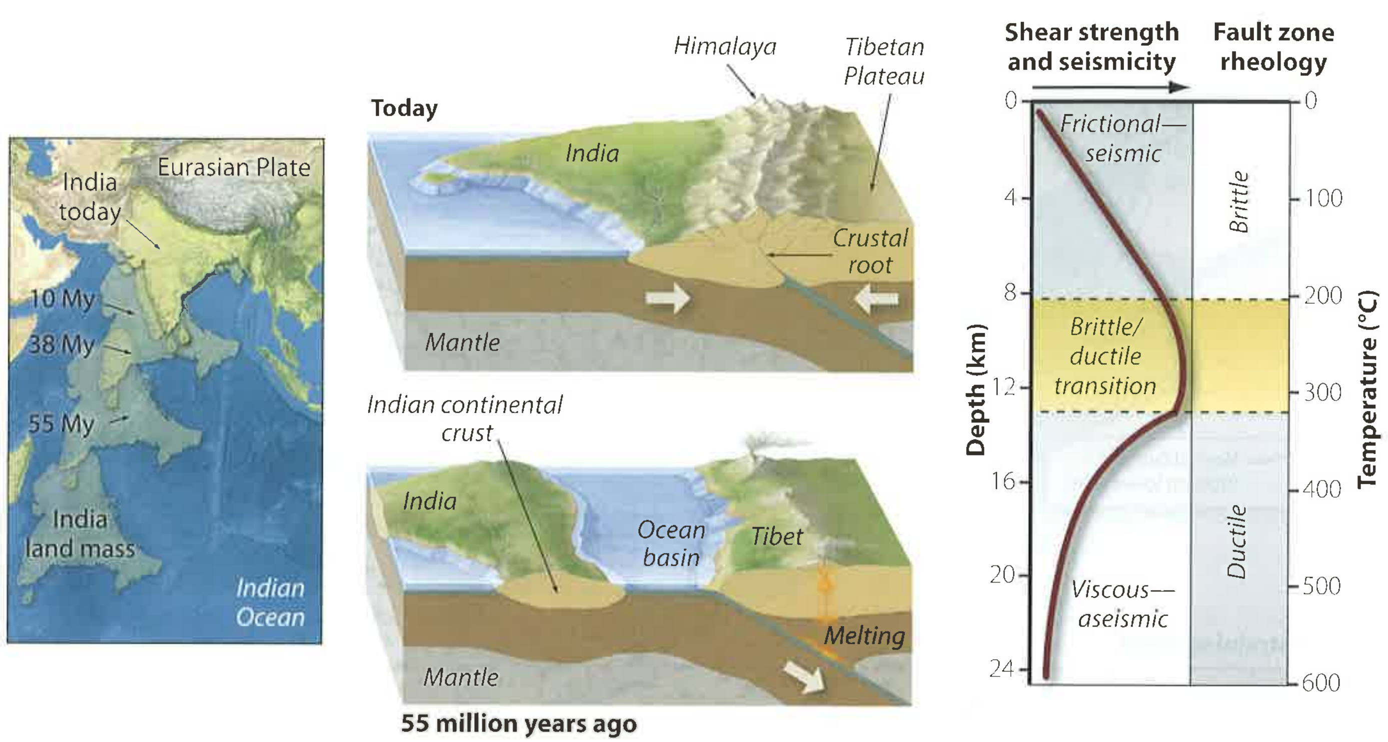
FIGURE 12.9 Restraining and Releasing Bends. Interaction of strike-slip fault geometry and topography produces zones of structural uplift that result from compression at restraining bends whereas subsidence occurs in extensional pull-apart (releasing) bends.



PHOTOGRAPH 12.11 Sag Ponds. The Denali Fault (Alaska) runs obliquely across this image, and a line of sag ponds mark its trace (image is several kilometers across).



PHOTOGRAPH 12.12 Inselberg. An inselberg rising above the low-relief coastal plain of Namibia near the dunes of Sossevlei. Sport utility vehicle is shown for scale.



As India moved into Asia, the continental lithosphere of both continents was not dense enough to subduct and ongoing collision resulted in **crustal thickening.** Over time, the **crustal root** grew thicker and the Tibetan Plateau rose and, as the collision continues, it grows wider. The thickening crust, rising mountains, and high rates of erosion affect the shape and behavior of the Himalaya. Rapid erosion advects warm rock toward the surface, and the thickened crust causes rock at depth to cross the brittle/ductile transition. This change in the process and rate by which the rock deforms allows rock to more easily and rapidly extrude out the sides of the range, removing some of the mass brought in by continental collision and limiting the height of the Tibetan plateau.

The shear strength and rheology (flow characteristics) of the lithosphere vary with depth due to the effects of increasing pressure and temperature. In the uppermost brittle crust, frictional strength increases to a depth of around 8 to 10 km. At higher pressures and temperatures below about 13 km depth, shear strength decreases and the crust becomes ductile, producing aseismic viscous flow. Tectonic convergence that produces exceptionally thick crust can lead to lateral flow of ductile material at depth.

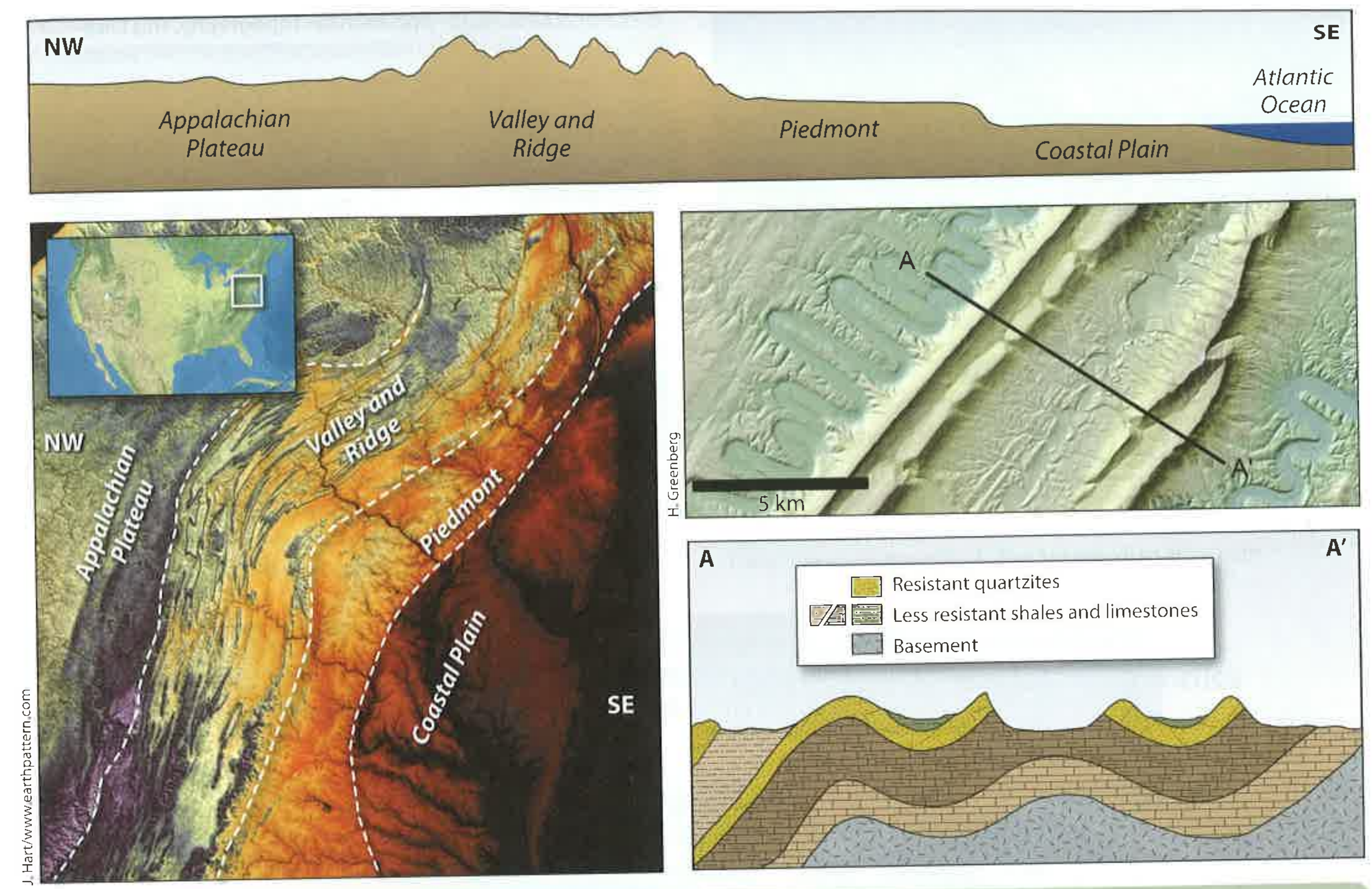
FIGURE 12.10 Formation of the Tibetan Plateau. The Himalaya and Tibet formed as a result of crustal thickening sufficient to change solid-Earth behavior at depth, limiting the crustal thickness supportable by the underlying mantle and thereby forming the high-elevation, low-relief Tibetan Plateau.

such as failed rift margins. Earthquakes that do occur can be strong and affect large areas because of the relative strength of cratonic, continental crust. Such mid-continental quakes have caused liquefaction of weak, wet sediments (for example, in the New Madrid Fault Zone in south-central North America) and left distinctive linear fault scarps (as at Tennent Creek in northern Australia).

Structural Landforms

After tectonically driven rock uplift declines or ceases, structure and lithology exert a pronounced influence on topography as rates of weathering and erosion outpace rates of rock uplift, folding, and deformation. Different lithologies or structures determine slope form as preferential erosion of weaker rocks leaves more resistant rocks standing out in relief. In regions with large variations in erosion resistance, steeper slopes tend to form on more erosion-resistant rocks, whereas gentler slopes characterize more erodible formations.

The topographic expression of deformation structures, such as folds, results from differential erosion of rocks. For example, geologic structure is readily apparent in the Appalachian Mountains of eastern North America, where large differences in the erodibility of the underlying geology govern ridge and valley patterns [Figure 12.11]. Resistant



The tectonically inactive Appalachian Mountains rise from the sedimentary coastal plain and consist of the broad, low-elevation upland of the **Piedmont,** the higher-elevation **Valley and Ridge** province that defines the spine of the range, and the low-relief **Appalachian Plateau** on the western edge of the range. The Appalachian Mountains have been tectonically quiescent since the opening of the Atlantic Ocean.

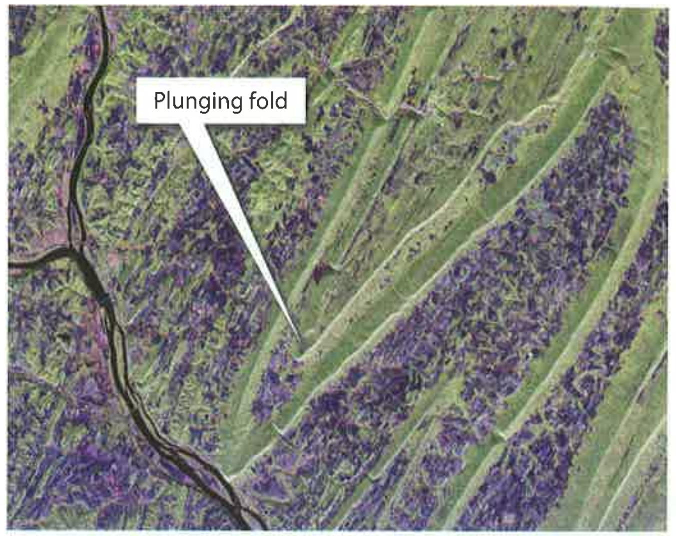
The topography of the Valley and Ridge province reflects the underlying geology, with ridges formed by erosion-resistant rocks (such as quartzites) and valleys formed by more erodible shales and limestones. The correspondence between rock strength and topographic expression led many geologists to presume, correctly or incorrectly, that old mountains have reached a state of equilibrium and erode everywhere at the same rate.

FIGURE 12.11 Tectonically Inactive Mountains—the Appalachians. The valley and ridge morphology of the Appalachian Mountains reflects the strong influence of geological structure and lithology.

units (sandstone and quartzite) form ridges and weaker units (shale and limestone) form valleys [Photograph 12.13]. Over time, stream patterns become adjusted to the underlying geologic structure, and streams preferentially flow along the weakest rocks or in fracture zones. In general, there is greater structural control on landforms in ancient, no-longer-active orogens than in active orogens because in rapidly uplifting regions like the Himalaya and Taiwan uniformly steep (threshold) slopes obscure the underlying geologic structure and patterns of rock uplift.

As erosion cuts through hard, ridge-forming beds, it can expose weaker underlying rocks, which can then erode quickly. This can lead to a topographic inversion, in which geologic structure and topography are out of alignment. Where erosion breaches the crest of a structural anticline, exposure of more erodible rocks along the crest of the anticline can produce a topographic low along the structural high through excavation of an **anticlinal valley.** Similarly, a **synclinal ridge** reflects the preservation of structurally low but erosion-resistant rocks as a topographic high in the structural trough of a fold.

The inclination or dip of underlying beds can greatly influence landform development [Figure 12.12]. **Monoclines** occur where the dip of strata increases locally, but the beds do not turn over, producing a structural and topographic step. The relative inclination of stratified (layered) rocks



PHOTOGRAPH 12.13 Appalachian Topography. This false-color radar image shows the large, plunging folds of the Valley and Ridge physiographic province near Sunbury, Pennsylvania. The ridges are held up by resistant sandstone and the valleys are dominated by shale and limestone. Black trace at left of image is the Susquehanna River cutting across structure. The view is about 30 km wide and north is to the upper lefthand corner.

relative to the hillslope angle influences slope forms. Erosion of gently dipping beds leads to the development of pronounced dip slopes where the hillslope angle parallels rock bedding. **Cuestas** are asymmetric slopes that are elongated in the down-dip direction. More symmetric slopes where the underlying beds dip approximately 45 degrees are known as **hogbacks.** Distinctive **flatirons,** named due to their resemblance to the now antiquated household appliance, develop from differential erosion of a resistant rock layer that dips more steeply than the hillslope angle. In flat-lying strata, erosion-resistant cover beds that protect underlying rock lead to the development of **mesas,** flat-topped plateaus like those of northern Arizona. Mesas are often protected by an erosion-resistant cap rock that retards erosion on the mesa top, leading to steep-walled sides. Continued erosion of a mesa can lead to an isolated summit area known as a **butte,** like the famous outcrops of Monument Valley in Utah and Arizona. In arid landscapes, where soil mantles are thin and slope form reflects bedrock properties (see Chapter 5), the influence of differential erosion resistance in flat-lying layered sedimentary rocks can be expressed as **cliff-and-bench topography,** in which erosion-resistant rocks (like well-cemented sandstone and lava flows) form cliffs, whereas weaker interstratified rocks (like shale) form benches.

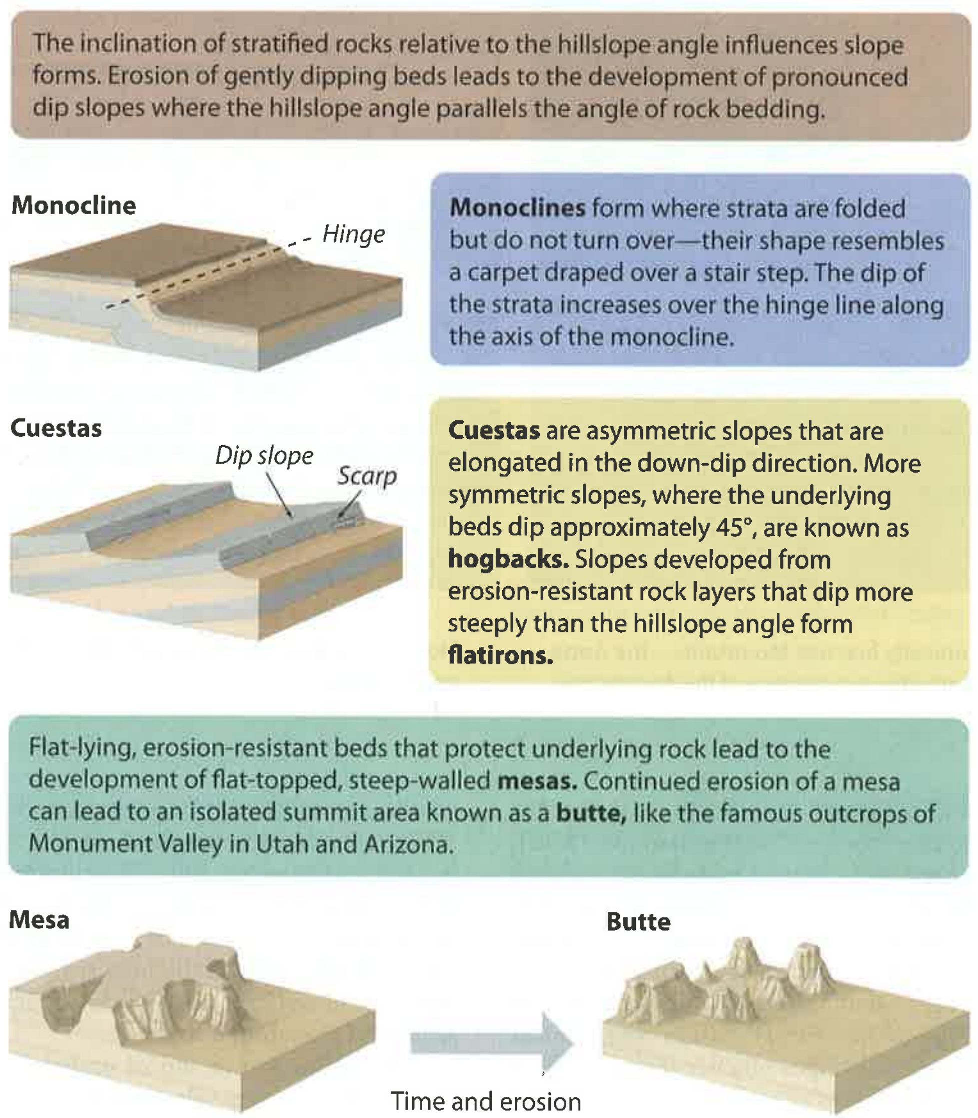


FIGURE 12.12 Structural Landforms. The orientation of dipping beds of rock can control the shape of landforms, especially in areas where soil cover is minimal.

(a)

(b)

PHOTOGRAPH 12.14 Joint-influenced weathering patterns and landforms. (a) Checkerboard weathering pattern caused by chemical weathering along intersecting joint sets in the El Capitan Granite, Yosemite, California. Trees provide scale. (b) Fins left by preferential erosion of joints in sandstone in the Maze District, Canyonlands National Park. The fins are 20 to 40 m high.

Joints, fractures in rocks along which little or no motion has occurred, can result from regional tectonic stresses or shrinkage upon cooling, and can have significant topographic expression. Joint-related landforms are particularly visible in arid environments with bedrock hillslopes. Regional tensile stresses may produce joints in even relatively strong rocks because the tensile strength of rock is far less than its compressive or shear strength. On a smaller scale, igneous rocks often have well-developed joint sets due to shrinkage upon cooling, for example, columnar joints in thick, slowly cooled, basalt flows (see Photograph 11.15). Joints can greatly influence differential weathering and erosion because they provide conduits along which water and plant roots can preferentially penetrate, break up, and erode rock masses (see Chapter 3). In rocks with well-developed vertical jointing, continual erosion along joint surfaces can produce checkerboard weathering patterns [Photograph 12.14a] or leave the unjointed rock standing in relief as narrow fins [Photograph 12.14b], the jointed (more fractured) rock having been removed more rapidly by erosion.

Landscape Response to Tectonics

Tectonics affects landscapes in a variety of different ways. Tectonically induced base-level fall and uplift can increase topographic relief and steepen river profiles and hillslopes (see Chapter 7). Uplift of the land surface relative to local or global base level (surface uplift) will trigger an increase in local slope that sweeps up through a river system, driving a wave of incision toward the headwaters of a drainage basin as the increased topographic gradient (slope) influences the rate and type of erosional processes.

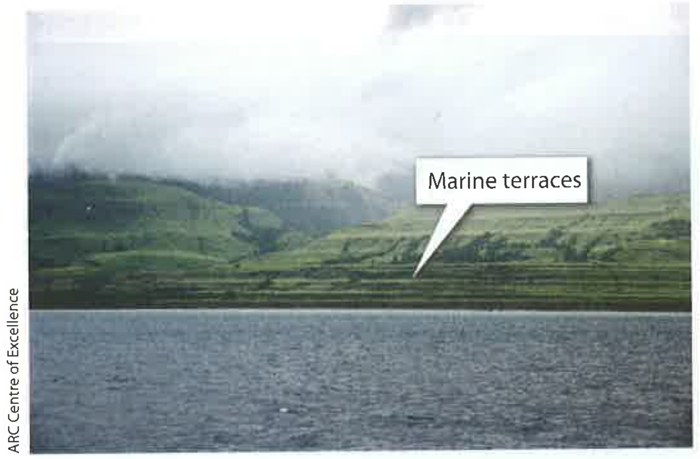
Tectonic forcing can reorganize drainages and change drainage-basin boundaries. For example, continental rifting opens new basins and splits existing drainages. Tectonic tilting can reverse a river's flow direction or cause one stream to erode headward more rapidly than another, eventually beheading the captured stream and creating a new and expanded drainage network. Such truncated basins are common in tectonically active areas and are often identified by fluvial gravels stranded at high elevation, far above any modern-day stream channels. A good example can be found in the northern Galilee in Israel, where basaltic gravel from the Syrian highlands sits atop the local limestone bedrock, even though today there is no fluvial connection between the basalt outcrops and the gravels. More subtle responses to tectonic forcing are common in drainage basins where a rise in base level leads to aggradation and basin filling, whereas a falling base level causes incision and the creation and preservation of terraces.

Topographic slope drives erosional processes more directly than elevation. The effect of elevation on erosion occurs indirectly through its control of climate. Flat surfaces at high altitude erode slowly, whereas steep slopes at low elevation erode relatively quickly.

Coastal Uplift and Subsidence

Changes in relative sea level, whether from global sea-level change, tectonic rock uplift, or subsidence, leave their mark on coastal landscapes. The coastlines of many tectonically active continental margins consist of rocky coasts where actively eroding sea cliffs define the shoreline angle (see Chapter 8). During times when sea-level rise matches the rate of coastal rock uplift, wave action can abrade a broad, gently sloping (~1 degree) wave-cut platform that extends seaward from the base of the sea cliff to below the level of tidal influence. If sea-level rise is less than the rate of coastal rock uplift, then the sea cliff and wave-cut platform are abandoned, which results in the formation and exposure of a low-relief marine terrace with a fossil sea cliff separating it from the next higher terrace.

Over multiple glacial/interglacial cycles of global sea-level rise and fall (due largely to variations in the amount of glacial ice), sequences of marine terraces can form on actively rising coastlines, with younger terraces closest to sea level and older, increasingly eroded terraces at progressively higher elevations. The series of marine terraces formed by uplifted coral reefs on the coast of New Guinea provides a classic example of landforms resulting from the interaction of changing sea level with tectonically driven surface uplift [**Photograph 12.15**]. On this and many other rising, tectonically active coastlines, such as those of northern California and Japan, the heights of individual marine terraces can be correlated with sea-level high stands; such correlation allows one to estimate coastal uplift rates back through time from the relative elevations of different marine terraces [**Figure 12.13**].



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PHOTOGRAPH 12.15 Marine Terraces. Flights of marine terraces along the rapidly uplifting Huon Peninsula on the northeastern coast of New Guinea. The highest terrace is now several hundred meters above sea level and hidden under the low clouds.

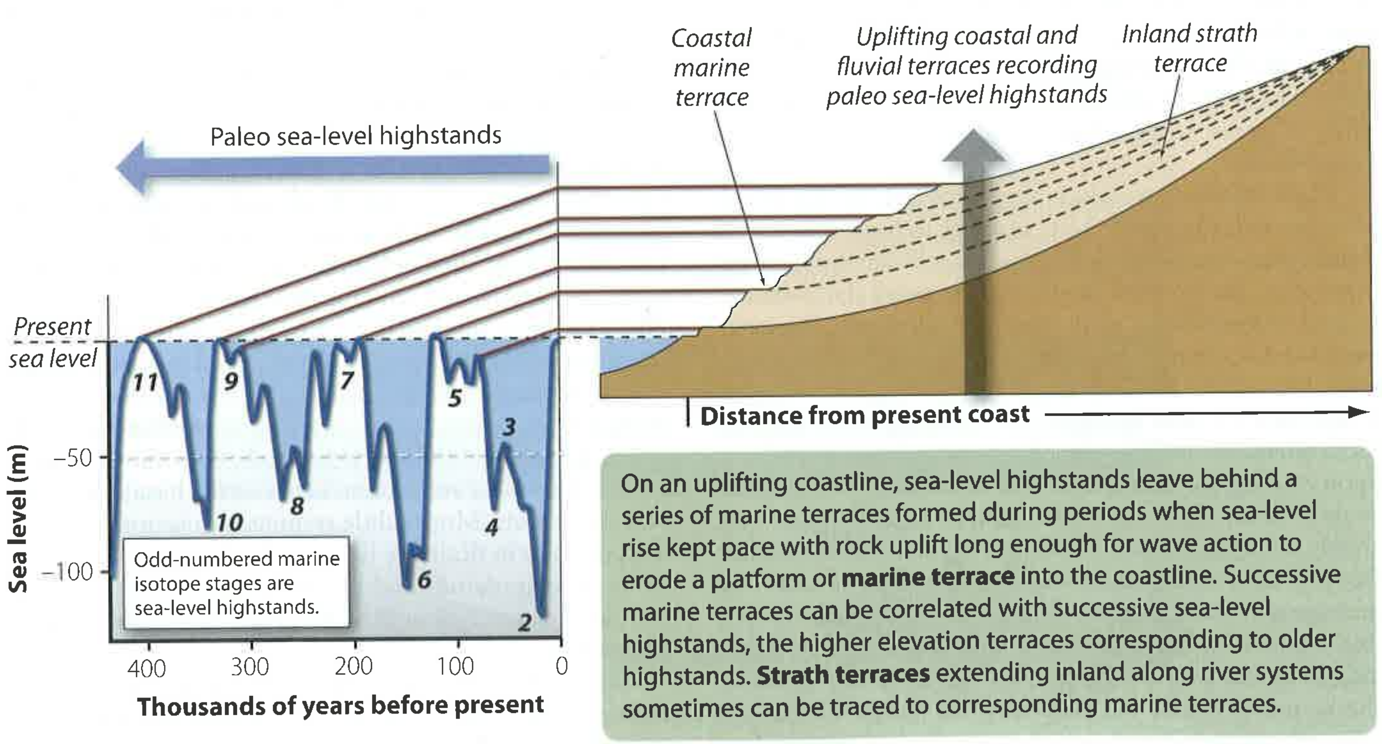

FIGURE 12.13 Marine Terraces Along Uplifting Coastlines. On an uplifting coastline, the elevation of marine terraces and inland bedrock strath terraces may be correlated with past sea-level highstands to estimate their age and the local uplift rate.

Coastal subsidence can occur for a variety of reasons. Subsidence can accompany large subduction-related earthquakes along active margins. Drowned forests and marsh deposits covered by layers of tsunami-deposited sand along the coasts of northern California, Oregon, and Washington testify to long periods of slow coastal uplift separated by episodes of rapid subsidence during large earthquakes [**Photograph 12.16**]. Such subsidence can result in brief periods of marine inundation even on an actively uplifting coast. Long-term subsidence due to sedimentary loading and compaction from ongoing deposition characterizes large estuaries such as the Mississippi River delta. Subsidence along the mid-Atlantic coast of North America reflects the slow decay of the glacial forebulge, formed as a result of the displacement of mantle material from beneath the Laurentide Ice Sheet in the Late Pleistocene.

B. Atwater/USGS

PHOTOGRAPH 12.16 Ghost Forest. Ghost forest along the Copalis River, southwestern Washington. The dead trees (snags) were victims of tidal submergence caused by tectonic subsidence during the great CE 1700 Cascadia subduction-zone earthquake. The trees died when their roots were immersed in salt water as the land subsided during the earthquake.

Rivers and Streams

Rivers and streams respond to tectonic forcing through adjustments in slope. Spatial variations in rock uplift along a river profile can lead to local reaches steeper, or flatter, than expected along a river's longitudinal profile. A plot of the downstream values in the stream gradient index (see Chapter 7) can identify such anomalous reaches where faults or lithology might be influencing channel slopes. Methods for analyzing such deviations also include DEM-based drainage area-slope analyses [**Box 12.1**]. Steep sections of a river profile tend to erode faster than gentler reaches, causing knickpoints (see Chapter 7) to migrate upstream and in extreme cases can lead to one river capturing and diverting the flow from another.

The relationship between a drainage basin and its base level can change through tectonic subsidence, uplift of the land, or a rise or fall in sea level. Base level changes greatly affect the locus of sedimentation in coastal zones and alluvial rivers. In contrast to how the effects of a base-level drop propagate up through the channel network, base level rises primarily affect reaches near river mouths, drowning deltas and turning coastal valleys into estuaries and bays. Rivers respond to a rising base level through deposition in their lower reaches, building a depositional wedge or ramp of sediment [**Figure 12.14**]. Above this depositional zone, the channel network remains uninfluenced by the change other than now being graded to a higher base level.

Along rivers discharging to an uplifting coastline, flights of strath (bedrock) terraces abandoned above the modern river floodplain record progressive incision of the river. Such terraces can be correlated by height, just like marine terraces, and often can be traced up through a river network. By dating terraces at different elevations, one can determine river down-cutting rates and, by inference, surface and rock uplift rates.

River-channel morphology can serve as a sensitive indicator of tectonic deformation because channel patterns are closely related to both channel slope and sediment supply (see Chapter 6). For example, in response to localized steepening of their longitudinal profile, braided channels tend to incise and convert to a single-thread channel, whereas meandering channels may increase in sinuosity or braid before incising in response to tectonic steepening. More extreme changes in channel gradient can convert an alluvial channel reach to a bedrock channel reach. Greater increases in slope lead to development of distinct knickpoints or knickzones where channels narrow, steepen, and incise gorges or canyons.

The effect of vertical fault offset along a river profile depends on the magnitude and sense of offset. Sufficient upward displacement on the downstream side of a reverse fault can create a low-gradient reach upslope or even impound a lake, whereas upward displacement on the upstream side of a normal fault will create a knickpoint, or waterfall. The relationship between a river's course and the underlying bedrock structure provides clues as to whether the course of the river predates or postdates the topography (see Chapter 7). Lateral deformation of drainage patterns, via strike-slip fault offset, can produce asymmetry in drainage-basin form if sustained over geologic time.

Hillslopes

The steepness of hillslopes can respond to changes in tectonic forcing—up to a point. Soil-mantled slopes can only become so steep before landsliding begins to limit further steepening. In low- to moderate-gradient landscapes, slope steepness generally increases with increasing rock uplift rates, a characteristic of **sub-threshold slopes.** However, the angle of soil-mantled and well-fractured bedrock slopes can only increase to an upper limiting or threshold mean hillslope angle between 30 and 40 degrees, depending on soil and rock strength. Once hillslopes steepen to **threshold slopes,** erosion rates will respond to further increases in rock uplift through more frequent landsliding as hillslopes cannot steepen further to keep pace with ongoing river incision. The contrasting behavior of sub-threshold and threshold slopes (see Chapter 5) means that hillslope angles in low-gradient, sub-threshold terrain can

BOX 12.1 Drainage Area-Slope Analyses

A formal way to assess the adjustment of channel slopes, and therefore river profiles, to tectonic activity comes through positing a balance between rates of rock uplift (U_R) and river incision (E) to predict the form of steady-state river profiles. Generally, the erosive potential of a river may be expressed as a function of its drainage area (A) and its local slope (S), as a steeper river with a larger drainage area, and thus greater discharge, and will have greater power to cut down into rock:

$$E = K \cdot A^m \cdot S^n \qquad \text{eq. 12.A}$$

where the scaling exponents m and n are in many cases found to have values of 1.0 and 0.5, respectively, and K is a constant to characterize bedrock erodibility and the role of climatic factors and basin geometry in scaling the river's discharge. For the idealized case of a steady-state river profile eroding everywhere along its length at the rock uplift rate, $U_R = E$ and thus

$$U_R = K \cdot A^m \cdot S^n \qquad \text{eq. 12.B}$$

which may be rearranged to yield an expression for how channel slope would be expected to vary as a function of drainage area

$$S = (U_R/K)^{(1/n)} \cdot A^{-(m/n)} = K_s \cdot A^{-\theta_c} \qquad \text{eq. 12.C}$$

where $K_s = (U_R/K)^{(1/n)}$ and $\theta_c = m/n$. Equation 12.C predicts that the steady-state profile of an incising bedrock river will plot as a straight line on a logarithmic graph, with a slope equal to $-m/n$ and a coefficient (K_s) directly related to the ratio of the uplift rate to the bedrock erodibility, climate, and basin geometry (U_R/K). The coefficient K_s is termed the steepness index and the exponent θ_c is the profile concavity.

Subject to the assumptions of constant erodibility and rock uplift, area-slope plots can be used to identify locations along a bedrock river profile where uplift rates change; such changes will show up as breaks in slope on such plots **[Figure 12.B1]**. Portions of an upland river system with higher rock-uplift rate (U_R) will plot at higher slopes for the same drainage area than will reaches with lower rock-uplift rates. Values of K_s determined for different reaches of a river system can be used to estimate relative differences in the value of rock-uplift rates in different portions of a river system, if, of course, one has accounted for any differences in lithology, discharge, and bedrock erodibility, which may influence the denominator of the ratio that makes up K_s. Interpreting drainage area-slope analyses requires consideration of potential lithologic or tectonic factors that may influence the analysis, such as the presence of knickpoints propagating up through a channel network or large differences in bedrock erodibility.

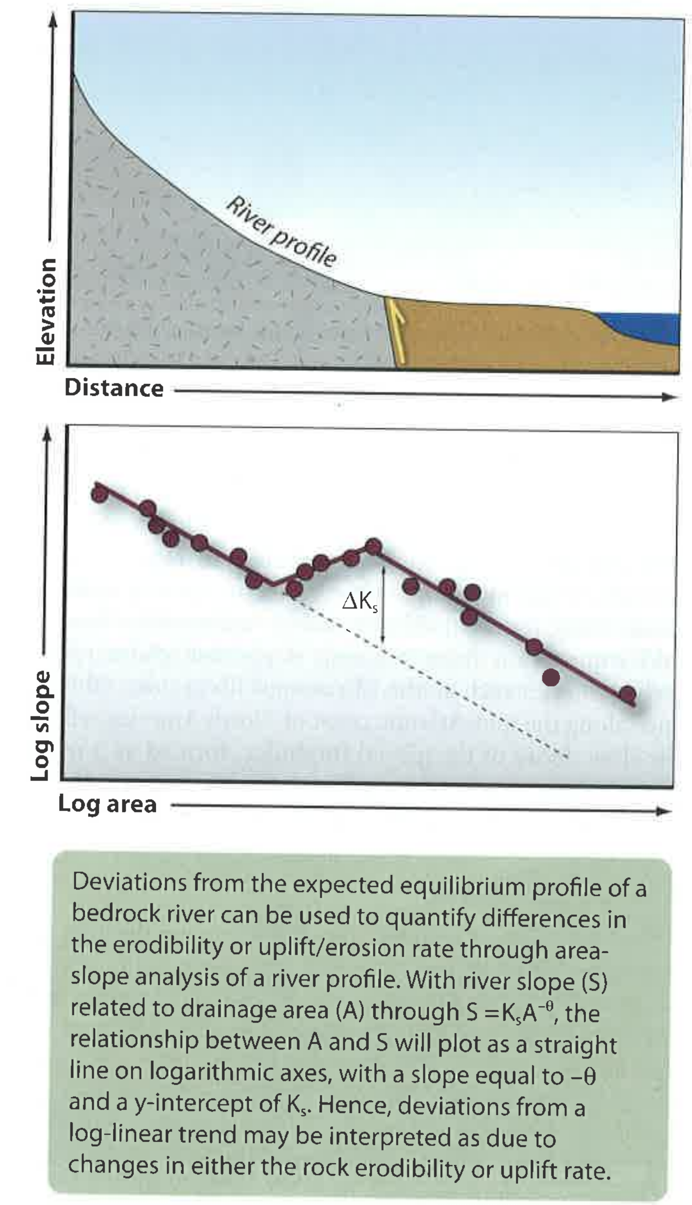

FIGURE 12.B1 Drainage area and slope are inversely related down river systems. Locations where the slope/area trend along a river diverges from the general trend can be used to identify locations where rock uplift rate or erodibility may have changed.

respond to an increased uplift rate or base-level fall through steepening. In contrast, the morphological response of landscapes with steep, threshold slopes would be most pronounced in stream profiles or in the stripping of soil to expose bare bedrock slopes, and the proliferation of deep-seated bedrock landslides.

The development of **inner gorges,** zones of steeper slopes low on valley walls that create a distinct valley-within-a-valley morphology **[Photograph 12.17]**. Inner gorges represent a response to either falling base level or an increase in uplift rate. Gorge walls steepen where rivers incise into bedrock faster than hillslopes erode. Where river incision

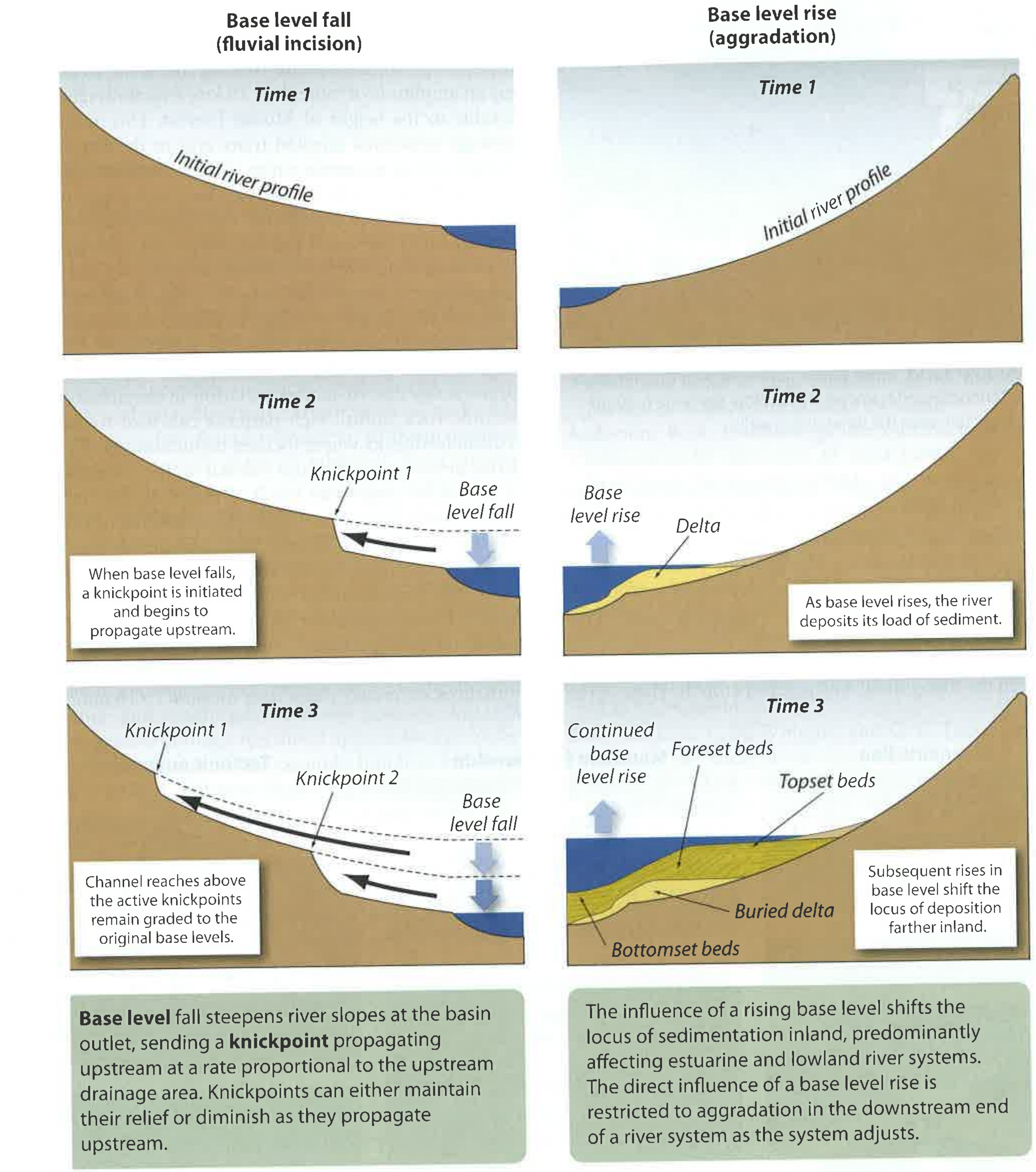

FIGURE 12.14 Fluvial Responses to Base Level Change. The rise and fall of base level result in very different river, estuarine, and sedimentary basin responses.

rates increase in response to increased tectonic uplift, the lower portions of valley walls steepen first, creating the distinct break in slope that defines an inner gorge.

The sediment supply of rivers can be affected by earthquakes or large storms that deliver large loads of hillslope-derived sediment to river systems in the form of seismically induced or precipitation-induced landslides. The resulting increase in downstream sediment loading may cause channel aggradation and braiding, and temporarily convert bedrock channel reaches into alluvial reaches.

Erosional Feedbacks

Coupling between erosion and tectonics produces a range of effects from continental-scale topography down to the

PHOTOGRAPH 12.17 Inner Gorge. Inner gorge in the Himalaya between Tenboche and Namche, Nepal. Note the large (~10 m) boulder stuck between the inner gorge walls.

scale of geological structures along individual river valleys [**Figure 12.15**]. For instance, erosion of deep valleys can focus exhumation, resulting in preferential rock uplift (through isostatic rebound) along major rivers. This can produce a **river anticline** defined by the crest of a structural fold running parallel to the river, with the greatest uplift centered along the course of the river.

This odd situation of a river flowing along the structural high, with the topographic low perched atop the spine of the structural anticline, characterizes major trans-Himalayan rivers, like the Arun River just east of Mount Everest. The anticlinal geologic structure running along the Arun River has an amplitude of more than 10 km, structural relief comparable to the height of Mount Everest. This and similar geologic structures oriented transverse to the trend of the compressional mountain range are the youngest deformational structures in the Himalaya. Their young ages indicate that they developed in response to incision along the rivers and thus that the rivers did not simply take advantage of preexisting structure to establish their courses across the rising range. In other words, the focused erosion along the course of major, range-crossing rivers unroofs rocks at speeds greater than in areas away from the river courses.

Sustained gradients in erosion also occur at much larger scales due to spatial variation in precipitation and tectonic rock uplift. Such patterns can lead to tectonic-erosion feedbacks where focused denudation is coupled to active deformation. Greater rainfall on the windward side of mountain ranges can result in differential exhumation that strongly influences structural development on either side of the drainage divide. This asymmetry deeply exhumes rocks on the wetter, windward side of ranges, such as on the rain-drenched western slope of the southern Alps of New Zealand. In the dramatic cases of where the Indus and Tsangpo rivers—the most powerful and erosive in Asia—slice through deep gorges at either end of the Himalaya, extremely rapid river incision (>10 mm/yr) has

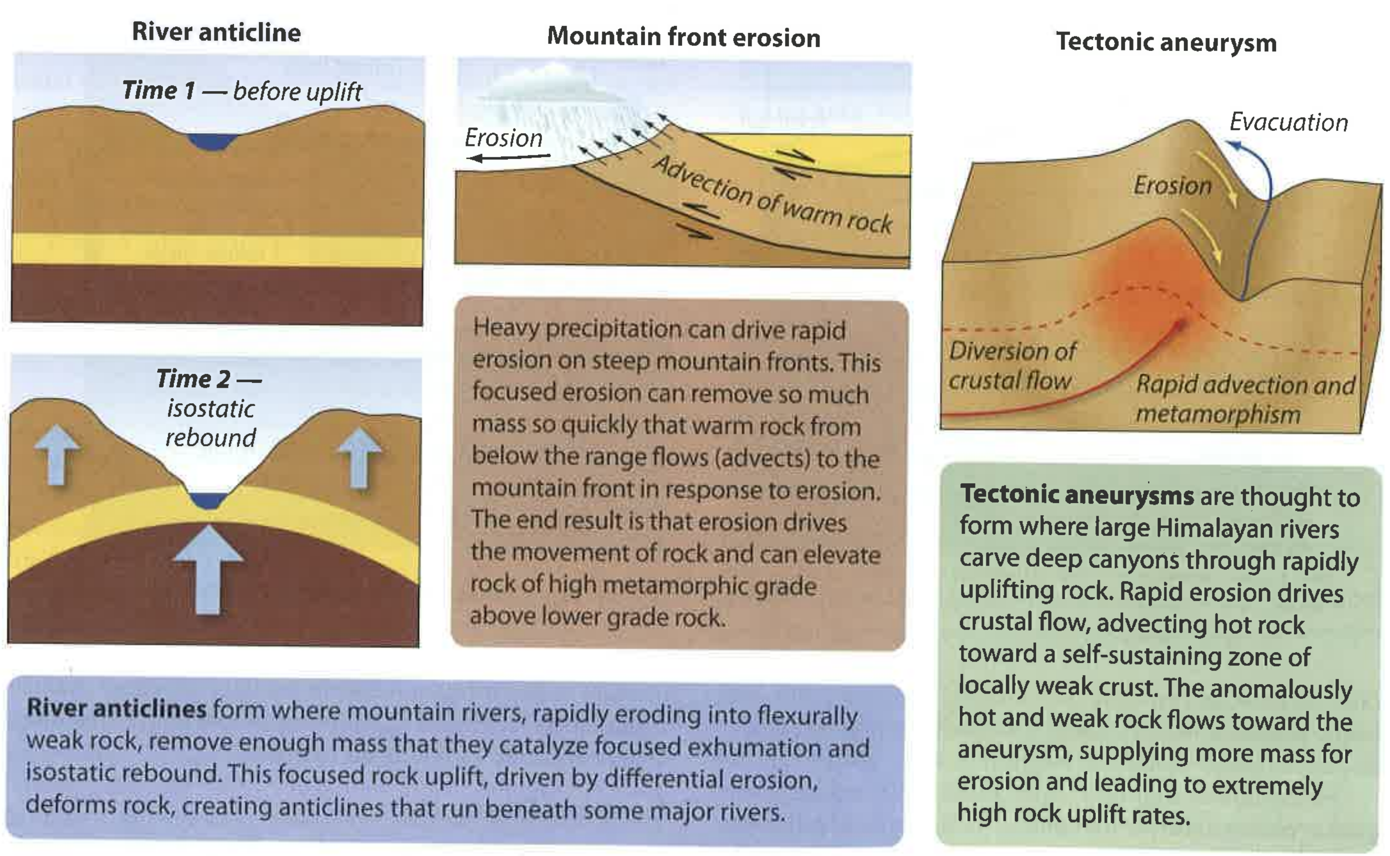

FIGURE 12.15 Tectonics and Geomorphic Feedbacks. Erosion influences solid-Earth processes by removing mass from the surface, thereby inducing crustal flow, deformation, and uplift. [Adapted in part from Zeitler et al. (2001).]

resulted in the development of deeply exhumed geologic structures expressed as a bull's-eye pattern of young, high-grade metamorphic rocks in the region surrounding the rapidly eroding gorges.

Applications

Understanding the role of tectonics in setting the rate and type of geomorphic processes and the distribution of landforms is key to understanding the geomorphology of many regions around the world. Explicitly considering and understanding the linkage between solid-Earth and surface processes has important applications to landscape management, sustainability, and natural-hazard reduction. The tools and approaches of tectonic geomorphology are central to understanding why certain types of rocks and landforms are often found together in different parts of the world. Structural patterns inherited from ancient tectonic deformation can greatly influence modern landforms and surface processes through sustained variability in erosion resistance. In rapidly uplifting areas of the world, such as the Himalaya, tectonic uplift and the limited strength of rocks keeps slopes poised to fail at threshold steepness. Frequent landslides make road maintenance in tectonically active mountain belts a challenge.

Understanding the dynamic nature of Earth's surface, as driven by tectonics, is the foundation for understanding the nature and magnitude of many geologic hazards. Tectonic geomorphology has direct applications to understanding seismic hazards. For example, building and development is limited, and more highly regulated, along active fault zones in California than it is elsewhere in the state. Mapping the distribution and alignment of fault-related landforms is a standard technique employed in seismic-hazard assessments. The physiographic expression of faults and fault zones can be used to map the traces of faults, evaluate seismic hazards, and identify locations at risk of direct fault offset. Tectonic geomorphology is useful for assessing earthquake recurrence intervals for faults along active mountain fronts in both densely populated areas, such as along California's San Andreas Fault and in relatively inaccessible regions, such as the High Himalaya. In particular, identifying the character, extent, and history of motion across fault zones allows geologists to estimate the size and characteristics of past and thereby future earthquakes. Geomorphological techniques are commonly used in site investigations for seismically vulnerable infrastructure, such as nuclear power plants and dams.

The broad patterns of rock uplift, governed by tectonic processes, provide the raw material and template upon which erosion works to sculpt landscapes. In this manner, tectonic setting influences both the tempo and spatial patterns of landscape-forming processes, specifically, the rate at which rock weathers and soil erodes. Tectonics controls differential rates of erosion through slope steepness and the extent and rate of soil formation. Thus, regions where rapid uplift outpaces weathering have thin soils that erode rapidly. In contrast, wet, tectonically inactive regions, where weathering outpaces exhumation, develop deeply weathered, thoroughly leached, nutrient-poor soils and thick zones of saprolite. The resulting differences in soil type and fertility greatly affect both ecosystem characteristics and dynamics and the ability of a landscape to sustain intensive land uses, such as farming. The terracing of mountainsides, a practice common to many different cultures around the world, reflects a human adaptation to accelerated erosion of the soil needed to farm on mountainsides steepened by tectonics.

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DIGGING DEEPER When and Where Did that Fault Last Move?

Understanding when, where, and how much a fault last moved is important both to scientists and society. Earthquakes associated with active fault offset have damaged infrastructure in the past and present an ongoing hazard. Characterizing the location, timing, and amount of fault movement is critical for determining the rate of solid-Earth deformation, an important geomorphic control on landscape evolution.

In lightly vegetated and undeveloped terrain, the surface expression of faults—their scarps—can be mapped in the field and by using aerial photographs. Fault scarps are one of the most geomorphically distinct surface expressions of plate tectonics and a direct way to document deformation and infer seismic shaking in the past. However, in forested terrain or in urban areas, fault traces were difficult to map until the advent of LiDAR. Now, high resolution LiDAR-based digital elevation models (DEMs) are being used to map scarps beneath dense vegetation or in areas that have been heavily disturbed by development. For example, in Washington State, LiDAR clearly revealed the trace of the Tacoma Fault Zone (Sherrod et al., 2004; **Figure DD12.1**), which carbon-14

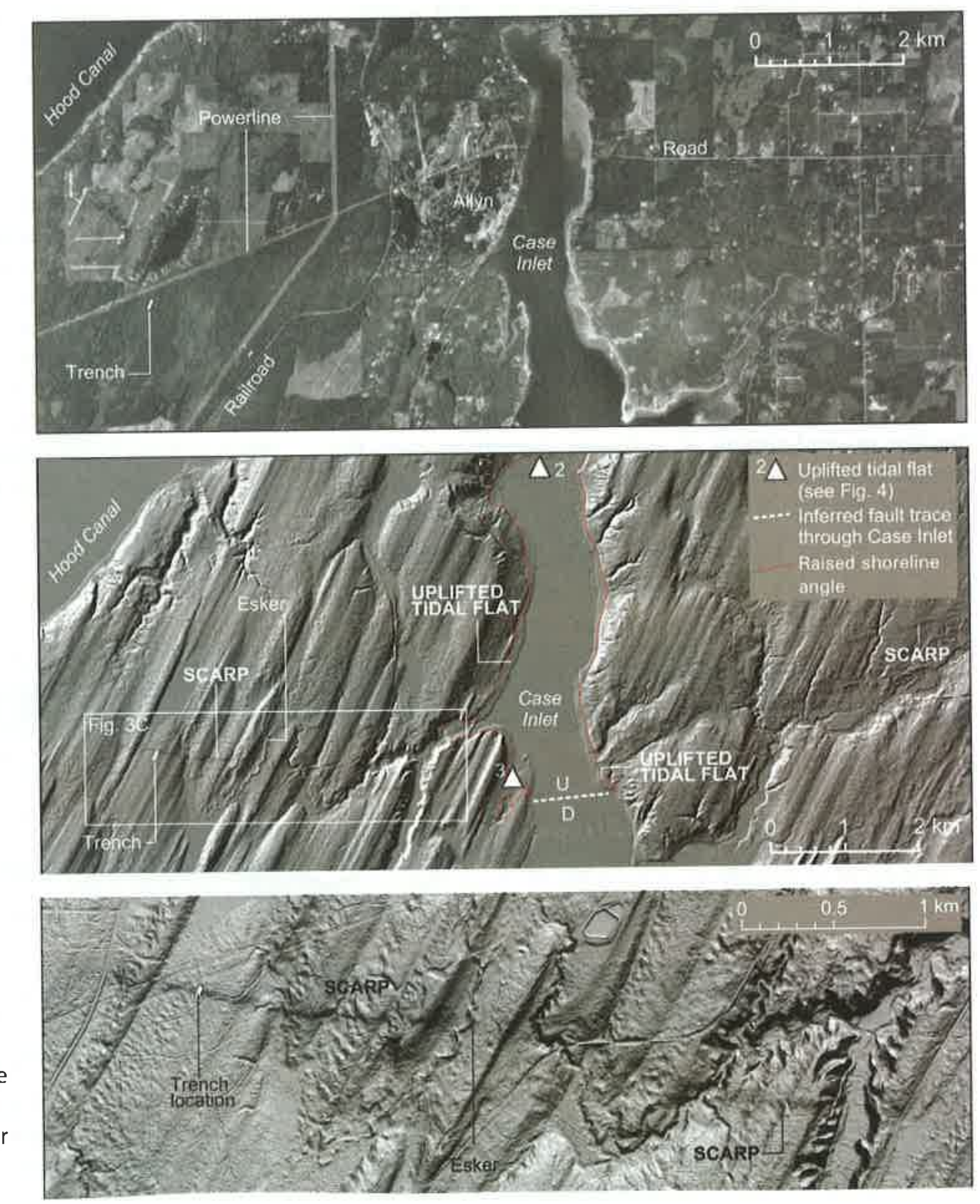

FIGURE DD12.1 (a) Look closely at this orthophotograph of the landscape near Tacoma, Washington; it is unlikely you can find the fault scarp. (b) In this LiDAR DEM, the fault scarp is visible (and labeled). The elongated ridges, trending NE/SW, are glacial drumlins. (c) The LiDAR image is greatly enlarged, showing the linear fault scarp even more clearly. [From Sherrod et al. (2004).]

DIGGING DEEPER When and Where Did that Fault Last Move? *(continued)*

dating of associated organic material indicates has been active in the past 1000 years.

Geomorphologists can estimate the age of fault scarps by studying how they erode and change shape predictably over time. Scarps evolve from steep and angular to gentler and more rounded [**Figure DD12.2**]. In scarps formed in unconsolidated materials such as sand, gravel, and glacial till, material moves from the steep fault scarp face and from the sharp upper inflection to form a depositional wedge of colluvium at the base of the scarp. Most of this mass appears to move by diffusive processes similar to creep (see Chapter 5); thus, the rate of mass movement off the scarp face is proportional to the gradient or slope of the scarp. Researchers have calibrated the change in fault scarp shape over time by determining or assuming a diffusion coefficient so that scarps of unknown age can be dated (Hanks et al., 1984). Diffusion dating of fault scarps is complicated by the observation that diffusion coefficients differ as a function of scarp height and orientation (Pierce and Colman, 1986).

Clues to the timing and magnitude of past earthquakes and the offset they caused are preserved both on the fault scarp itself and in the adjacent colluvial wedge (material shed off the scarp and deposited at its base). Trenching colluvial wedges provides an indication of the number of events and their displacement as each event will often bury or displace identifiable soil horizons. Radiocarbon analysis of buried organic material, as well as luminescence dating techniques when organic matter is absent, are used along with detailed trench-wall mapping to infer the history and timing of sediment deposition onto the wedge. Since most wedge sedimentation occurs during and just after each earthquake steepens the fault scarp, understanding the history of wedge sedimentation and soil development allows geomorphologists to infer faulting history [**Figure DD12.3**].

In the absence of a colluvial wedge, or if the wedge contains no datable material, then geomorphologists measure the offset of distinct landforms. Offset moraines, pluvial lake shorelines, and river terraces are common datums (Wallace, 1977) that can be dated directly, or their age can be assumed based on climatic inferences [**Figure DD12.4**]. In some cases, offset historical structures, such as a Crusader castle wall in Israel (Ellenblum et al., 1998), provide chronologic control and act as strain gauges.

Bedrock scarps, because they generally lack colluvial wedges, remained a dating challenge until the development of cosmogenic nuclide techniques. In 2001, Gran et al. made ^{36}Cl isotopic measurements in samples collected at regular intervals down the limestone exposed on a 9-m-high normal fault scarp in northern Israel. The scarp bisects a village, and homes are built at the top of the upthrown block to take in the view. Because cosmic rays penetrate below the ground surface and because a variety of different combinations of displacement and earthquake timing could generate similar isotope concentration profiles down the scarp, Gran et al. (2001) fit many different models to their data in a process termed optimization modeling [**Figure DD12.5**]. This allowed them to conclude that the fault had moved several times during the Holocene, most recently in the past few thousand years. Motion was greatest in the mid-Holocene, a time when archaeologists identified fatalities in a nearby cave, the collapse of which was likely triggered by shaking on the fault they dated.

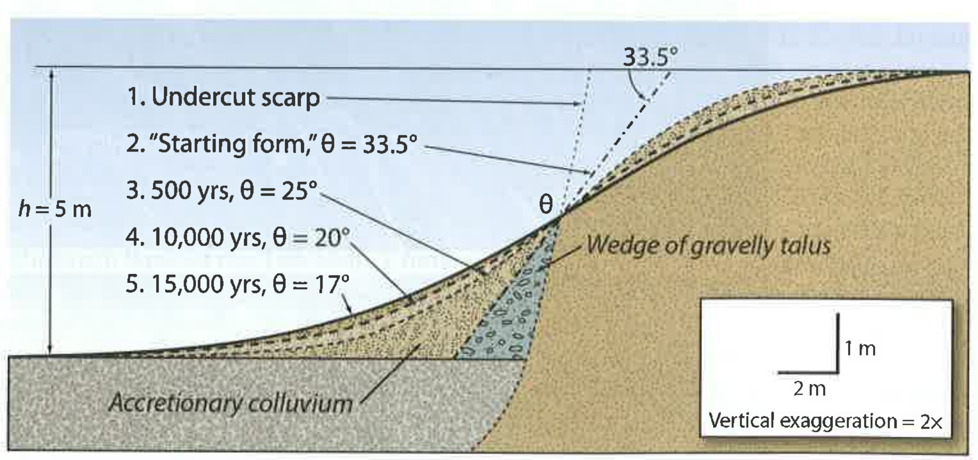

FIGURE DD12.2 Schematic diagram of a one-event scarp in unconsolidated material and its adjacent colluvial wedge (labeled "accretionary colluvium"). Scarp shapes, as predicted by the diffusion equation over time, are shown along with average slopes (in degrees). The assumed "starting form" of a 33.5-degree slope is the estimated angle of repose for colluvium. Over time, the scarp gets less steep and the wedge gets larger and thicker. [From Pierce and Coleman (1986).]

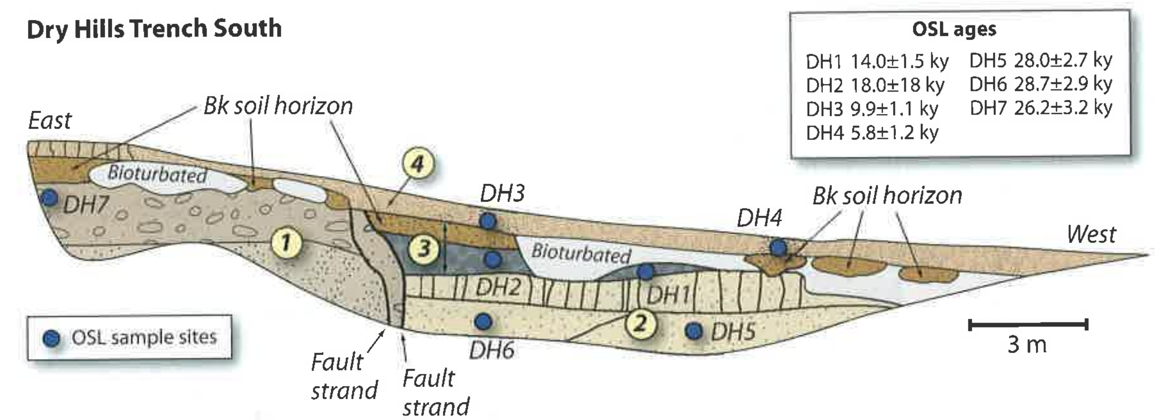

FIGURE DD12.3 Trench log depicting the colluvial wedge formed by erosion of a normal fault in northern Great Basin of western North America. The upthrown block is on the left; the downthrown block on the right. The two strands of the fault are the solid black, nearly vertical lines. The trench reveals three faulting events dated by optically stimulated luminescence (OSL). Unit 1 is not exposed in the hanging wall (graben) but only on the upthrown block; it and unit 2 are the same age. The earliest event caused the deposition of unit 2 about 28,000 years ago. Note that unit 2 is capped by a soil as indicated by the short vertical lines. The next event shed material off the scarp, forming what is now unit 3 (14,000–18,000 years ago). The latest event shed unit 4 between 5,000 and 10,000 years ago. [From Wesnousky et al. (2005).]

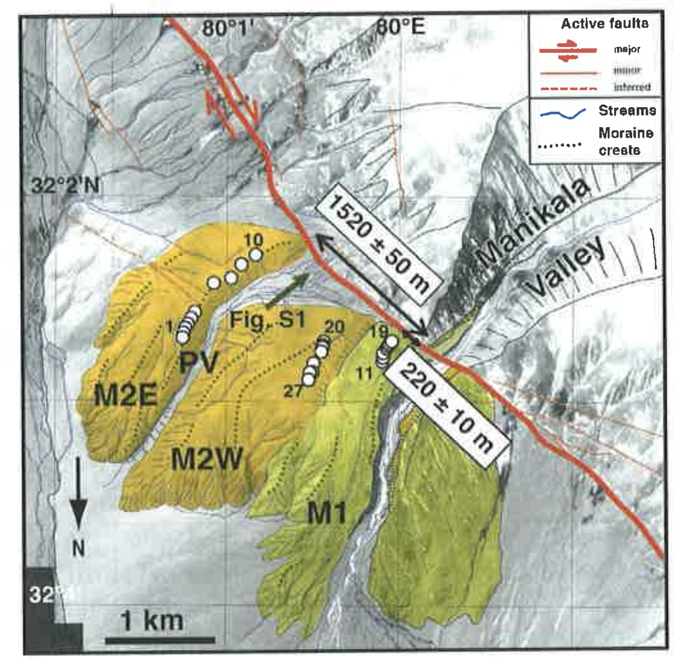

FIGURE DD12.4 In Tibet, the Karakorum Fault is a major, 1200 km-long feature north of the Himalaya. The fault laterally and vertically offsets many moraines deposited during cooler or wetter times. Here, ice flowing down the Manikala Valley formed two sets of moraines at different times. The fault has since offset the moraine complexes laterally from the source valley. Cosmogenic exposure ages (sample sites shown by white circles) show the earlier moraines, labeled M2, are ~140,000 years old and the younger moraines (M1) are ~20,000 years old. Using the measured offsets of 1520 ± 50 m (M2) and 220 ± 10 m (M1) for the two sets of moraines and their ages, Chevalier et al. (2005) suggest a long-term slip rate of about 1 cm/yr. This is many times larger than the deformation rate measured by InSAR (see Chapter 2) over a recent 8-year period, implying that slip on the fault varies over time. [From Chevalier et al. (2005).]

Chevalier, M. L., F. J. Ryerson, P. Tapponnier, et al. Slip-rate measurements on the Karakorum Fault may imply secular variations in fault motion. *Science* 307 (2005): 411–414.

Ellenblum, R., S. Marco, A. Agnon, et al. Crusader castle torn apart by earthquake at dawn, 20 May 1202. *Geology* 26 (1998): 303–306.

Gran, S., A. S. Matmon, P. R. Bierman, et al. Displacement history of a limestone normal fault scarp northern Israel from cosmogenic ^{36}Cl. *Journal of Geophysical Research* 106 (2001): 4247–4265.

Hanks, T., R. C. Buckham, K. R. Lajoie, and R. E. Wallace. Modification of wave-cut and faulting controlled landforms. *Journal of Geophysical Research* 89 (1984): 5771–5790.

Pierce, K. L., and S. M. Colman. Effect of height and orientation (microclimate) on geomorphic degradation rates and processes, late-glacial terrace scarps in central Idaho. *Geological Society of America Bulletin* 97 (1986): 869–885.

Sherrod, B. L., T. M. Brocher, C. S. Weaver, et al. Holocene fault scarps near Tacoma, Washington, USA. *Geology* 32 (2004): 9–12.

Wallace, R. E. Profiles and ages of young fault scarps, north-central Nevada. *Geological Society of America Bulletin* 88 (1977): 1267–1281.

Wesnousky, S. G., A. D. Barron, R. W. Briggs, et al. Paleoseismic transect across the northern Great Basin. *Journal of Geophysical Research* 110 (2005): B05408 doi: 10.1029/2004JB003283.

DIGGING DEEPER When and Where Did that Fault Last Move? *(continued)*

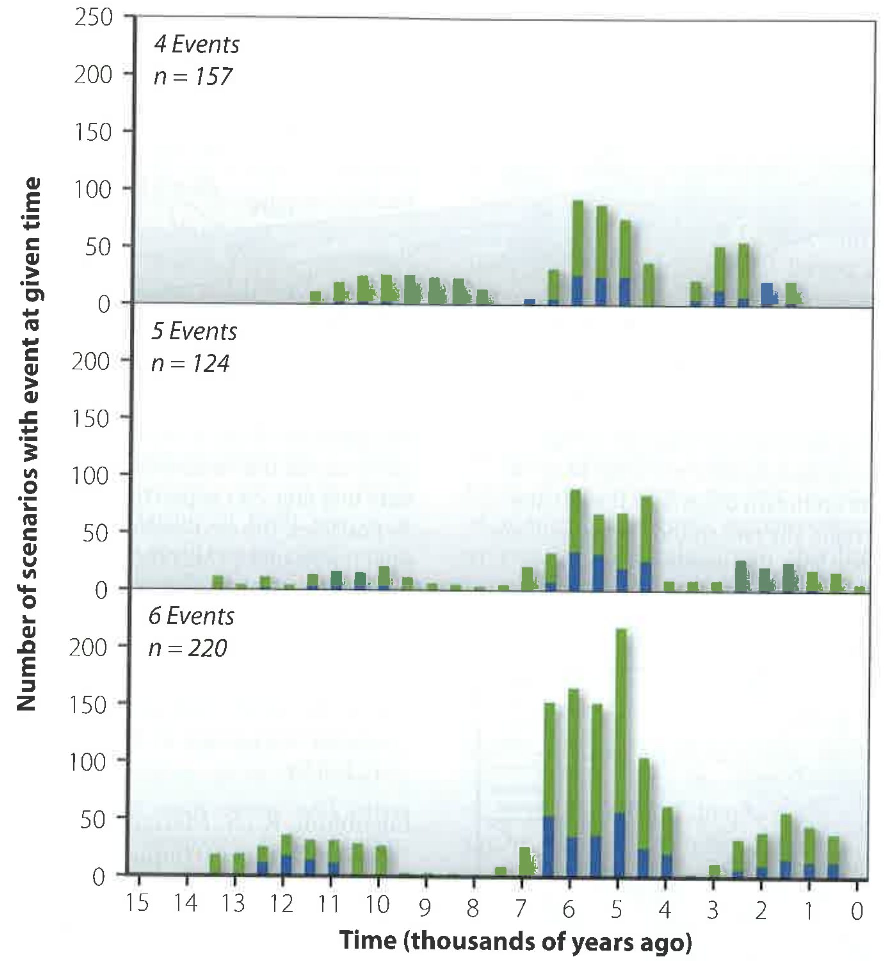

FIGURE DD12.5 Measuring the ^{36}Cl distribution on a fault scarp does not yield a unique age estimate for prior motion; rather, many different models (timing, size, and number of fault offsets) can be fit to the data. By running multiple models and looking for consistency between results, one can have higher confidence in the estimated timing of paleoearthquakes. These histograms are a summary of many different models; the best fit models are shown in blue. The shapes and modes of the three histograms are similar, indicating that no matter if one assumes that 4, 5, or 6 events created the present-day fault scarp, the mid-Holocene (4000 to 7000 years ago) was the most active time for earthquakes. [From Gran et al. (2001).]

WORKED PROBLEM

Question: The average elevation of the High Himalaya is about 5 km, with peaks rising substantially higher. The average long-term erosion rate in the Himalaya is about 1–2 mm/yr and the average erosion rate of inactive orogens, such as the Appalachian Mountains, is only about 0.02 mm/yr. How long would it take to erode the world's highest mountain range down to sea level once tectonic rock uplift ceased, and how much rock would be eroded off in the process?

Answer: A simple estimate obtained by dividing the height of the range (5 km) by the average erosion rate (1–2 mm/yr) yields just 2.5 to 5 million years. However, this estimate is far too short due to neglect of isostatic rebound. Recall that surface uplift equals the uplift of rock less the amount of rock eroded (eq. 12.1). The effect of isostatic compensation on net surface elevation may be estimated using equation 12.3: $U_S = E\,[(\rho_c/\rho m) - 1]$. For the case of $\rho_c \approx 2.7$ g/cm^3 and $\rho_m \approx 3.3$ g/cm^3, the ratio $\rho_c/\rho_m = 0.82$ indicates that once tectonically driven rock uplift ceases, erosionally driven rock uplift will compensate for 82 percent of the mass removed from the surface by erosion. Hence, equation 12.3 may be recast as $U_s = -0.18$ E. Rearranging and solving for E, yields $E = -5.6\ U_s$, indicating that 5.6 km of rock needs to be eroded in order to result in 1 km of net surface lowering. Hence, 28 km of rock needs to be eroded to reduce a 5 km high mountain range to sea level (5 km · 5.6). At a pace of 1–2 mm/yr, this would take 14–28 million years. But this itself represents an underestimate of the total time involved because the erosion rate of the range would decline to well below 1 mm/yr as the topographic relief of the range diminished and hillslopes relaxed to gentler gradients once tectonic forcing ceased. For example, the current erosion rate of the southern Appalachian Mountain range is about 0.02 mm/yr. At this pace it would take almost 420 million years to reduce the 1500 m relief of the southern Appalachians to sea level [1500 m/(0.18 · 0.00002 m/yr)].

KNOWLEDGE ASSESSMENT Chapter 12

1. Explain why many of our planet's major surface features correspond to current or ancient boundaries between tectonic plates.
2. Explain the difference between active and passive tectonic control on landscapes.
3. In what tectonic settings do lithology and structure exert strong control on geomorphology? Explain why this is the case.
4. How do the drivers of rock uplift change over time?
5. Explain the role of erosion in driving both rock and surface uplift.
6. Explain how changes in base level can be inferred from the landscape.
7. Use a sketch to explain and define the different types of uplift.
8. List and explain several examples of how density contrasts drive geomorphic change.
9. Provide several examples illustrating the geomorphic expression of active faulting.
10. What is an anticlinal valley and how does it form?
11. Why is it difficult to read underlying structure from landscape features in areas like the Himalaya and Taiwan?
12. What are the geomorphic expressions of joints and in what climate zones are such expressions easiest to detect?
13. Give two examples of different tectonically-controlled drainage patterns and explain what controls their geometry.
14. When and where does coastal subsidence happen?
15. Make a sketch showing the three most important geomorphic features of an uplifting coast.
16. How can fluvial terraces be used to estimate uplift rates?
17. How do hillslopes respond to increasing uplift rates?
18. What is an inner gorge and what might inner gorges tell us about uplift rates?
19. What can be learned from the longitudinal profile of a river?
20. Explain how the stream gradient index works and why it is useful for studying the effect of tectonics on landscapes.
21. Sketch an area/slope plot and explain why it is geomorphically useful.
22. Provide two examples of how tectonics creates knickpoints and knickzones.
23. What is a river anticline and how does it form?
24. Compare and contrast the large-scale geomorphology (characteristic landscape features) of compressional orogens, extensional rift zones, and continental interiors.
25. Explain sedimentary wedges, noting where they are found.
26. Explain how rift zones control the orientation and character of drainage networks.
27. Explain colluvial wedges, noting where they are found.