GEOLOGY 101 FINAL EXAM STUDY GUIDE - SECOND HALF OF COURSE

Study Material extracted from Don Barrie's Website: **http://classroom.sdmesa.edu/dbarrie/Geol100ChapStudyGuides.htm**

Chapters 10, 13 and 14 - Mountain Building and Geologic Structures

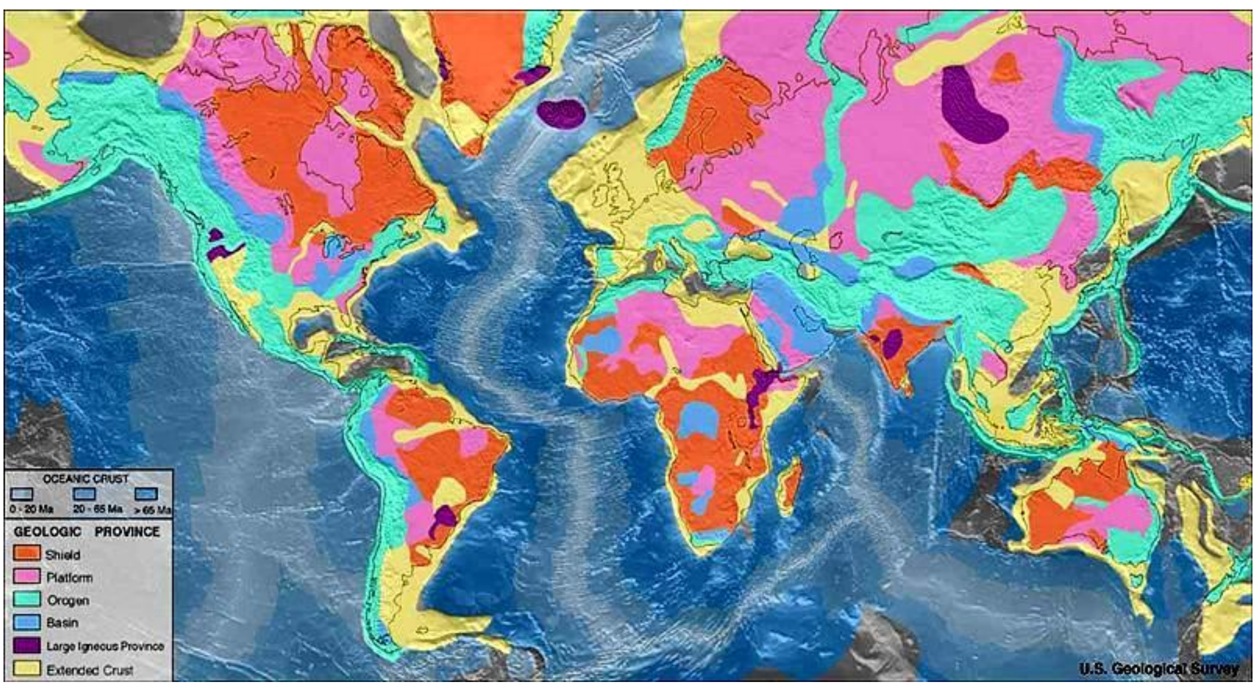
Mountain Belts and Continental Crust

At this point in the course, you should appreciate that mountain building generally occurs at convergent plate boundaries (subduction and collision zones), where compressional stress predominates, and also at divergent plate boundaries (seafloor spreading and continental rifting), where tensional stress predominates.

As with most topics in geology; however, the details can get somewhat complicated. In this chapter, we'll focus on various processes that occur in conjunction with mountain building.

Your text describes a mountain belt as a chain thousands of kilometers long composed of numerous individual mountain ranges. Mountain belts are characterized by diverse rocks that show evidence of intense deformation (folding, faulting, and metamorphism), vertical uplift, and often a history of igneous activity.

Figure in your text shows Earth's major mountain belts. Here's a map (below) showing the world's mountain belts in more detail. The light blue areas represent active mountain belts (orogens):



Mountain Belts (Orogens) of the World (shown in light blue).
(image courtesy of Wikipedia; http://en.wikipedia.org/wiki/Orogeny)

You might notice that the Earth's major mountain belts generally correspond with convergent plate boundaries, both active an inactive. This is just what we'd expect, since plate convergence is what causes most mountain building.

Isostasy

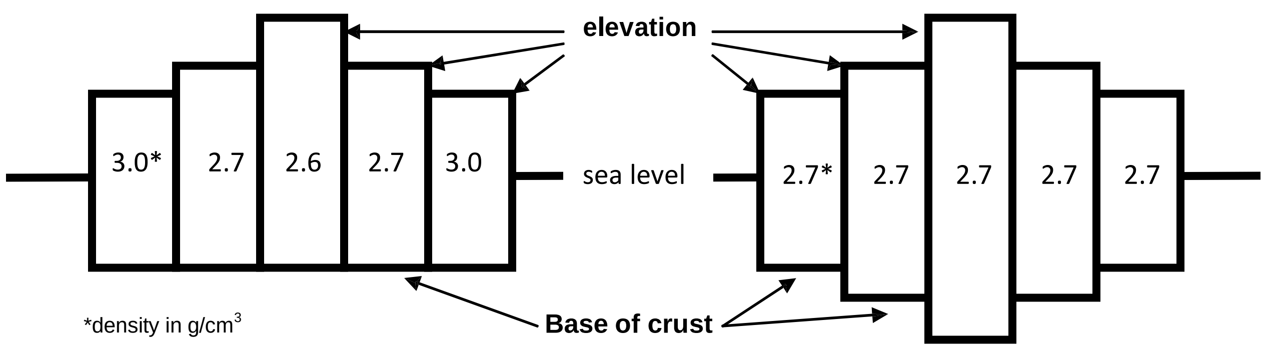
So far, we've mainly focused on horizontal motions of the Earth's lithosphere, and we've seen that such motions can be understood in terms of plate tectonics. In order to understand how mountains are built, we also need to introduce an important process that accounts for much of the vertical uplift that characterizes mountain belts: isostasy.

As defined in the glossary of your text, isostasy refers to the balance or equilibrium between adjacent blocks of crust resting on a plastic mantle.

I prefer to think of isostasy as the buoyant equilibrium that exists between the lithosphere and asthenosphere, such that lithospheric plates float within asthenospheric mantle at an elevation that depends on both their thickness and density.

Isostasy is such an important concept that it's appropriate to delve into it in more detail. In the late 1800s, two different models of isostasy were developed, the so-called Pratt and Airy models.

These two models are illustrated below:



Pratt Isostasy Model. **Airy Isostasy Model.**

In both the Pratt and Airy isostasy models (named after the folks who developed them), the elevation of the Earth's surface is explained in terms of crustal blocks that float atop a plastic mantle.

In the Pratt model, different surface elevations are understood in terms of crustal blocks of differing density. In this case, the blocks with lower densities float higher than those with higher densities, in the same way that a block of styrofoam floats higher in a swimming pool than a block of wood. In this model, the bottom of each block occurs at the same depth.

In the Airy model, different surface elevations are understood to be the result of the different thicknesses of crustal blocks, with thicker crustal blocks sticking up higher and projecting downward more deeply than thinner ones. In this model, the depth to the bottom of each block differs from block to block.

So…which model is correct? Both models, it turns out, are accurate in some respects but not in others.

For example, the Pratt model explains in a very general way why the continents stick up higher above sea level than the ocean basins. Since continental crust is less dense than oceanic crust, we'd expect it to ride higher atop the mantle than oceanic crust. This is just what the Pratt model of isostasy predicts. In other words, the continents are high and dry not just because they're above sea level. Rather, the continents are elevated because they're composed of rock types (e.g., granitic and metamorphic rock) that are more buoyant (less dense) than oceanic crust, which is mainly basaltic.

But wait…in the Pratt model, notice that the base of the crust is the same everywhere. Seismic investigations later revealed that this just isn't the case. In general, the base of the Earth's crust beneath the continents (i.e., the Moho) occurs at a much greater depth than beneath the ocean basins.

In the Airy model of isostasy, the depth to the crust-mantle boundary (Moho) varies, however, this model assumes that the densities of crustal rocks are everywhere about the same, which just isn't the case. The Airy model predicts that regions of thickened continental crust should be associated with higher elevations, which generally is the case. The really important prediction of the Airy model is that Earth's mountain belts should possess crustal roots (i.e., thicker crust that projects downward into the mantle like the root of a tooth). Seismic data confirm that the world's major mountain belts such as the Himalaya are indeed underlain by thicker crust.

By the way, an easy way to remember the basic difference between the Pratt and Airy models is the phrase, "Pratt is flat but Airy varies."

Currently, the preferred model of isostasy applies to the lithosphere and asthenosphere and combines aspects of both Pratt and Airy. In essence, we now recognize that continental lithosphere rides higher atop the mantle than oceanic lithosphere because it's both thicker and less dense (Pratt); also, the depth to the base of the crust and lithosphere varies from place to place, and many of Earth's mountain belts are underlain by lithospheric roots (Airy).

An historical note:

You might remember from the chapter on plate tectonics that Alfred Wegener's theory of continental drift wasn't widely accepted by his contemporaries, particularly geologists in North America. One reason for this may have been that North American geologists in the 1920s and 1930s were reluctant to abandon their oversimplified understanding of isostasy.

As U.C. San Diego scholar Naomi Oreskes discusses in her book, The Rejection of Continental Drift: Theory and Method in American Earth Science (Oreskes 1999), *the Pratt model of isostasy was generally preferred by geologists in North America over the Airy model because it was easier to work with. As mentioned above, a prediction of the Pratt model is that the depth to the base of Earth's crust should everywhere be the same; however, Wegener argued that the leading edges of the continents were thickened as they drifted across the globe. Wegener's critics in turn argued (quite correctly) that thickened continental edges would imply a variable depth to the base of the crust—a conclusion clearly at odds with Pratt isostasy. The result is that Wegener's critics rejected his ideas in part because they were incompatible with Pratt isostasy. In effect, they rejected Wegner's ideas rather than rethink their commitment to Pratt isostasy.*

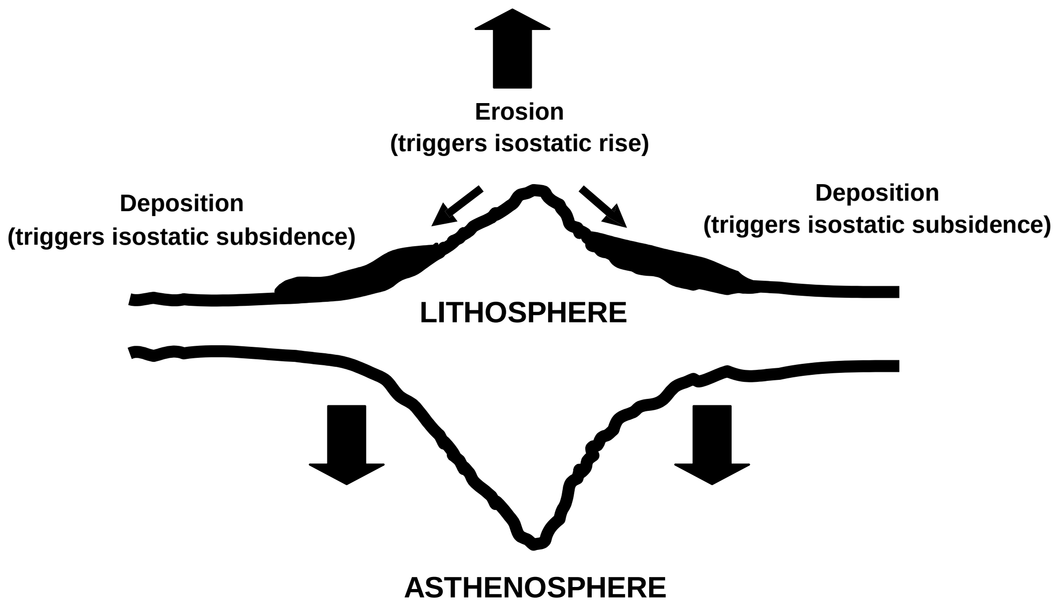
In the end, Wegener's prediction that drifting continents become thickened at their leading edges has been upheld (although for different reasons that Wegener suggested), whereas the Pratt model of isostasy has been falsified. Score one for our hero, Alfred Wegener ☺.

Large vertical motions of the lithosphere occur when the lithosphere is out of isostatic balance with the underlying asthenosphere.

In general, mountain belts develop in response to lithospheric thickening. Such thickened lithosphere creates an isostatic imbalance. In response, the lithosphere within a mountain belt typically rises to counteract this imbalance (Figure below).

The lithosphere can also move vertically downward if loaded from above. For example, when thousands of meters of ice accumulate on a continent during an ice age, the lithosphere of the continent sinks in response. When the ice age ends and the glacial ice melts, the lithosphere rises in response. Here's a simple animation of **Isostasy— Figure 5.14.**

Although it may seem counterintuitive, erosion also leads to isostatic rise. As a mountain belt erodes, the lithosphere rises in response due to unloading, although usually not up to its former elevation, as shown here:



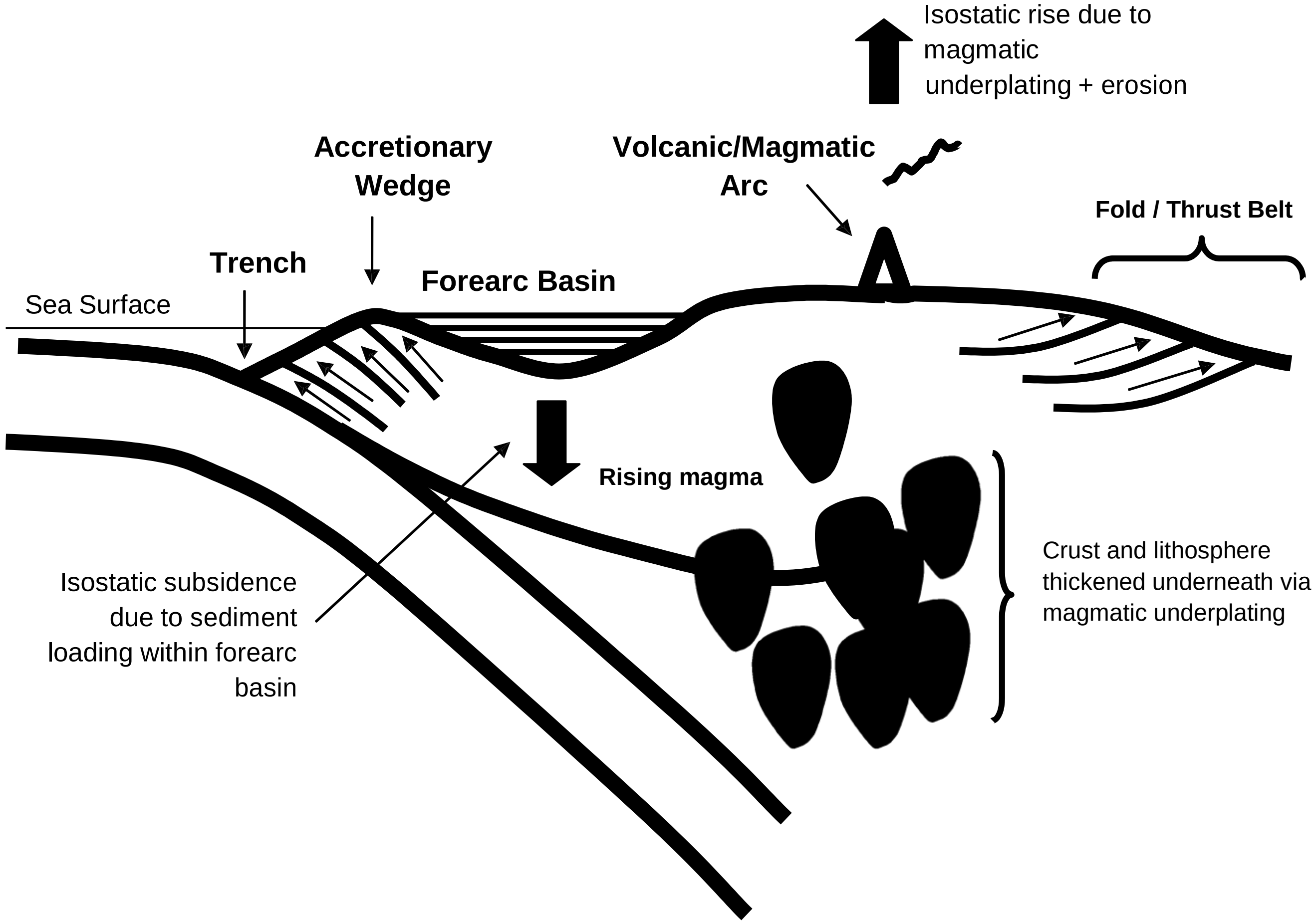
Erosion, Deposition, and Isostasy.

Let's consider some specific plate tectonic settings where mountain belts develop:

Ocean-Continent Convergence (plate subduction)—marginal, "Andean-type" mountain belts

Figure below shows a geologic cross section through a typical "Andean-type" mountain

belt. Here's a simplified sketch (next page):



Simplified cross section through an Andean-type mountain belt.

Next to the trench, we find a complexly folded and faulted region called the accretionary wedge, where seafloor rocks from the subducting plate are being scrapped off and plastered onto the non-subducting plate.

Thrust and reverse faulting along with intense shearing characterize accretionary wedge rocks. As you might imagine, the rocks within an accretionary wedge are highly deformed and mixed up—they've been through a lot! In fact, a rock called "mélange" is commonly found in accretionary wedges. Mélange consists of a complex mixture of various rock fragments jumbled together in a matrix of sheared, fine-grained mudrock.

Here's a **photo of mélange** from an ancient subduction zone in the Appalachian Mountains. Notice the large rock fragments surrounded by fine-grained mudrock. Notice, too, how sheared and broken up this stuff is. Think of it as tectonic gobbledygook!

Here's another photo of some mélange from the Franciscan subduction complex, near San Francisco, CA:

Melange with Chert Clasts, Marin Headlands, CA.
Image courtesy of Bruce Molina, United States Geological Survey
Image source: Earth Science World Image Bank
(http://www.earthscienceworld.org/images)

Moving landward from the accretionary wedge, we encounter an elevated region called the volcanic / magmatic arc. This is where igneous activity is most intense. Intrusive igneous rocks are being emplaced at depth, whereas volcanic rocks of generally intermediate composition are erupting at the surface, creating large stratovolcanoes.

The lithosphere is thickest beneath the volcanic / magmatic arc due to addition of vast amounts of hot, buoyant magmatic material to the underside of the arc as this material rises from near the subducting plate, in effect, thickening the lithosphere from below. Your textbook authors refer to this process as magmatic underplating.

Isostatic rise associated with the thickened lithosphere of the magmatic arc creates tall mountains (Figure 5.6). Here's a photo (below) of the mighty Sierra Nevada, a northwest trending mountain range in eastern California. Most of the rocks in the Sierra are granitic rocks that formed within a Mesozoic-age volcanic / magmatic arc east of an ancient subduction zone that once existed just offshore of central and southern California:

Sierra Nevada Mountains.
(photo courtesy of Wikipedia; http://en.wikipedia.org/wiki/Sierra_Nevada_(U.S.)

In between the accretionary wedge and the magmatic arc is the forearc basin, which receives vast amounts of sediment eroded from the magmatic arc and sometimes also the accretionary wedge.

Sedimentary rocks within the forearc basin are much less deformed than those in the accretionary wedge.

As more and more sediments are added to the forearc basin, isostatic subsidence occurs, making room for still more sediments. Thus, the forearc basin sinks isostatically (due to loading from above), whereas the magmatic arc rises (due to thickening from underneath).

So, to summarize, Andean-style mountain belts are characterized by a deep trench, a complexly folded and faulted accretionary wedge adjacent to the trench, a forearc basin inboard from the trench, and a volcanic / magmatic arc inboard from the forearc basin.

In some cases, a fold and thrust belt develops landward of the arc, where layered rocks are horizontally shortened, stacked on top of each another like shingles, and thrust toward the continent in response to compressional stress associated with ocean-continent plate convergence (Figure 5.6).

A good example of a subduction-related fold and thrust belt is the northern Rocky Mountains, where gigantic stacks of sedimentary rock have been thrust eastward where the Earth's crust has been crunched together like a bunch of shingles or dominoes.

The origin of the Rocky Mountains has been a perplexing problem for geologists because the Rockies are so far inland from the Andean-type subduction zone that persisted along the west coast of North America throughout Mesozoic and into Cenozoic time. In a typical subduction zone, the subducting plate slides beneath the non-subducting plate at a fairly steep angle; however, in late Mesozoic time, the angle of the subducting plate may have been significantly flattened, shifting the focus of mountain building processes (e.g., magmatism, compressional faulting) far inland.

The geology of central California fits the Andean-style subduction model quite well.

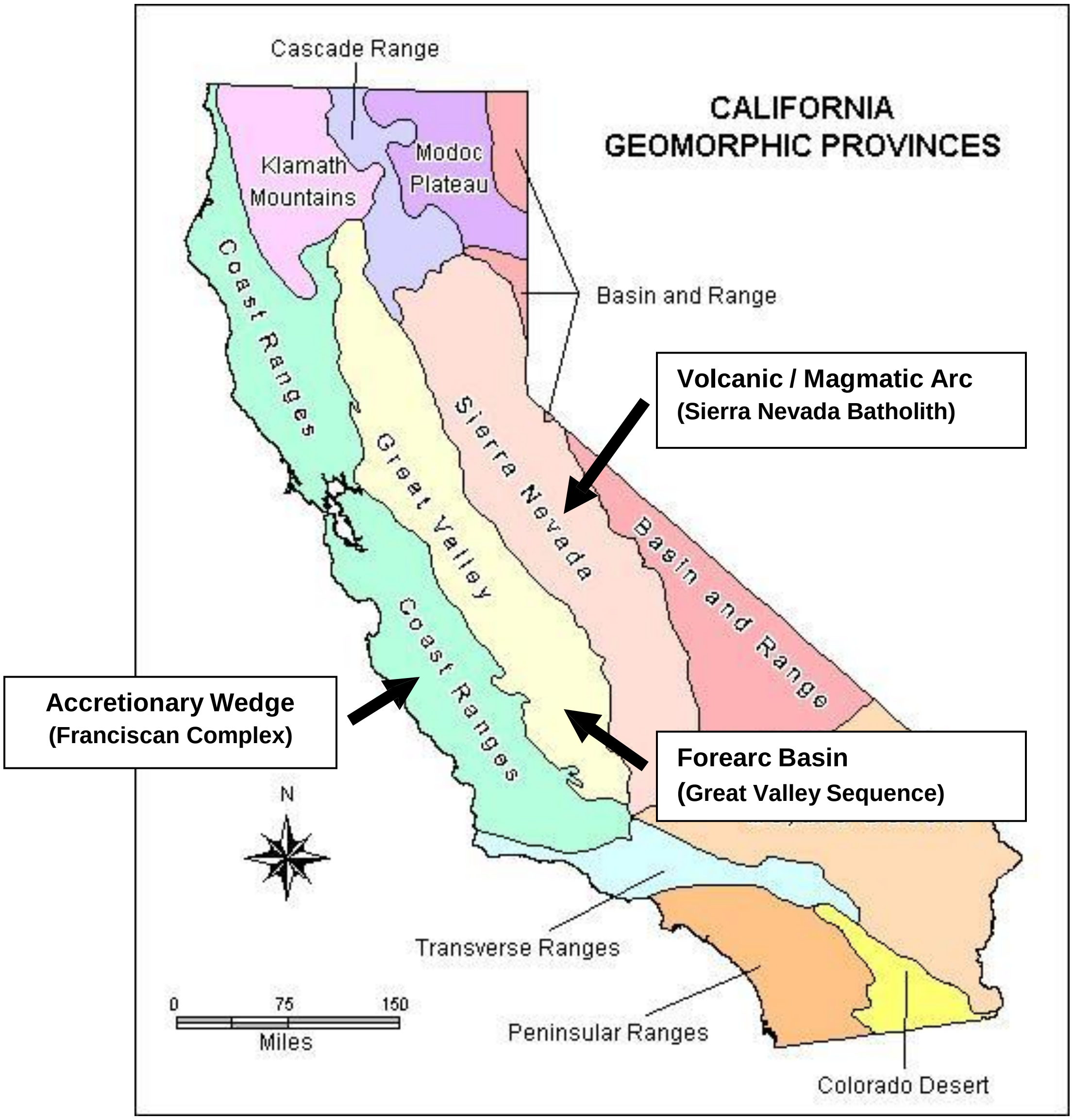
Along the central California coast, the Coast Ranges, including the Diablo Range east of San Francisco and the Santa Cruz Mountains, consist of an intensely deformed mixture of various rock types of mostly oceanic affinity known collectively as the Franciscan Complex.

With the development of plate tectonic theory, geologists recognized that Franciscan rocks correspond to Mesozoic-age accretionary wedge deposits.

Moving east, California's Central Valley region is underlain by thick sequences of relatively undeformed marine and non-marine sedimentary rocks that accumulated within a Mesozoic forearc basin.

Finally, the Sierra Nevada mountains east of the Central Valley represent the remains of a Mesozoic volcanic / magmatic arc. Known collectively as the Sierra Nevada batholith, Sierran granitic rocks are the cooled, crystallized remains of ancient magma chambers that once fed an active volcanic arc.

Here's a sketch to help you visualize these relationships:



California's geomorphic provinces.

Image courtesy of California Geologic Survey

http://www.conservation.ca.gov/cgs/geotour/Pages/Index.aspx

Before plate tectonic theory came along, geologists had a hard time making sense of Franciscan rocks in particular. How could such a complex, intensely deformed group of rocks have developed? With the advent of the plate tectonic model, the complexity of Franciscan rocks began to make sense. In effect, Franciscan rocks represent material scraped off an ancient, now-subducted oceanic plate and plastered onto the edge of the North American plate.

Arc-Continent Convergence (terrane accretion)

Imagine an old island arc, embedded in an oceanic plate, moving toward a subduction zone.

When the island arc reaches the subduction zone, it can’t subduct due to its high buoyancy, and it eventually clogs the subduction zone.

With continued compression, a new subduction zone may be initiated seaward of the arc, and the arc eventually becomes plastered onto the edge of the continent. When this happens, the island arc is effectively welded onto the edge of the non-subducting plate, causing the plate to grow laterally—a process called terrane accretion.

Here’s an interesting **animation of the accretion process.**

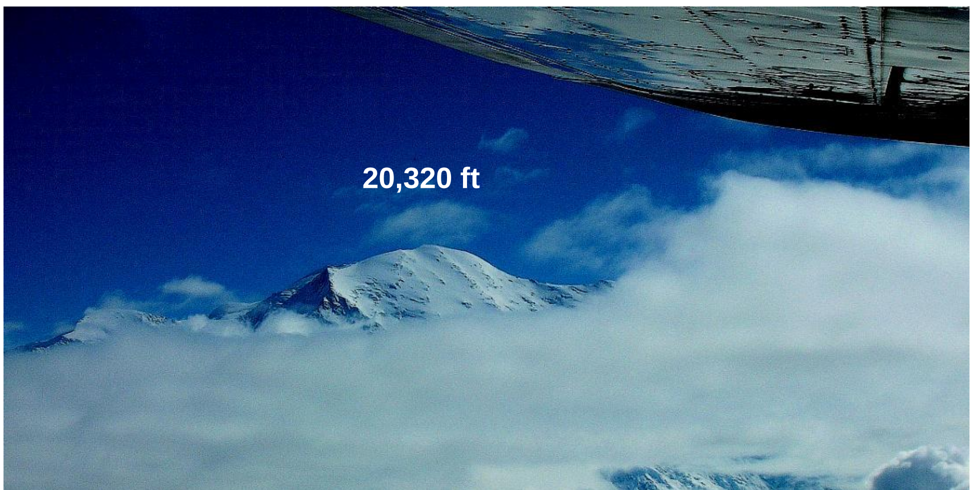
A terrane is a fault-bounded package of rocks that have a different geologic history than the rocks in adjoining regions. Notice the variant spelling. A *terrane* is an accreted, tectonic sliver, whereas the word, terrain, simply refers to the character of the land (e.g., hilly vs. flat terrain).

Island arcs can become terranes. So can sea floor rocks, accretionary wedge rocks, and even entire microcontinents. If a terrane originates from far away, it’s often referred to as exotic by geologists to remind us that the rocks we’re looking at came from someplace else.

Much of the land in North America west of the Rocky Mountains, including most of Alaska, is thought to be composed of exotic terranes that have been accreted to the North American plate since Mesozoic time.

Both ocean-continent convergence and arc-continent convergence (terrane accretion) create mountain belts near plate edges.

Speaking of accreted terranes, here’s a photo of North America’s tallest peak, Mt. McKinley. I shot this from a small plane on a visit to Alaska several summers ago:



Mt. McKinley (Denali).

Mt. McKinley (or Denali, as the Athabascans call it) is part of the Alaska Range and consists of 56 million year-old granitic rock that was intruded into older, exotic island arc terranes during their accretion onto the North American plate.

Denali is still rising at about 1 mm/year, pushed up by compressional tectonics and isostatic rise associated with nearby subduction of the Pacific plate beneath Alaska.

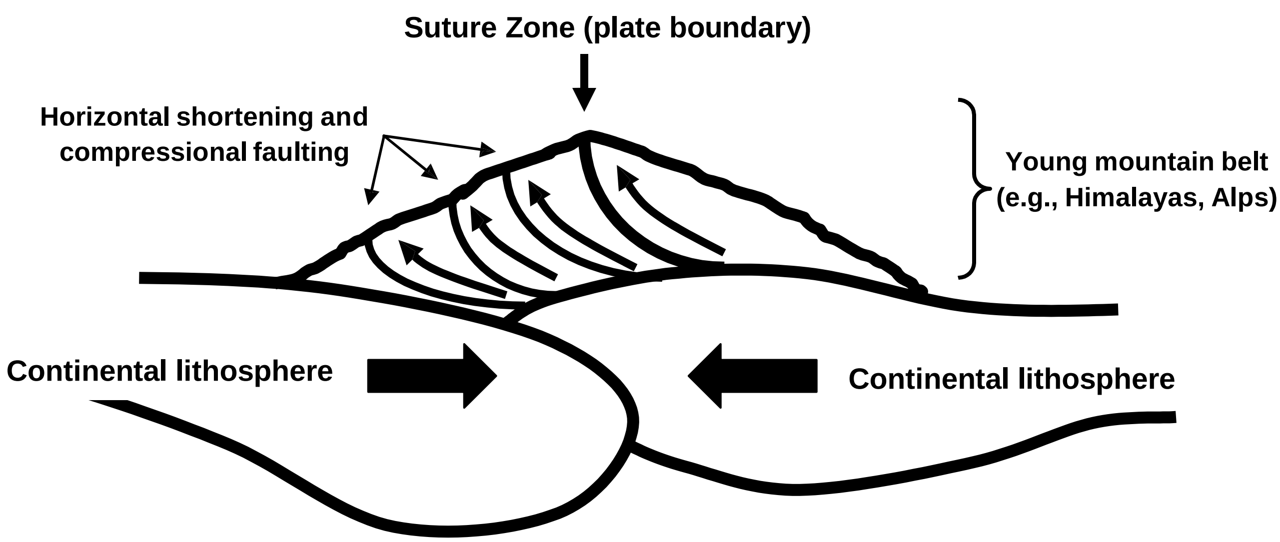
Continent-Continent Convergence (plate collision)—interior, “Himalayan-type” mountain belts

Mountain belts found within plate interiors as opposed to plate edges are thought to be the result of continent-continent convergence, where two continental plates collided and welded together to become one.

The Alps, Urals, Appalachians, and Himalayas are all examples of mountain belts formed by plate collision. In such settings, compressional stress and associated horizontal shortening can be intense, a cross section through the Alps. Like subduction zones, the lithosphere at collision zones becomes thickened, but in a different way than at subduction zones. Instead of becoming thickened from underneath by the addition of hot, buoyant magma, lithospheric thickening at collision zones occurs primarily by horizontal shortening, folding, and stacking of thrust sheets, one atop another. As the lithosphere thickens, its isostatic balance is upset, and isostatic uplift occurs as a result.

The lithosphere beneath the Himalayas is approximately 200 km thick, about as thick as anywhere on Earth, where the Indo-Australian plate is attempting (unsuccessfully) to subduct beneath the southern portion of the Eurasian plate, effectively doubling up lithospheric thickness in this region.

Here’s a simple sketch of a typical collision zone (“Himalayan-type” mountain belt):



Cross section of a collision zone. (continent-continent plate convergence)

In the above diagram, the lithosphere is thickened beneath the mountain belt as a result of accordion-style, horizontal shortening. At shallow depths within the crust, where the rocks are cool and brittle, thrust and reverse faulting predominate. At greater depths, where the rocks are warmer and more plastic, regional metamorphism and folding occur.

Post-Orogenic Uplift and Block-Faulting

You can just skim this section of Chapter 5. I won't hold you responsible for this material.

Continental Growth

This last section of Chapter 5 discusses how continents evolve. As your text states, "Continents grow bigger as mountain belts evolve along their margins."

Hopefully by now you can appreciate how amazing this simple statement really is. Not only can plate tectonic processes like subduction, terrane accretion, and collision explain the growth and development of mountain belts, such processes can explain how the continents themselves grow!

Nowhere is this better illustrated than along western North America, where dozens of large and small, fault-bounded slivers of continental and oceanic crust (terranes*)* have been mapped in recent years. In effect, much of North America can be thought of as a collage of accreted, exotic terranes.

San Diego itself is built atop a 100-million year old exotic island arc terrane. The exposed granitic rocks in the greater San Diego area formed in an ancient island arc that once lay perhaps thousands of kilometers offshore. Eventually, these rocks were accreted onto the leading edge of the North American plate. There's a very famous fault zone (famous among geologists, anyway ☺), the Cuyamaca-Laguna Mountain Shear Zone, just east of I-8 near Sunrise Highway, that marks the eastern boundary of this terrane.

What is true of western North America is probably also true, to some extent, of most other continents as well: the continents grow larger by terrane accretion as geologic time progresses. Eventually, they collide, creating supercontinents like Pangea. Following this, the supercontinents break up to repeat the cycle all over again.

Here's a link to a fascinating New York Times article where you can scroll through various **plate configurations from 200 million years ago to 250 million years in the future**, when the next supercontinent, dubbed Pangea Ultima, may form…Wow!

References Cited:

Oreskes, Naomi. 1999. *The Rejection of Continental Drift: Theory and Method in American Earth Science.* Oxford University Press.

Geologic Structures

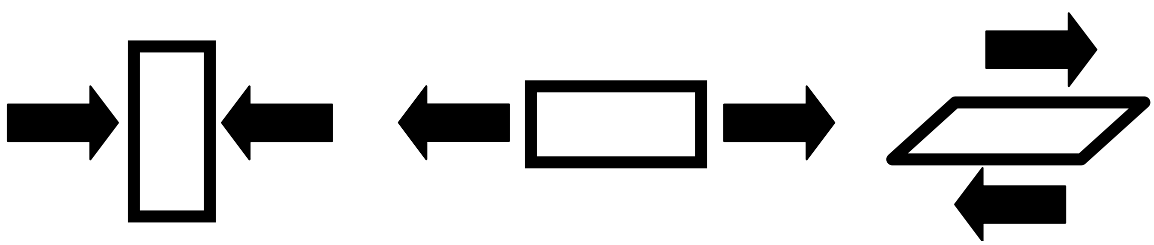
Geologic structures include faults, folds, and joints. Before we discuss the details of each type of structure, let's distinguish between stress and strain.

Stress, defined as force per unit area, is what causes rocks to deform (change shape).

You might recall from the chapter on metamorphic rocks that stress also causes the development of foliation (mineral alignment).

Below figures summarize the different types of stress, including compressional, which results in shortening or flattening; tensional (also called tensile), which results in stretching and elongation; and shear, which results in a sort of smearing.

Here are some sketches to help you visualize the different types of stress:



Compressional stress **Tensile stress** **Shear stress**

Types of stress

In the above diagram, notice that different types of stress are distinguished in terms of their effects.

In other words, stress can't be observed directly. Instead, we infer the type of stress from its effects:

- If horizontally directed compression stress is applied, shortening occurs in a direction parallel to the direction of maximum compressional stress.

- If horizontally directed tensile stress is applied, elongation occurs parallel to the direction of maximum tensile stress.

- If horizontally directed shear stress is applied, smearing occurs parallel to the direction of maximum shear stress.

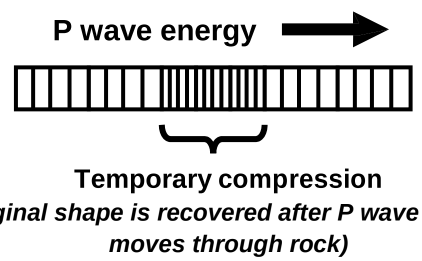
Deformation caused by stress is called strain.

The type of rock, the temperature and pressure, and even the rate of stress all influence how a rock will accommodate strain.

In summary, rocks can behave in an elastic, ductile (plastic), or a brittle fashion when stressed:

- Elastic strain occurs if the material recovers its initial shape after the stress is reduced or removed.
- Plastic or ductile strain occurs when the rock bends but doesn't break.
- Brittle strain occurs when the rock breaks.

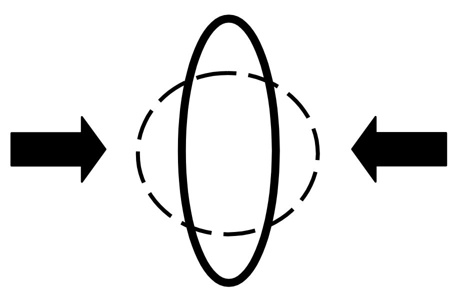
Here are some quick sketches of the various strain behaviors discussed above:



Elastic strain.

Seismic waves are a good example of elastic strain, as shown in the diagram above. As a P-wave moves through a rock mass, for example, the rock mass will compress, but after the P-wave moves through, the rock mass will uncompress and return to its original shape.

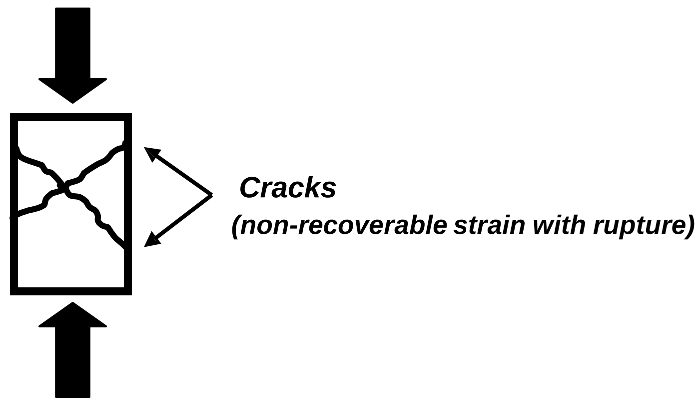
Under increased stress, a rock mass will continue to strain elastically, but only to a point. Beyond the elastic limit, a rock mass (or a blob of silly putty) will bend, as shown below:



Initially spherical ball of silly putty is permanently squashed upon compression ***(non-recoverable strain without rupture)***

Plastic (ductile) strain.

Finally, if stress is applied rapidly, as shown below, a rock mass will permanently rupture, creating cracks (brittle strain), as shown below:

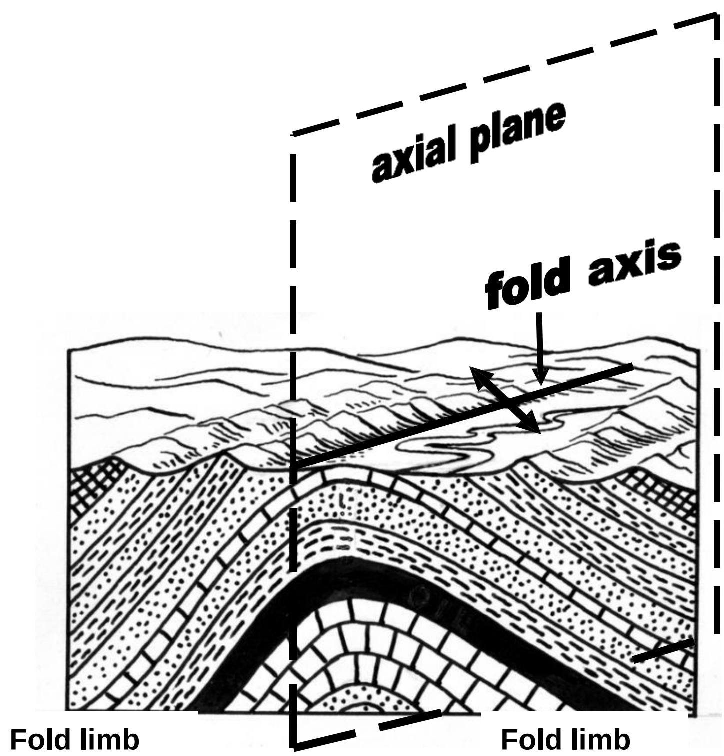


Brittle strain.

In general, rocks subjected to tectonic stress at plate boundaries will exhibit brittle behavior at or near Earth's surface and ductile behavior in Earth's interior where temperatures and pressures are higher. Low rates of stress are more conducive to ductile behavior, whereas higher stress rates, particularly at low temperatures and pressures, typically result in brittle behavior.

Folds

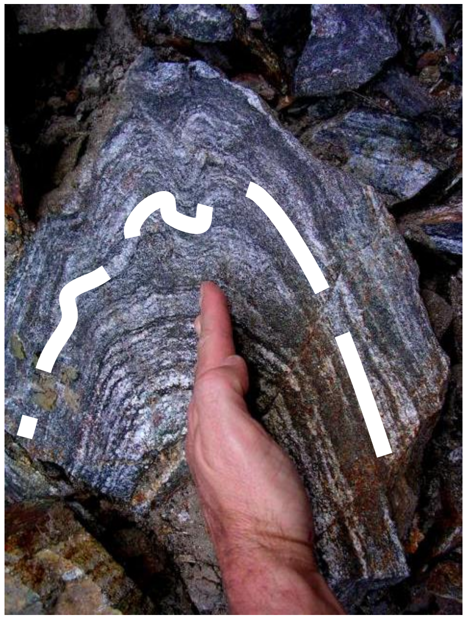
As shown in Figure 6.11, folds are wavelike bends in layered rock. Upward-arching folds are called anticlines, whereas downward-arching folds are called synclines. To remember this, try associating the word "anticline" with the letter, A, which sort of looks link an upfold. On the other hand, "syncline" sort of reminds one of the word, "sink", which resembles a downfold. Here's a sketch of an anticline:



Anticline
(modified from Wikipedia; http://en.wikipedia.org/wiki/Anticline)

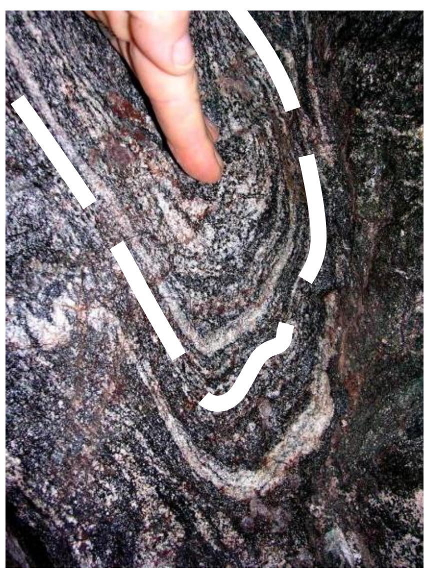
In the above sketch, the imaginary *plane of symmetry* that bisects the fold is called the axial plane. The *line of symmetry* formed by the intersection of the axial plane with the folded surface is called the fold axis, or hinge line. The sides of the fold, which tilt away from the axis, are called the fold limbs.

Here's a photo of a small anticline (below). My hand represents the orientation of the axial plane of the fold.



Anticline.

…and a small-scale syncline:



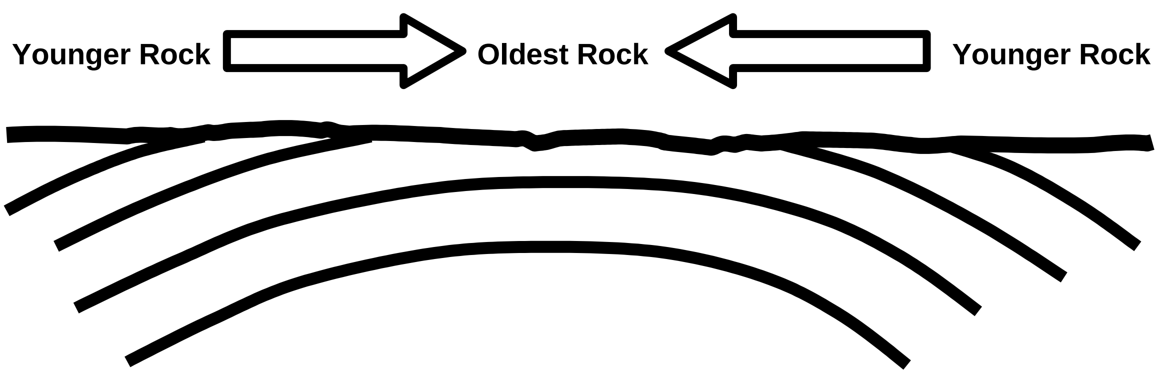
Syncline.

…and a spectacular anticline - syncline pair:

Anticline – Syncline Pair, Split Mountain Gorge
Anza-Borrego Desert.

An important aspect of anticlines and synclines is how the relative age of the folded layers change as one moves from the limbs toward the axis of partially eroded folds exposed at the Earth's surface.

The diagram below shows a partially eroded anticline (upfold) exposed at Earth's surface. Imagine walking from each limb toward the fold axis. How does the relative age of the rock layers you cross change?



Anticline shown in cross section.

As shown in the above diagram, the oldest layers are exposed in the center (axial region) of an eroded anticline. In other words, the relative age of the folded layers increases as one moves from the limbs toward the axis of eroded anticline. ***The strict definition of an anticline is an upfold with the oldest rocks exposed in the axial region of the fold.***

Likewise, eroded synclines display the opposite age trend: the rock layers become progressively younger as one moves from the limbs toward the axial region of a syncline.

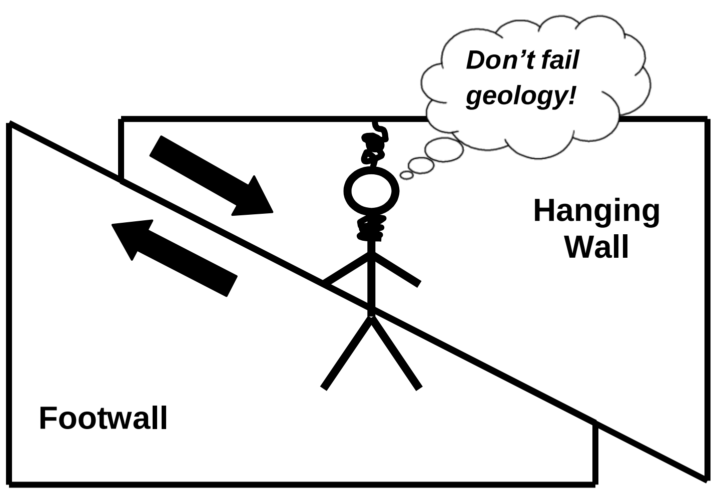
Your text goes into great detail about fold geometry (plunging vs. non-plunging folds, domes and basins, open vs. isoclinal folds, etc.). Much of this material is covered in greater depth in the laboratory course (Geology 101), so I won't hold you responsible for the specifics of fold geometry in this course.

Folds can develop in a variety of ways in response to any type of stress. Here's an interesting link that shows two ways that folds can develop: **Flexural Slip vs. Flow Folds** . Since most folding is associated with ductile behavior, folded rocks typically form at depth under generally increased pressures and temperatures. You might remember that these are generally the same conditions necessary for metamorphism, so it should come as no surprise to learn that metamorphic rocks are often folded.

Faults

Faults are brittle fractures along which movement has taken place. For dip-slip faults, movement is parallel to the dip of the fault plane and typically shows a strong component of vertical motion. In strike-slip faults, movement is primarily horizontal.

To help distinguish different types of dip-slip fault, it's helpful to contrast the hanging wall (head wall) and the footwall. The hanging wall is the side of the fault above the fault surface, whereas the footwall is the side below the fault surface. Your text explains that this terminology comes from mining—the side of the fault a minor hangs a lantern on will be the hanging wall. Here's an easy way to distinguish the hanging wall and the footwall of a dip-slip fault. In the cross section sketch below, the hangman's noose is in the hanging wall, whereas the hangman's feet are touching the footwall:



Definition of hanging (head) wall vs. footwall.

In the diagram above, notice that the hanging/head wall is sitting on top of the fault…but also moving *down* relative to the footwall.

Now that you're familiar with the terminology, here's how we distinguish between normal and reverse faults:

- For a **normal fault**, the **hanging wall moves down relative to the footwall**. Normal faults develop in response to tensile stress.

- For a **reverse fault**, the hanging **wall moves up relative to the footwall**. Reverse faults develop in response to compressional stress.

A thrust fault is simply a low-angle reverse fault. Like other reverse faults, thrust faults also develop in response to compressional stress.

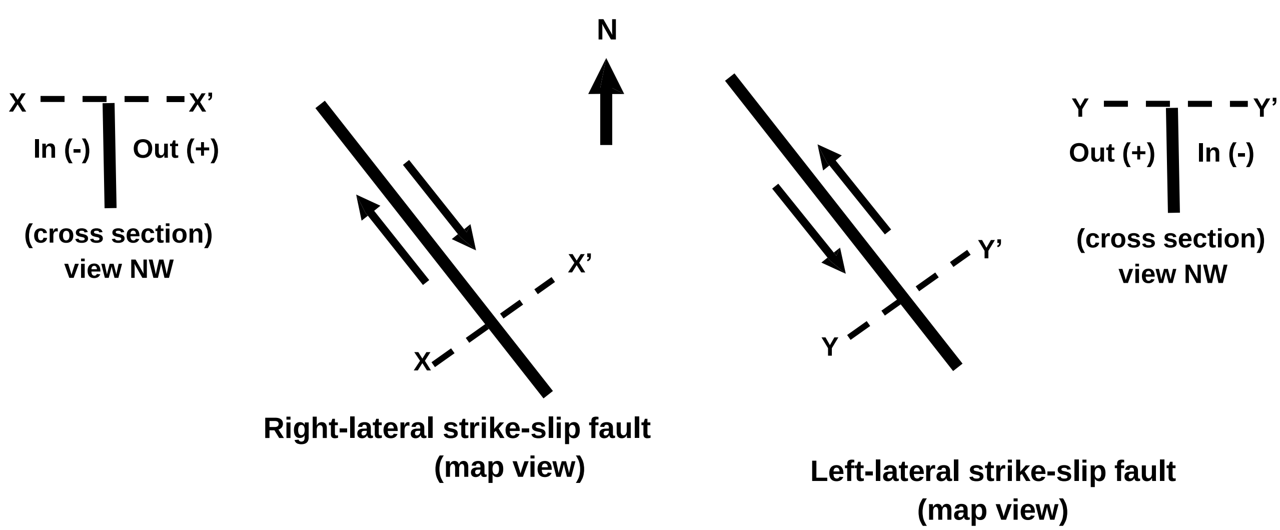
Make sure you understand how the sense of movement across a dip-slip fault allows us to define the type of dip-slip fault it is.

By the way, is the hangman fault drawing shown above a normal or reverse fault? If you answered "normal fault" please email me for an honorary A+!

Strike-slip faults are also distinguished on the basis of the relative movement across the fault, except that the concept of headwall and footwall doesn't apply.

Right-lateral strike-slip faults, where the opposite side of the fault moves to an observer's right as the observer faces the fault. A left-lateral strike-slip fault occurs when the opposite side of the faults moves to an observer's left as the observer faces the fault.

Here are some simple drawings of right- and left-lateral strike-slip faults in map view and cross section:



Strike-Slip faults in map view and cross section.

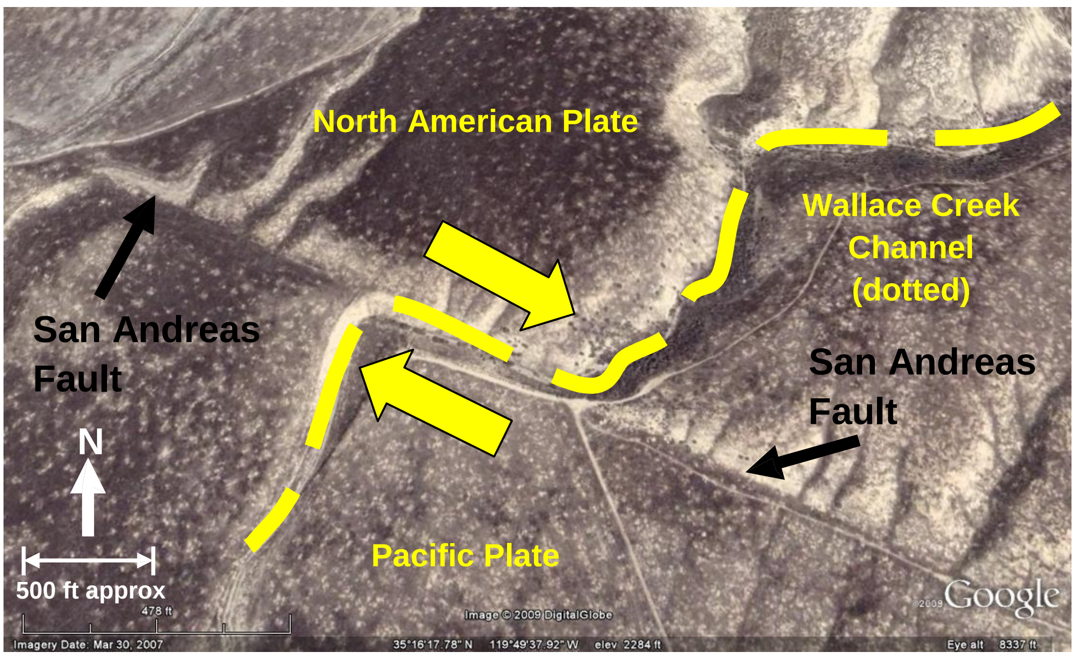
In the cross sectional drawings above, the side of the fault labeled out (+) is moving out of the plane of the page, whereas the side labeled in (-) is moving into the plane of the page when viewed in a northwest direction.

The famous San Andreas Fault is an example of a right-lateral strike-slip fault. To download an animated, interactive image of the location, check out this **[Virtual Tour of the San Andreas Fault.]**

In the above-referenced animation, the channel of Wallace Creek has actually been offset by the fault! At this location, the San Andreas Fault forms the main boundary between the Pacific and North American plates.

Geologists and geology students come from all over the world to have the experience of standing right on top of the plate boundary between the Pacific and North American plates at this location. I've done it myself at this very spot! Pretty cool.

Here's a Google Earth shot of this famous site:



San Andreas Fault at Wallace Creek (image courtesy of Google Earth).

Fault Types shows animations of normal, reverse, thrust, and strike-slip faults.

Joints

The only difference between faults and joints is that joints don't show evidence of slippage whereas faults do.

Joints are arguably the most common geologic structure and can form in a variety of ways.

Here's an example of joints developed as a result of pressure release (unloading):

Unloading joints in granodiorite. Image courtesy of Marli Miller, University of Oregon Image source: Earth Science World Image Bank (http://www.earthscienceworld.org/images)

Joints can also form as a result of tectonic stress associated with regional uplift. In the photo below, the two visible joint sets form an x-shaped pattern, which probably indicates that this rock mass is being horizontally stretched as it's being uplifted.

Jointed sandstone, Anza-BorregoState Park.

Under certain conditions, lava flows can cool slowly enough for shrinkage stresses to be distributed evenly throughout the cooling lava, resulting in these incredible columns:

Columnar basalt, Mexico
(photo courtesy, U.S. Geological Survey)

Make sure you can distinguish and describe the following geologic structures: anticlines, synclines, normal faults, reverse faults, thrust faults, right-lateral strike-slip faults, left-lateral strike slip faults, and joints.

Let's end this chapter by highlighting the general relationship between various geologic structures and plate tectonic setting. First, a disclaimer: all of the geologic structures mentioned so far can be found at any type of plate boundary. That said, we can make some broad generalizations about the relationship between plate boundary processes and geologic structures.

At convergent boundaries (subduction zones and collision zones), reverse faults and thrust faults are expected to form at shallow depths, where the rocks are cool and behave in a brittle fashion under compressional stress. At greater depths, we might expect folding to be common where the rocks are warmer and under higher pressures.

At divergent boundaries, where the lithosphere is being stretched due to tensile stress, normal faults are common near Earth's surface, where the lithosphere is cool and brittle.

At transform boundaries, strike-slip faults are common. Where strike-slip faults bend, localized compression and extension can develop, which may give rise to associated tensile structures (normal faults) or compressional structures (reverse faults and folds).

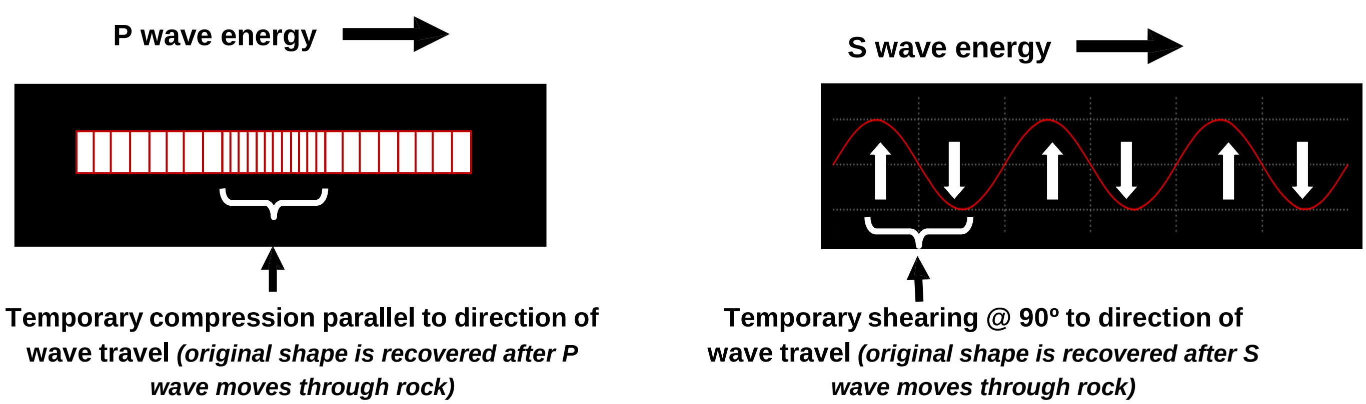
Chapter 11: Earthquakes

An earthquake occurs, quite literally, when the earth quakes. Most earthquakes are caused by subsurface fault rupture. As a fault moves, energy waves are generated that propagate outward in all directions, similar the movement of ripples in a pond outward in all directions from where a pebble tossed into the pond landed. In the case of a pebble landing in a pond, the impact of the pebble deforms the water surface, and energy waves are created as a result. In the case of an earthquake, the deformation of Earth's interior via fault movement creates seismic energy waves. Your text mentions other ways that earthquakes can be generated, including magma movement, mineral transformations, and even dehydration of water-rich minerals like serpentine. For the purposes of this chapter, however, we'll focus on fault-produced earthquakes.

One thing that might not be obvious about is that the seismic waves propagate outward from the earthquake source, or focus, in 3 dimensions. Imagine a series of energy spheres moving outward from the earthquake focus. If the earthquake location is to be shown in map view, it's often more convenient to plot the point on the ground surface directly above the earthquake focus—a point called the earthquake epicenter. Notice again that the epicenter of an earthquake doesn't correspond with the surface trace of the fault on which the earthquake originates.

Most earthquakes happen suddenly, which makes them difficult to study in real time. Luckily for geologists, most aspects of earthquakes can be studied by the energy pulses they create, called seismic waves (Figure 7.5). Seismic waves come in two basic types: (1) body waves that travel in Earth's interior, and (2) surface waves that travel at or near Earth's surface. For the most part, we'll focus on body waves, which themselves come in two types, including primary (P) waves and secondary (S) waves. As discussed below, two types of surface waves include Love and Rayleigh waves.

Here's an online link that demonstrates the movement associated with P, S waves, and Surface (Love and Rayleigh) waves: **Seismic Waves—** Here's a diagram showing the way that P and S waves deform the solid Earth as they move through:



Elastic properties of P and S waves.

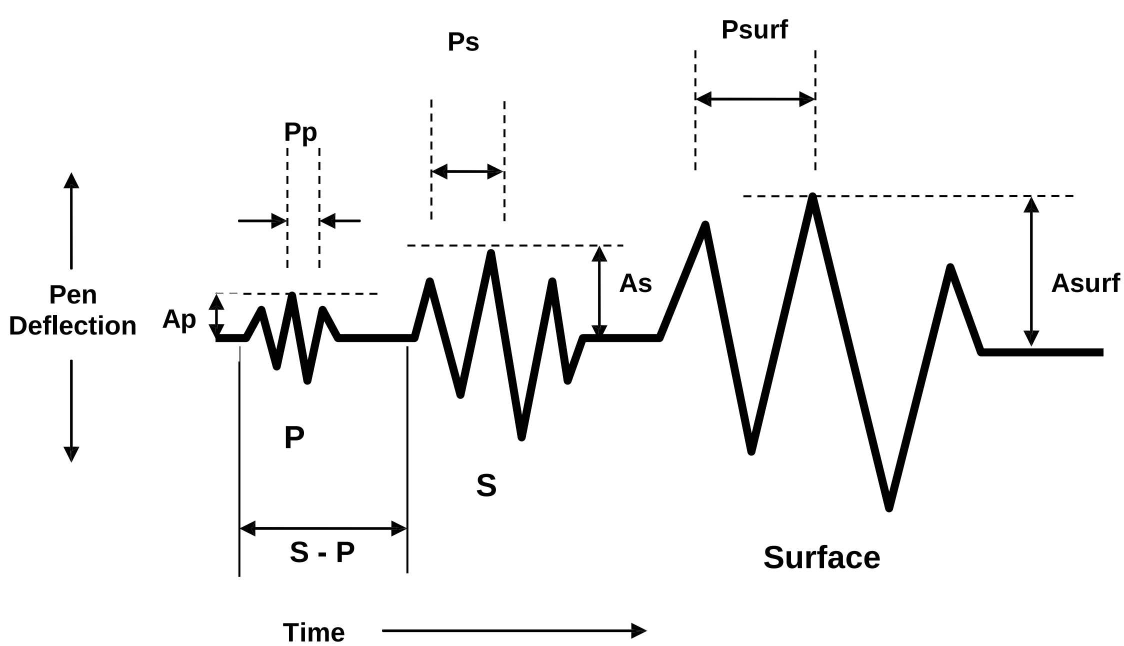
Notice that P waves cause compressional, elastic vibration parallel to the direction of wave movement. S waves, by contrast, propagate by elastic shearing, which causes Earth's interior to vibrate perpendicular to the direction of wave movement. Love waves are like sideways S waves--they propagate sideways in a horizontal plane. Rayleigh waves are a bit like rolling ocean waves.

In terms of the relative velocities of various seismic waves, P waves travel the fastest (4-7 km/sec), followed by slightly slower S waves (2-5 km/sec), followed by surface waves (slowest). P waves can travel through solids, liquids, and gases, whereas S waves can only travel through solids. Since liquids and gases don't possess shear strength, S waves (which propagate via shearing) can't pass through liquids or gases.

Earthquake waves are recorded on instruments called seismographs. The permanent record of an earthquake recorded on a seismograph is called a seismogram. As we'll see below, many aspects of earthquakes can be studied from seismogram records. Here's a link that illustrates **How a Seismograph Works (scroll down to 8.3).**

Notice that the drawing pen is essentially stationary with respect to the rest of the instrument. Truth be told, the animation you just looked at illustrates how older seismographs work. In many modern seismographs, an electronic sensor in Earth's shallow interior converts ground shaking into an electrical signal, which in turn gets converted into pen vibration or even just a squiggle on a computer screen. In any case, the squiggly lines (i.e., wave arrivals) on the seismogram tell us a lot about earthquakes.

Your text doesn't have a detailed figure of a labeled seismogram record, so here's a typical (albeit simplified) seismogram with various seismic wave features labeled:



Typical seismogram record.

The following important points apply:

- The seismogram record shown above is essentially an x-y plot of pen deflection verses time.

- The three wave arrivals are easy to spot, with P waves arriving first, S waves arriving next, and Surface waves arriving after the S waves.

- Each wave set consists of several individual waves of each type.

- The height of each wave is called wave amplitude (shown by vertical arrows—**Ap, As,** and **Asurf** for P, S, and Surface wave amplitudes). Notice that wave amplitude for a particular wave set is measured from the background level of the seismogram up to the top of the tallest wave for a particular wave set (not from trough to crest).

- The time for each wave to complete one wave cycle is called wave period (shown by horizontal arrows, measured from the crest of one wave to the crest of the next wave—**Pp, Ps,** and **Psurf** for P, S, and Surface wave periods).

- The time gap between the first S waves and the first P waves (**S - P** on seismogram) is called the -S minus P time difference‖, or -delta t (Δt)‖. Notice that the S–P time difference is just the gap in time between the arrival of the first P waves the first S waves.

- In general, P waves will be the first to reach a distant location (because they travel at the fastest velocity), but P waves will shake the ground less violently than S and Surface waves (because P wave amplitudes are typically smaller). In addition, P waves will cause the ground to shake in more of a rapid, vibratory manner than S and Surface waves because P wave periods are typically lower than S and Surface wave periods.

- Since P waves travel faster than S waves, Δt increases with increasing distance from the earthquake origin. This can actually be felt in some cases. Have you ever felt a quake as a single jolt, like someone backed a truck into a building? If so, you probably felt an earthquake that originated nearby, before the P waves could get too far ahead of the S waves. On the other hand, have you ever felt a quake that starts slowly at first, then, after a few seconds, really becomes violent? In this case, the initial shaking you felt probably marked the arrival of the first P waves, with their smaller wave amplitudes, followed a few seconds later by the more violent (larger-amplitude) S waves. In this case, the quake was far enough away that the P waves got a few seconds ahead of the S waves.

A True Story…

Back in 1989, there was a moderately large, Magnitude 7.1 earthquake in the Santa Cruz Mountains that shook the entire San Francisco Bay region, resulting in severe damage and a number of deaths—the Loma Prieta earthquake.

At the moment the earthquake struck, my mom, who lives in the San Francisco area, was talking on the phone with her sister in New Orleans. For the first few seconds of the quake, my mom was not worried, casually commenting to her sister that, -We're having an earthquake.‖

At this point, my mom was only feeling the P waves from the distant quake, which had relatively small wave amplitudes. A few seconds later, the shaking became more violent, and dishes began to rattle…

At this point, mom told her sister, -Oh, this is a pretty decent size quake!‖ By then, she was undoubtedly feeling the first S wave arrivals with their larger wave amplitudes…

Several seconds later, the last thing my aunt heard before the phone line went dead was my mom screaming into the phone, -OH MY GOD, THIS IS TERRIBLE!!‖ At this point, the large-amplitude surface waves had arrived. The phone line was severed, and my aunt wasn't able to reach my mom for several days!

Luckily, my mom wasn't hurt, our house survived with only some minor cracking, and I got a good story out of it!

Your book has a well-written section on measuring and locating earthquakes. In general, the S-P time difference (Δt) can be used to figure out the distance to an earthquake, because as distance to an earthquake increases, so does the S-P time difference.

Near where an earthquake occurs, the P waves will arrive only slightly ahead of the S waves, and the S-P time difference will be small. But for distant locations, the P waves will arrive well ahead of the first S waves.

Such information can tell us the distance to an earthquake source but not the direction. A minimum of three seismogram records is needed to pinpoint, or triangulate, the location of the earthquake on a map (i.e., the epicenter). Here's a link to an animation showing how an earthquake epicenter can be located: **Earthquake Location—.**

Another aspect of an earthquake is its size, which can be measured in two ways: (1) the damage caused, or (2) the amount of energy released. I won't say much about the first method; you can read about this on your own (see—Modified Mercalli Intensity Scale).

For many scientific applications, the size of the earthquake is represented by a single number, called magnitude, which is related to the energy released. Earthquake magnitude can be calculated from seismic wave amplitudes. In general, the larger the wave amplitude, the more energy released, and thus the larger the earthquake magnitude.

Richter Magnitude

The most famous earthquake magnitude scale is of course the Richter scale. To calculate Richter magnitude, the largest amplitude of a seismic wave is measured at a distance of 100 km from the earthquake epicenter.

You might see a problem already. What if we're not lucky enough to have a seismograph set up at exactly 100 km from the earthquake epicenter? Well, actually, this isn't a problem, because we know enough about how seismic wave amplitudes shrink (attenuate) with distance as they lose energy that if we can measure the wave amplitude at any distance, we can apply a correction factor to calculate what the amplitude would be at exactly 100 km from the earthquake.

For the mathematically inclined, here's Richter's original definition of magnitude, courtesy of the **[Southern California Earthquake Center](#)**:

–The (Richter) magnitude of any shock is taken as the logarithm of the maximum trace amplitude, expressed in microns, with which the standard short-period torsion seismometer would register that shock at an epicentral distance of 100 kilometers.‖

Nobody uses the true Richter scale anymore, because the –standard short-period torsion seismometer‖ Richter originally used to measure wave amplitudes has gone the way of the dinosaur.

Today, much better, more sensitive instruments are used. Even so, modern earthquake researchers use a variety of Richter-like scales to measure earthquake magnitude, many of which are also based on the maximum amplitude of various seismic waves.

Here's the equation for calculating Richter Magnitude (below)...don't worry! You don't need to memorize this equation, and I won't ask you to solve it. I'm just including it so you can see how the math relates to what we've been discussing:

$$\textbf{Richter Magnitude } (\mathbf{M_L}) = \mathbf{Log_{10}A + 3{*}Log_{10}(8\Delta t) - 2.92}$$

A is max. wave height (amplitude).

Δt is the S–P time difference.

Don't be concerned with the logarithmic nature of the above equation. The term, Log_{10} is just a mathematical operation Richter used to shrink down very large numbers to a more manageable size.

Bottom line, if we know both **A** (max. amplitude) and **Δt** (S – P), we can calculate $\mathbf{M_L}$ (Richter magnitude). How do we determine **A** and **Δt?** From a seismogram record!

In effect, the $\mathbf{3{*}Log_{10}(8\Delta t) - 2.92}$ term in Richter's equation is just a distance correction factor to shrink or stretch the wave amplitude to what it would be at exactly 100 km from the earthquake epicenter.

Moment Magnitude

The most objective measure of earthquake magnitude is called Moment Magnitude, and is based on a calculation called seismic moment:

$$\textbf{Seismic Moment } (\mathbf{M_o}) = \textbf{Fault Slip x } \underbrace{\textbf{Rock Strength}}_{\textit{shear modulus, actually…don't worry about the details ☺…}} \textbf{ x Fault Plane Area}$$

Using the above calculation for seismic moment (**M_o**)**,** seismologists have devised the moment magnitude scale:

Moment Magnitude (M_w) = 2/3 x $Log_{10}(M_o)$ – 10.7

Once again, don't worry about the math. Just appreciate that the larger the seismic moment (**M_o**), the larger the moment magnitude (**M_w**)**.**

In general, larger moment magnitudes are associated with longer fault ruptures, stronger rocks, larger fault offsets, and greater energy release.

On the one hand, moment magnitude gives a much better estimate of the energy released during large earthquakes than Richter magnitude.

Unfortunately, it can take several weeks to calculate the moment magnitude of an earthquake after it occurs, because the area of the fault plane is typically estimated from the distribution of aftershocks. But once calculated, the moment magnitude of a quake is the best measure of the energy released during the quake.

Earthquake Magnitude and Energy

Because the Richter and moment magnitude scales are logarithmic (i.e., exponential), earthquake energy increases rather dramatically with increasing magnitude.

For every 1-point increase on the Richter scale, seismic wave amplitude increases by 10 times; however, energy release increases by approximately 32 times. You can see this effect clearly on Figure 7.12 for moment magnitudes as well. A magnitude 2 quake releases about as much energy as 120 pounds of explosives, whereas a magnitude 6 earthquake releases about as much energy as 120 *million* pounds of explosives, or almost as much energy as an atomic bomb blast! That is, a magnitude 6 quake releases about a million times more energy than a magnitude 2 quake! Wow!

Here's a quick, back-of-the-envelope calculation you can do to estimate the approximate difference in energy release between two earthquakes of different Richter magnitude:

Larger magnitude – smaller magnitude = result ➡ 30^{result} = difference in energy release.

Let's do a quick example.

Approximately how much more energy is released in a Mag 7 quake than a Mag 4 quake?

7 – 4 = 3 ➡ 30^3 = 30 x 30 x 30 = 27,000

So a Mag 7 quake releases approximately 27,000 times more energy than a Mag 4 quake!

If you have the interest and the time, here's a wonderful online exercise you can do that illustrates how seismologists actually determine earthquake location and magnitude: **Measure and Locate an Earthquake**…it'll take you about 20 minutes. In fact, I may make this a 5-pt exercise (I'll let you know!). After completing this exercise, you'll be given the opportunity to receive a personalized certificate as a ―virtual seismologist.‖ Pretty cool!

By this point in the course, you should be able to explain the general relationship between earthquakes and plate boundaries. Simply put, most earthquakes are concentrated along linear and arc-shaped belts that define active plate boundaries. In fact, as your text points out, plate boundaries are identified and defined by earthquakes. Combining our knowledge of plate tectonic theory with our understanding of earthquakes, we can make some broad generalizations about the earthquake activity (seismicity) associated with various plate boundary types:

- At divergent plate boundaries, earthquakes are typically shallow, restricted to a narrow band, and much lower in magnitude than those at convergent and transform boundaries. The reason that earthquakes at divergent boundaries are typically small is that rock is typically weak in tension. That is, when subjected to tensile stress along a divergent plate boundary, rock breaks quite easily, before large stresses can build up. This results in small earthquakes.

- At transform boundaries, earthquakes are typically shallow, range in size from small to large and are confined to a narrow bands along individual transform faults. Along the transform boundary between the Pacific and North American plates, the seismicity occurs in a very wide band because this boundary is defined by dozens of individual faults, not just the San Andreas Fault.

- At collision (i.e., continent-continent convergence) zones, earthquakes are shallow to intermediate in depth, range in size from small to large, and are characterized by broad zones of seismicity.

- At subduction zones (characterized by continent-ocean convergence and ocean-ocean convergence), earthquakes range in size from small to large and occur at shallow, intermediate, and deep depths.

In fact, very deep earthquakes *only* occur at subduction zones. This is because the cold, brittle oceanic plate that subducts can still behave in a brittle (breakable) fashion to depths of up to about 700 km. Everywhere else, Earth's mantle is too plastic at that depth for the rocks to break. If the rocks can't break, earthquakes can't occur.

It's actually possible to track the movement of subducting plates because they produce an inclined zone of seismicity that dips into the mantle, underneath the adjacent continent or island arc. These dipping seismic zones, called Benioff zones, were actually discovered before the development of plate tectonic theory, and remained a mystery until the subduction process was understood. Can you imagine the excitement of the first geologists who realized that these strange, dipping seismic zones—Benioff zones—were seismic proof of the subduction process?! Another connection between earthquakes and plate tectonics is that many devastating tsunamis like the 2004 tsunami off the coast of Sumatra are caused by vertical seafloor motion associated with large subduction zone quakes. Here's an animation of how a submarine earthquake can generate a tsunami: **Tsunami Animation— .**

Feel free to skim the very interesting section on the effects of earthquakes, which include ground motion, ground-surface displacement, landslides, and tsunamis.

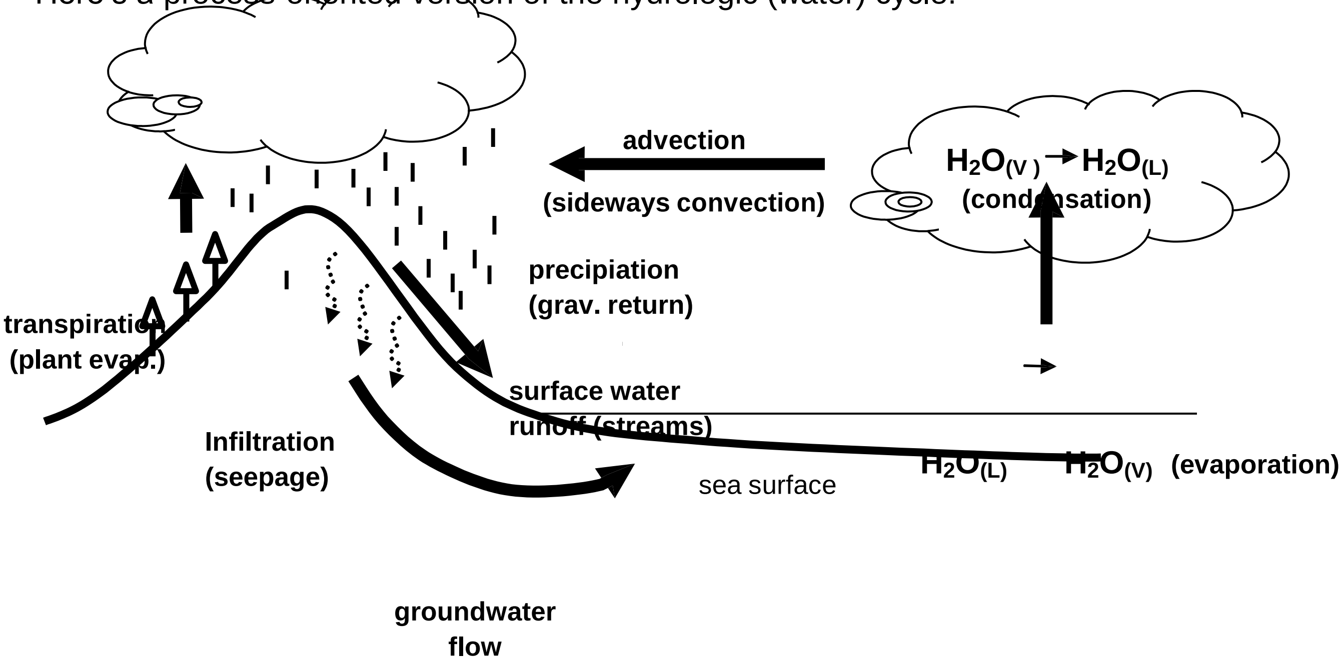
Chapters 16 and 17 - Surface Waters and Groundwater

Rivers, Streams and Floods

Running water is perhaps the most important erosional agent that sculpts Earth's landscapes. Even arid environments owe their distinctive look to the effects of running water. This chapter focuses on the various processes associated with running water.

The global movement of Earth's water can be envisioned as a series of interconnected processes that cause water to change state (i.e., melt, freeze, evaporate, condense) and to move (convect, advect, infiltrate, flow) from the oceans, to the atmosphere, to the continents, and back to the oceans again, as shown in the hydrologic cycle.

Here's a process-oriented version of the hydrologic (water) cycle:



Hydrologic cycle.

In the diagram above, each arrow represents a process by which water is transformed or moved.

Evaporation, condensation, and transpiration represent changes in state, whereby water is converted from liquid to gas (evaporation, transpiration) and back to liquid (condensation).

Convection and advection represent the mass movement of water within the atmosphere.

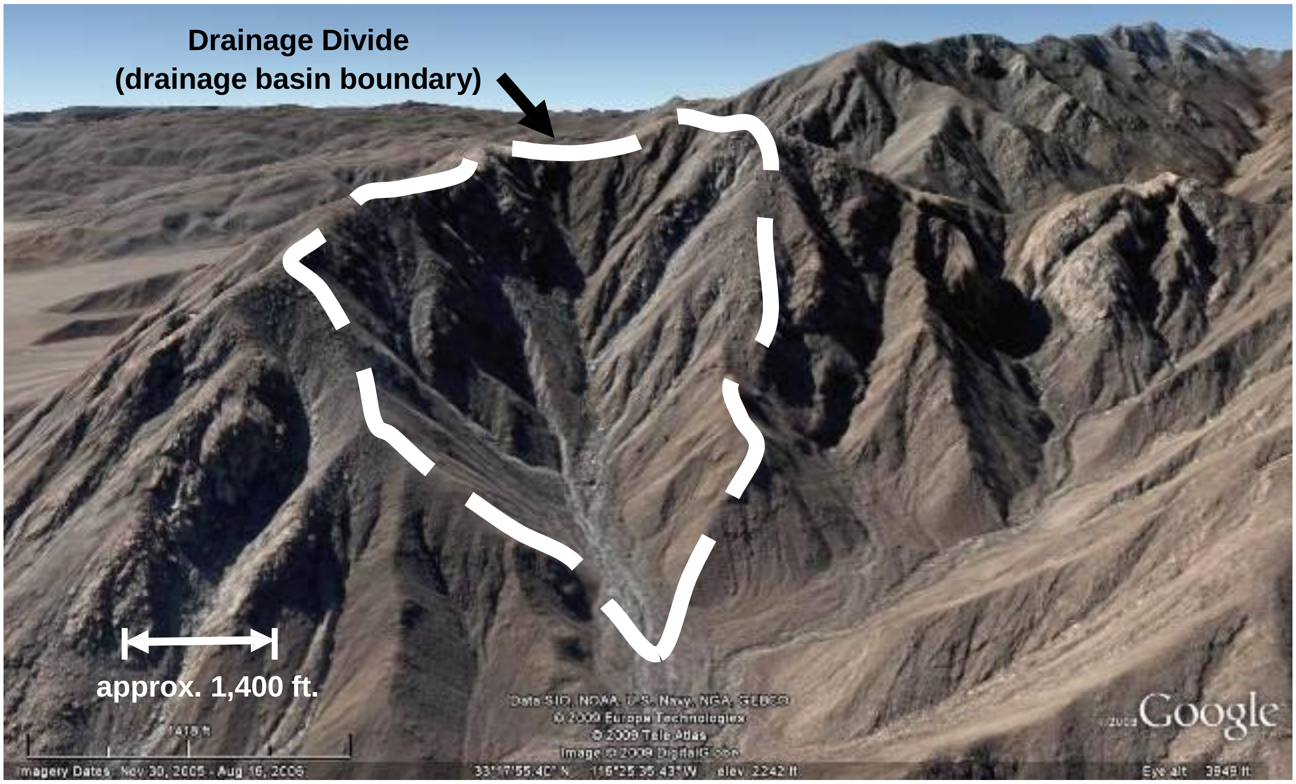
Precipitation, surface water runoff, infiltration, and groundwater flow represent the mass movement of water at or near Earth's surface.

This chapter focuses on the movement of running water in stream channels under the influence of gravity.

As landscapes develop, they’re typically carved up by a complex pattern of connected stream channels that comprise a drainage basin—the total area drained by a stream and its tributaries.

As shown in Figure 16.5, drainage patterns typically reflect the underlying geology of an area, with a dendritic pattern developing in uniformly eroding rock, a radial pattern developing on a conical mountain, dome, or volcano, a rectangular pattern developing in jointed rock, and a trellis pattern developing on a landscape characterized by ridges and valleys.

Here's an image of a small drainage basin just west of Borrego Springs, California:



**Small drainage basin in San Ysidro Mountains,
near Borrego Springs, CA** (image courtesy of Google Earth).

In the above image, several small tributary stream channels can be seen joining the main stream toward the bottom of the drainage basin, creating a dendritic drainage pattern.

Any rain that falls within the boundaries of the drainage basin will ultimately flow into the main stream and out of the drainage basin at its downstream end.

When rain water falls on a landscape (or ice melts) to become surface water runoff, it first flows over the ground surface as non-channelized, sheet flow.

Eventually, small irregularities in the land surface concentrate the flowing water into small channels called rills, which eventually merge into larger channels called streams.

Streams are very dynamic; they can change dramatically from one season to the next, or even overnight. Therefore, it's helpful to describe various aspects of the stream that are subject to change—stream variables:

1. **velocity** (how fast the water in the stream moves)
2. **gradient** (the slope of the stream bed)
3. **channel shape** (geometry) and **roughness** (irregularity)
4. **discharge** (volume of water flowing past a given point in a unit of time, e.g., ft^3/sec).
5. **capacity** (the total amount of sediment transported by a steam)
6. **competence** (the size of the largest particle transported by a stream).

There are complex relationships among the variables listed above. Typically, changing one variable results in a change in the others. Although it's not always easy to predict how the other variables will change given a chance in one or more variables, we can make some generalizations:

- At a given point along the course of a stream, **velocity, capacity, and competence all increase as discharge increases**. As more and more water flows through the stream channel, the water typically moves faster, because it experiences less friction from the sides and bottoms of the stream channel. Higher water velocity means that the stream can transport a greater amount of sediment, so capacity increases. Higher water velocity also means that the stream can move larger particles, so competence also increases. Both capacity and competence increase dramatically with increasing discharge.

- **Most of the sediment transport in a stream is accomplished during floods** when stream discharge and stream velocity (and therefore competence and capacity) are many times their level during lower discharge. During a flood, the power of a stream increases, and so the stream bed is eroded (scoured). As the flood subsides, discharge and stream velocity (and therefore competence and capacity) decrease, and sediment is deposited in the bed of the stream.

- Stream velocity is also affected by stream gradient, channel shape, and channel roughness. In general, **stream velocity increases as gradient increases**, because the slope of the stream bed is steeper. **Stream velocity is generally faster in a narrow, deep, semi-circular channel** because frictional resistance is less in such a channel than in a wide, shallow channel (Figure 16.8). **A stream flows more rapidly in a smooth channel** and more slowly in a rough, boulder-strewn channel, which creates greater frictional resistance for the moving water.

- Although it seems counter-intuitive, **the linear velocity of the water in a stream generally increases on average in the downstream direction**, from the dashing mountain stream to the broad peaceful river. Even though stream gradient decreases in the downstream direction, the broad lowland river has a much greater discharge and smoother channels, so it flows much more freely (i.e., the water doesn't have to dash around boulders in the stream). The net result is that linear velocity actually increases somewhat in the downstream direction.

Streams erode their beds and banks in three ways, including

1. **hydraulic action** (the force of the water itself)

2. **abrasion** (grinding away by friction between moving rock particles and the stream bed/banks)

3. **solution** (dissolving by water)

Once sediment is added to the bed of the stream, the stream transports it downstream in three modes, including:

1. **bed load** (heavy particles that travel on the stream bed)

2. **suspended load** (smaller particles that remain in the water above the stream bed by turbulence)

3. **dissolved load** (soluble products of chemical weathering)

Modes of Sediment Transport— is a nice animation of these three modes of sediment transport.

A true story…

In my younger days, I spent a lot of time whitewater kayaking on the Kern River, a large stream that drains the western slope of the southern Sierra Nevada mountains. The Kern is a wonderful place for whitewater because one can find everything from very easy to very difficult rapids within a few minutes of Kernville, a small mountain town on the Kern River, about an hour's drive east of Bakersfield. Rapids are typically rated on a scale from Class I (very easy) to Class VI (very dangerous). As a kayaking instructor, some of the rafting guides I worked with would tell their passengers that Class I is so easy you could bring your mother in law along, whereas Class VI is so difficult, you'd definitely want to bring your mother in law along! (just kidding, by the way…truth be told, my mother in law rocks!).

One aspect of the river that kayakers and rafters pay a lot of attention to is the discharge on a particular day. At increased discharge, a leisurely Class III rapid can become a challenging Class IV rapid, or even a dangerous Class V rapid. The whitewater shop in Kernville always posts the discharge so boaters can better judge the difficulty of the rapids they're going to do…

One weekend in late spring a number of years back, some friends and I drove up to Kernville for some whitewater kayaking. The Kern is a snowmelt-fed stream, and by this time, most of the snow had already melted, so the discharge on Saturday was quite low, perhaps 600 cubic feet per second (cfs). The water was slow and clear that day, and we had a grand ol' time bopping along between the rocks clogging the channel.

At the breakfast café across from the paddling shop the next morning, we couldn't believe our eyes when we saw the discharge…In the span of maybe 12 hours, the discharge had increased from about 600 to 6,000 cfs! Whoa! Apparently, a warm, late-season rain storm had melted massive amounts of snow up-drainage during the night. By Sunday morning, the discharge had skyrocketed down in Kernville. I was itching to do some challenging whitewater. My buddies decided to pass, but I decided to give it a try (first *mistake). At the put-in, I met a young woman I'd never boated with before but figured we'd be okay judging from how highly she spoke about her whitewater skills (second mistake)…*

Well, the river had totally changed! What was a fairly easy stretch of Class III+ whitewater on Saturday had become a torrent of Class IV+ rapids on Sunday, with tall waves, big drops, logs, and…lots of "holes" or "keepers"…recirculating curls of whitewater just downstream from waterfalls and large rocks. These nasty buggers are like a kayaker's version of a roach motel…kayakers check in but don't check out ☹…Being on the water that day was like being inside a washing machine.

First, the water velocity had increased dramatically with increased discharge, so not only were the rapids larger and more difficult, they came at us much faster. Second, capacity and competence had also increased…so much so that the water was the color of coffee (due to increased suspended sediment load), and it rumbled audibly from the large boulders rolling along in the main channel just underneath our boats (increased bedload).

Within minutes after getting on the water, my erstwhile companion went swimming, meaning she flipped upside down and ejected from her boat. The situation was now serious, because just downstream we saw a "horizon line", a placid-looking line across the river where one can't see what's downstream. The problem with horizon lines, of course, is what one does find downstream…a waterfall or a hole (keeper). In panic, my companion climbed up on the back of my boat, weighing me down like an anchor, while we drifted toward—and over—the horizon line into the hole below. Luck was with us that day, and through no skill on my part we both plowed through the hole without getting stuck. Finally, I was able to get my companion (and myself) to shore…we fished her boat out of some brush about half a mile downstream but never did recover her paddle…

The moral of this story? Don't be a dumb-ass and do difficult whitewater with a stranger…On the geologic side, the whole experience gave me a better appreciation for how dramatically stream variables like water velocity, capacity, and competence can change with discharge.

Meandering vs. Braided Channels

Two basic types of stream channels are meandering channels and braided channels.

Meandering (sinuous) channels tend to be narrow and deep in comparison with braided channels, and typically develop under low stream gradients in fine-grained sediments.

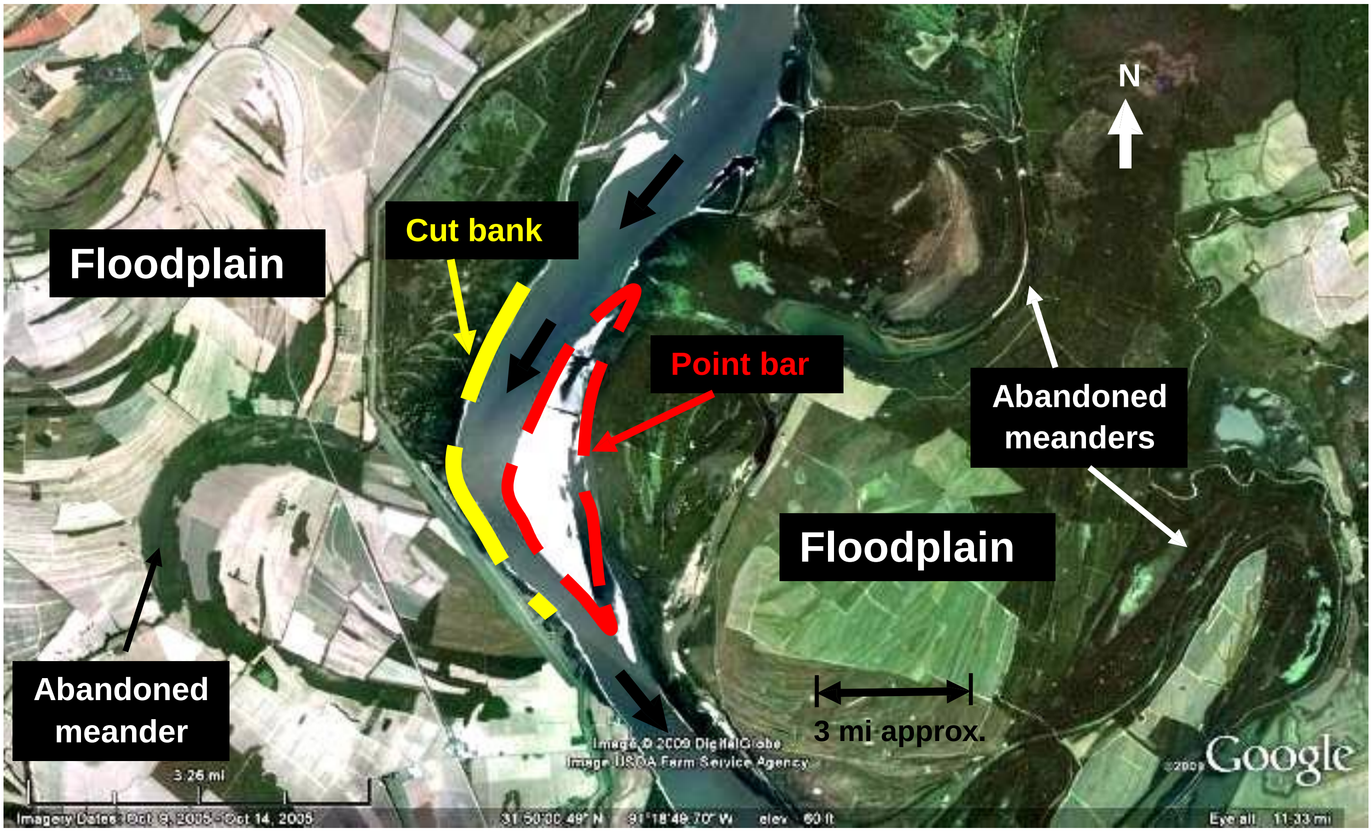
Typically, the water moving through a stream meander travels more slowly along the inside of the meander, which leads to deposition and the creation of a point bar.

On the outside bend of a meander, the water moves more rapidly, which causes erosion and the development of a cut bank .

Click the following link to see an animated version of **Meander Development—.**

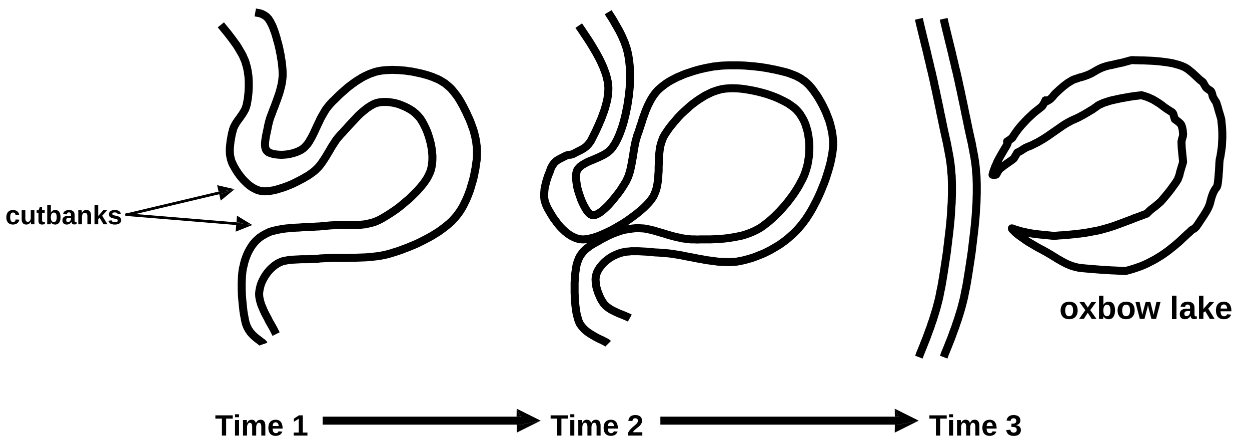
On occasion, the cut banks of two separate meanders can join to create a meander cutoff and an oxbow lake. Click on **Oxbow Lake Creation—** to see the animated version.

Here's an aerial image of the Mississippi River, a classic meandering stream, on the Louisiana/Mississippi border, upstream from New Orleans:



Mississippi River.
(image courtesy of Google Earth)

Abandoned meanders (formerly oxbow lakes) can be easily seen in the above image. Eventually, the abandoned meander on the left side of the image will be consumed by cut bank erosion as the channel migrates laterally across its flood plain, from right to left. Here's a sketch showing how an oxbow lake develops as cutbanks migrate toward each other and eventually merge:



Meander cutoff and oxbow lake development.

Braided stream channels are different from meandering channels in almost every respect.

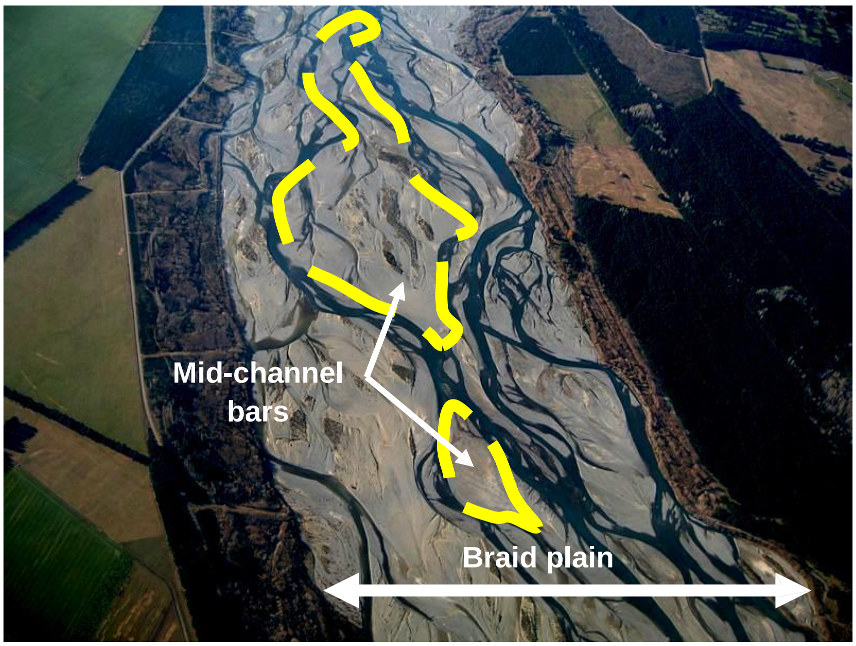
Typically, braided streams consist of an interconnected network of small channels that flow between mid-channel sand and gravel bars distributed across a wide braid plain.

Braided streams are heavily loaded with sediment, and typically have a greater proportion of bed load (sand and gravel) in comparison with the total load than meandering streams.

Sediment bars in a braided stream tend to be lenticular, with long axes roughly parallel to the stream flow, and tend to develop in the middle of the stream channel .

Your text points out that such mid-channel bars divert flowing water toward the sides of the channel, where it washes against the banks with greater force, eroding and widening the stream channel.

Here's an annotated photograph of a braided stream (below):



Braided stream in New Zealand.

(photo courtesy of Wikipedia; http://en.wikipedia.org/wiki/Braided_river)

Braided channels are favored by steeper gradients, higher sediment loads, and more easily erodible bank material.

Based on what you know about the differences between meandering and braided streams, see if you can select the correct answers in the table below (answers underneath the table…no peeking!).

	Channel Geometry*—ratio of width/depth	Bar Shape*	Bar Location*	Gradient*	Load Characteristics*—ratio of bedload/total load
Meandering Streams	High or Low (*select correct answer*)	Crescent- or Lens-shaped (*select correct answer*)	Mid-channel or Side-channel (*select correct answer*)	High or Low (*select correct answer*)	High or Low (*select correct answer*)
Braided Streams	High or Low *(select correct answer*)	Crescent- or Lens-shaped (*circle correct answer*)	Mid-channel or Side-channel (*circle correct answer*)	High or Low (*circle correct answer*)	High or Low (*circle correct answer*)

**Generalized, relative comparison between meandering and braided streams.*

Answers

In comparison with braided streams, meandering streams are generally not very wide compared to their depth (Low width/depth ratio); they tend to form (crescent-shaped) point bars on the sides of their channels (side-channel bars); they typically flow over relatively flat (low-gradient) terrain, and carry a lot of fine-grained sediment (silt and clay) in suspension compared to bedload (low bedload/total load ratio).

In comparison with meandering streams, braided streams are generally quite wide and not very deep (high width/depth ratio); they tend to form longitudinal (lens-shaped) bars in the middle of their channels (mid-channel bars) roughly parallel to the flow; they typically flow over steeper (high-gradient) terrain; and they typically carry a lot of medium-coarse grained sediment (sand & gravel) as bedload (high bedload/total load ratio).

So what do streams create? Valleys! In fact, most streams cut the valleys they occupy. Your book points out that stream-cut valleys are Earth's most common landforms.

Streams also modify the valleys they create by downcutting, valley widening, and valley lengthening.

As a stream erodes material from its bed and banks and transports this material downstream, it cuts a deeper and deeper valley, particularly in its upper reaches. The tendency of a stream is to downcut its bed vertically, toward base level, the lowest level to which downcutting can occur. Mass wasting and sheet erosion wear away the sides of a valley even as downcutting proceeds, so the result is typically a V-shaped valley.

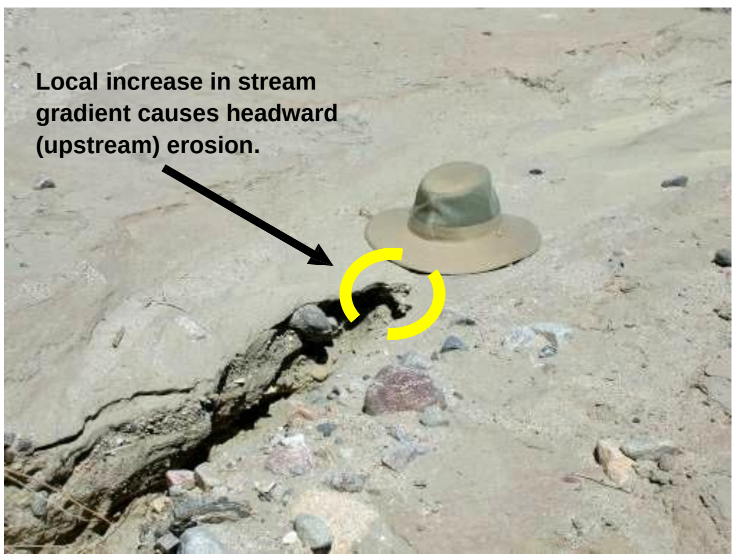
Valley widening is a dominant process in the lower reaches of a stream, near base level. As a stream migrates laterally across its flood plain, the walls of the valley are occasionally undercut via cut bank erosion and mass movement, both of which widen the valley.

A stream can also lengthen its valley in the downstream direction by building a delta or alluvial fan, or at its upstream end via headward erosion.

Here's an image of a portion of the Mississippi River delta downstream from New Orleans, where the river flows into the Gulf of Mexico:

Mississippi River delta.
(image courtesy of Google Earth)

Here's a photo illustrating headward erosion:



Headward erosion.

In the above photo, my hat marks the upstream end of a new channel being cut into an alluvial fan surface. Notice that the local gradient increases sharply at the head (upstream end) of the channel. As surface water runoff flows into the new channel, the head of the channel migrates upstream (toward my hat) via headward erosion, lengthening the channel with time.

Chapter 20 - Shorelines: Waves, Beaches, and Coasts

Living in San Diego as many of us do, much of our local geography and geology are the result of coastal processes operating today or in the not so distant geologic past. Our emphasis in this chapter will be on various coastal landforms and the processes that give rise to these landforms.

Feel free to skim over the first part of this chapter, which focuses on water waves and surf. These are important topics, but they're better addressed in oceanography, so we'll move on to wave refraction and longshore currents.

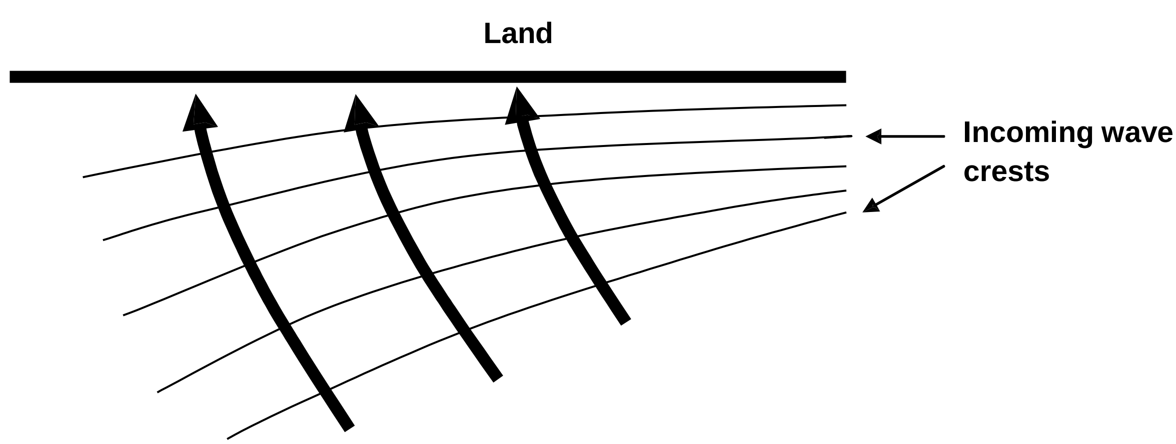
You might remember from our discussion of Earth's interior that refraction refers to the bending of waves.

In the context of this chapter, the bending (refraction) of wind-generated water waves near the coast is what creates longshore currents—currents that move both water and sediment parallel to the coast.

Just like seismic waves, wind-generated waves will refract when one part of the wave moves at a different velocity than another part of the wave. This is what happens when the waves approach the shoreline at an angle (Figure 20.5 A).

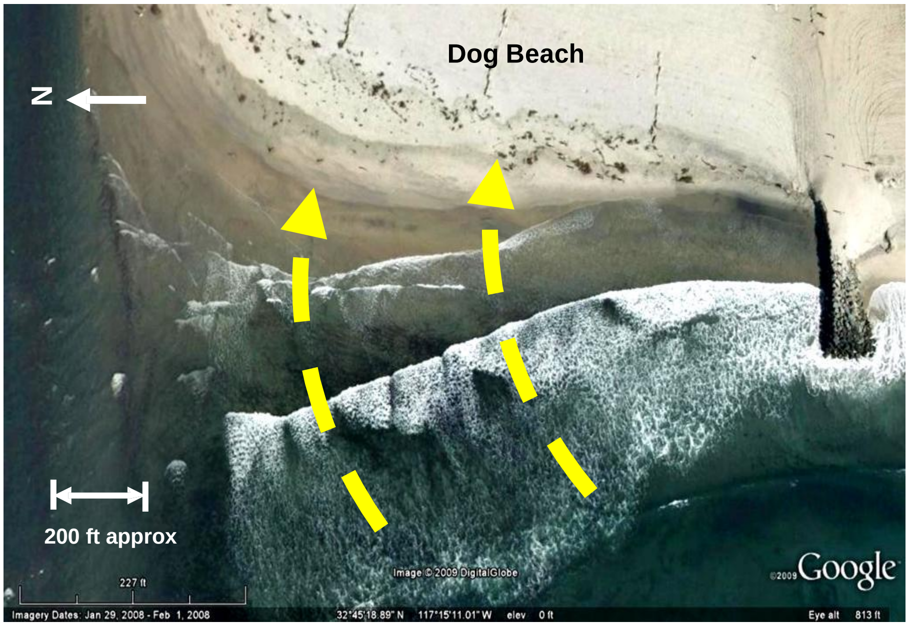
Along the crest of an incoming wave, bottom friction begins to slow the wave down as the wave moves into shallower water. If the incoming waves arrive at an angle to the coastline, this causes the portion of the wave crest in shallower water to slow down more than the rest of the wave. The result is that incoming wave crests bend toward the coastline.

This effect is illustrated in Figures 20.5 A and B, and also here:



Coastal wave refraction.

Here's an image of coastal wave refraction at Dog Beach in San Diego, just south of the Mission Bay jetty:

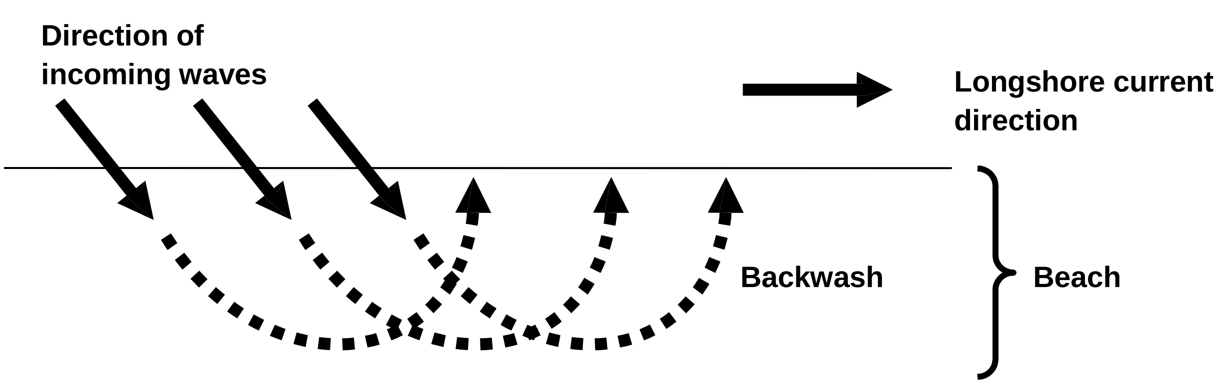


Wave refraction

(photo courtesy of Google Earth).

Coastal wave refraction is rarely sufficient to cause the waves to strike the coast directly head-on, and so even after refracting, the waves typically still strike the coast at a slight angle as they break and wash up onto the shore.

Rather than rushing up and down the beach face in a back-and-fourth line, the breaking waves rush up and down along a zigzagging path. It is the motion of the water in such zigzags that sets up a net flow of water parallel to the shoreline—what we call a longshore current. As your text points out, longshore drift is simply the movement of sediment parallel to the shore due to longshore currents. Here's a sketch of this process:



Longshore current development.

Here's an animation that demonstrates **Longshore Drift—Fig. 20.9.**

Longshore currents are associated with rip currents, but they're not the same thing.

In contrast to longshore currents, rip currents travel straight out to sea, perpendicular to the shore, not parallel to the shore. Rip currents are fed by the piling up of water on a beach as the water moves down the beach in the zigzagging pattern described above.

Every so often, so much water piles up on the beach that it rushes seaward as a single mass, which results in a strong, seaward current (Figure below).

As surfers know, rip currents typically die out just past the surf zone, and they're relatively narrow. Therefore, surfers will sometimes catch a rip current to get out to where the waves are bigger. If you're ever caught in a rip current as a swimmer, swim parallel to the shore and you'll eventually get out of it. Rip currents can be very dangerous if you're not aware of this!

Here's an image of some rip currents along the Baja, CA coast. The rip currents can be distinguished by gaps in the line of breaking waves and by plumes of sediment created as the currents wash nearshore material out beyond the surf zone.

Rip Currents along Baja, CA coast.
(image courtesy Google Earth)

A True Story…

A few summers ago, I got caught in a rip current. I was swimming with a friend at Mission Beach, heading into shore, and rather suddenly we noticed that no matter how hard we swam we weren't making any headway.

In fact, we felt ourselves being pulled out to sea! After about 30 seconds, my friend got tired and started to panic. It was at this point that I realized we were caught in a rip current, and as calmly as I could I told my friend to swim parallel to the shoreline. After only a few strokes, we exited the rip current, and easily made it into shore. Wow! I have a new respect for rip currents!

Now that we understand wave refraction and longshore currents, we'll turn our attention to coastal landforms—coastal features created by processes operating at or near the shoreline. To make things easier, we'll divide such features into depositional and erosional, because virtually all coastal landforms develop in response to erosion or deposition (or sometimes both).

Depositional landforms include beaches, spits, baymouth bars, tombolos, and barrier islands. You should read about each of these features in your text. I'll briefly describe each feature:

Beach—a strip of sediment along a coast, extending from the low tide line inland to a cliff or zone of permanent vegetation:

Torrey Pines State Beach
(from Wikipedia; http://en.wikipedia.org/wiki/Torrey_Pines_State_Beach)

Spit—a fingerlike ridge of sediment created by longshore drift. Here's an animation: **Spit—Fig.** and also a photo:

Sand spit, Washington state
(photo courtesy of Wikipedia; http://en.wikipedia.org/wiki/Spit_(landform))

Baymouth bar—a ridge of sediment (formerly a spit) that cuts off a bay from the open ocean, formed by sediment migrating across what was formerly an open bay. Here's an animation: **Baymouth Bar—Fig.** and also a photo:

Baymouth bar, Yasawa Islands
(photo courtesy of Wikipedia: http://en.wikipedia.org/wiki/Shoal)

Tombolo—a ridge of sediment formed by longshore drift that connects a nearshore island to the mainland. Here's an image of the Silver Strand, a tombolo that connects Coronado Island to mainland San Diego:

San Diego's Silver Strand, a tombolo
(image courtesy of Google Earth)

Barrier island—a ridge of sediment parallel to the shore that extends above sea level and is sometimes stabilized by vegetation:

Barrier islands off the Mississippi coast.
(image courtesy of Google Earth)

Erosional landforms include headlands, sea cliffs, sea arches, sea caves, wave-cut platforms, and stacks, as described below:

Headland—a point of land that extends seaward, created by wave erosion:

Headland west of La Jolla Cove (image courtesy of Google Earth).

Sea cliff—a steep (usually rocky) slope located at the inland edge of the shore, created by undercutting and mass wasting due to wave erosion:

Sea cliff at Black's Beach, San Diego (image courtesy of Google Earth).

Sea arch—a hole through a headland created by differential erosion:

Sea Arch, Newport, Oregon
(photo courtesy U.S. Geological Survey)

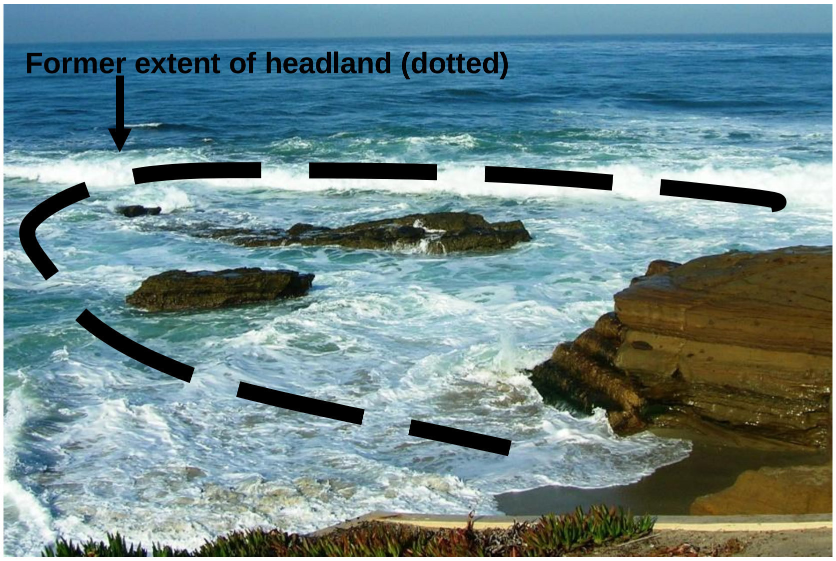
Sea cave—a roofed indentation created by wave erosion and undercutting:

Sea cave near La Jolla Cove.

Wavecut platform—a horizontal or gently sloping bench of rock formed by wave erosion and cliff retreat:

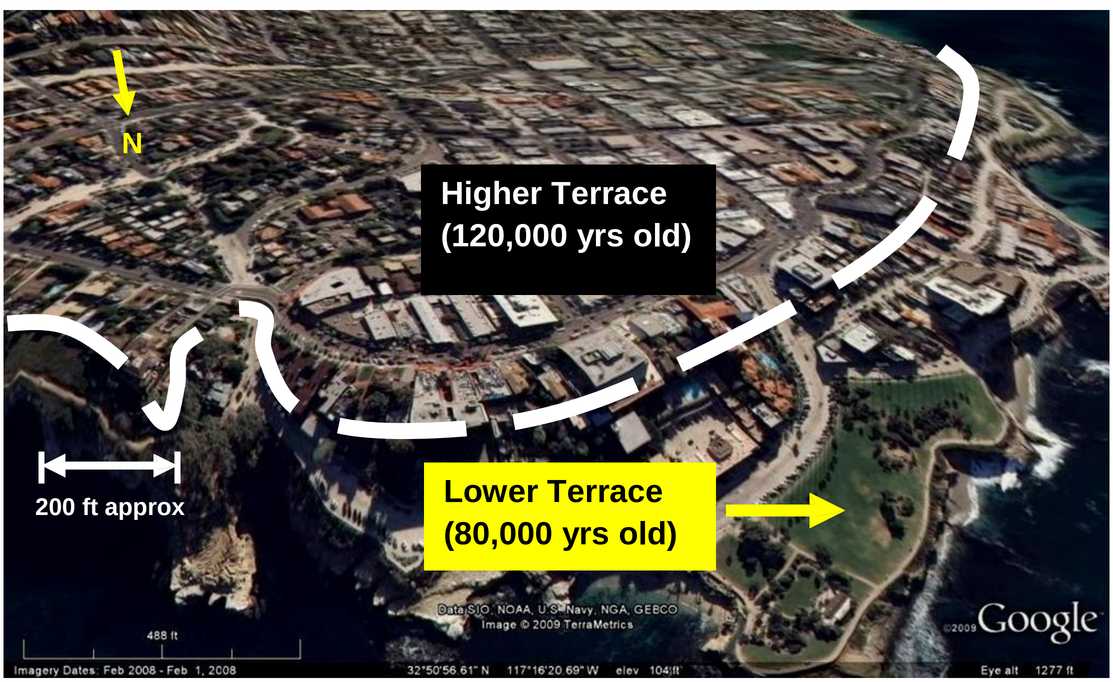
Wavecut platform, La Jolla coastline.

Stack—an erosional remnant of a former headland, created as the coast retreats inland due to wave erosion, undercutting, and mass wasting:



Sea stacks near La Jolla Cove.

Another important coastal feature that dominates metropolitan San Diego is called a marine terrace. A marine terrace is simply an uplifted wavecut platform upon which sediment is deposited. The town of La Jolla is built on two marine terraces, including a lower terrace dated at 80,000 years old and a higher terrace dated at 120,000 years old, as shown in the image below:



Marine terraces in La Jolla (image courtesy Google Earth).

In the above image, the grassy park next to La Jolla Cove is built on the lower (younger) terrace, whereas most of La Jolla is built on the higher (older) terrace.

As you've probably noticed, most of coastal San Diego is relatively flat. In many places you can recognize a stair-step topography. This is because San Diego is built upon a series of uplifted marine terraces.

Given that all marine terraces start out as wavecut platforms at or just below sea level, there's a relationship between terrace age and elevation. Higher marine terraces like the one Mesa College is built on are relatively old, whereas low-elevation terraces like those in coastal La Jolla are relatively young.

The age of the Linda Vista terrace, upon which Mesa College is built, is roughly 1.3 million years. In contrast, the age of the lower terrace at La Jolla Cove is only about 80 thousand years (Abbott 1999). Just think…the next time you visit Mesa College, you'll be walking on top of million-year-old sea floor! How cool is that?!

What determines whether a depositional or an erosional feature will develop along a particular stretch of coast? In a phrase, sediment supply vs. removal. If the supply of sediment along a stretch of coast is greater than the removal of sediment, a depositional feature will typically develop. However, if sediment is removed at a faster rate than new sediment is supplied, an erosional feature will probably develop.

An interesting implication of wave refraction and longshore drift is that an irregular coast will typically become straighter with time, because erosion (net removal of sediment) dominates at headlands, and deposition (net accumulation of sediment) dominates in bays and other coastal indentations.

Here's an animation that demonstrates this: **Coastal Straightening—Fig.** (scroll down to bottom of web page).

At headlands, incoming waves are refracted in such a way that wave energy is focused, or concentrated at the headland, which is attacked on all sides. In addition, incomplete refraction sets up longshore currents that tend to transport sediment away from the headland, resulting in erosion.

At coastal indentations such as bays, wave energy is dissipated, or lessened, within the bay because the energy is focused outward, toward the sides of the bay due to incomplete refraction.

Waves that reach the middle of the bay, furthest from the headlands on either side, don't typically possess sufficient energy to cause erosion, and so deposition results.

Erosion at headlands and deposition within bays eventually results in a straighter coastline.

Here in San Diego, for example, the coast around La Jolla is very irregular in shape, whereas to the north and south, the coast is straighter.

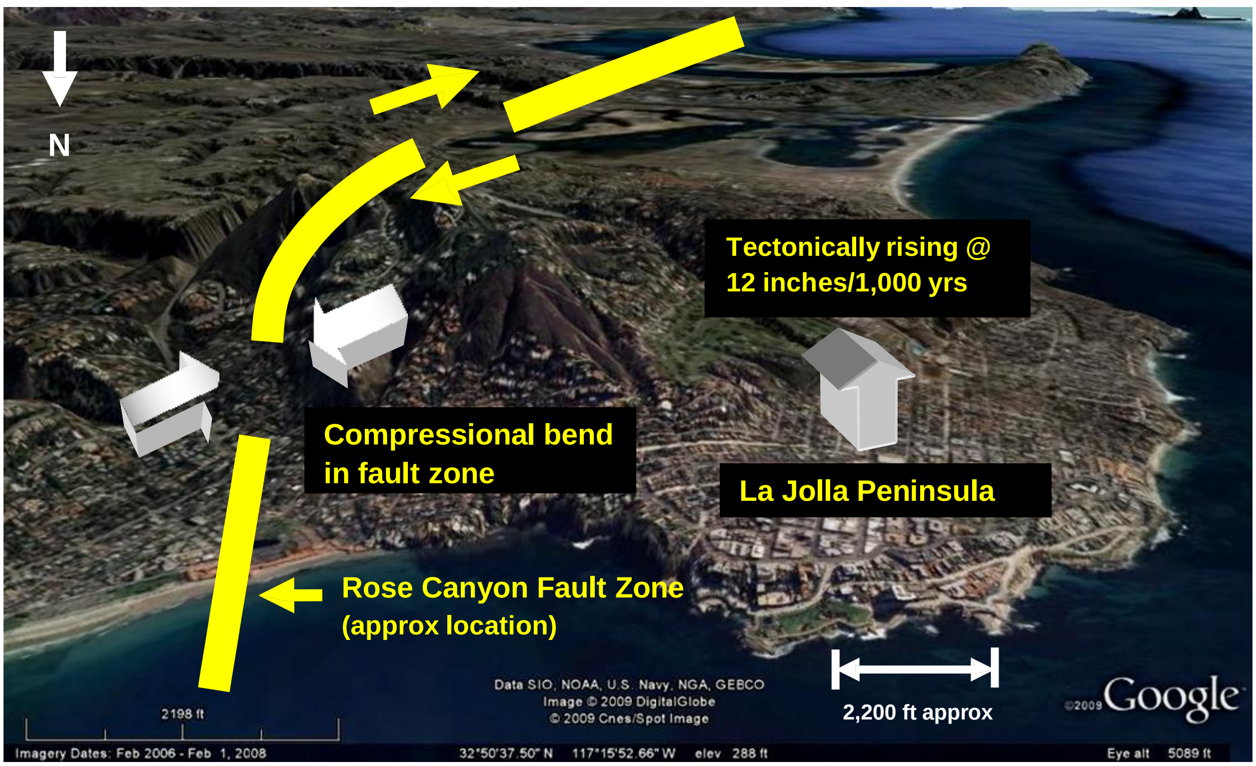
There’s a reason for this: in the La Jolla area, the coast is tectonically rising—it’s being shoved vertically upward along a compressional bend in the Rose Canyon Fault Zone.

This means that subsurface rock is constantly being lifted up (albeit very slowly from a human perspective), out of the water, before coastal straightening can modify the shape of the rising shoreline.

With continued uplift, the still-irregular coast is uplifted above sea level and quickly isolated from further coastal straightening as it rises skyward.

In contrast, the land north and south of La Jolla isn’t rising as rapidly (or at all in some spots); therefore, more time is available to straighten things out.

Here’s an oblique image of the La Jolla area (below). The Rose Canyon Fault Zone trends onshore at the south end of La Jolla Shores Beach, curves around the eastern side of Mt. Soledad, follows I-5 Freeway south, slices through North Island, and dies out in San Diego Bay:



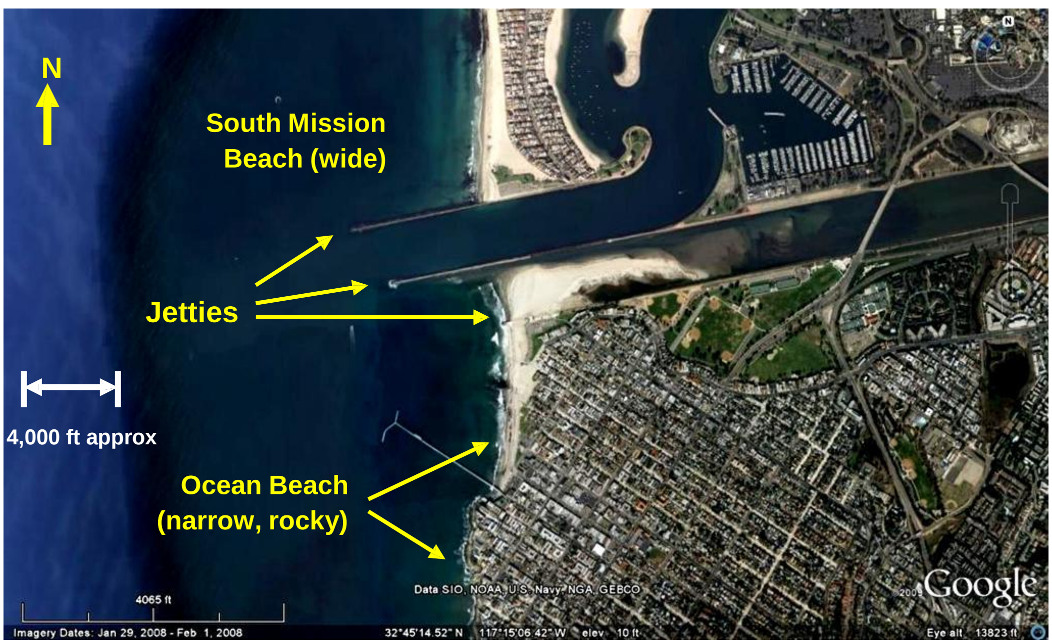
Approximate location of Rose Canyon Fault Zone, La Jolla, CA.
(image courtesy of Google Earth)

This same idea applies at larger scales as well. Along the Pacific coast, we’re at or very near an active plate boundary, and so much of the land is tectonically rising due to crustal compression and shortening. Coastal straightening processes therefore don’t have a long time to act, and so our rising coast is quite irregular in many spots. Along the Gulf and East coasts, which aren’t near a plate boundary, the land is rising very slowly or not at all in many spots, and so coastal straightening processes have more time to act. So plate tectonic processes even influence coasts!

One last topic from this chapter I'd like you to read about falls under the heading, Human Interference with Sand Drift. In summary, we've learned the hard way that it's not always a good idea to interfere with natural, longshore sediment drift. When we do, unanticipated consequences often result. For example, what happens when we build coastal structures like jetties and groins that jut out from the coast, interfering with longshore drift? As you can see in Figure below, there's often a buildup of sediment on the upcurrent side of the structure. On the downcurrent side of such coastal structures, sediment supply is cut off, and erosion may result.

Coastal erosion occurred here in San Diego following construction of the Mission Bay jetties. Mission Beach is built on a sand spit that separates the Pacific Ocean from Mission Bay. Following dredging of the Bay in the 1940s for recreational purposes, jetties were built to prevent the entrance channel to Mission Bay from filling in with sediment derived from longshore drift and the San Diego River.

The jetties block the longshore drift in this area, which is toward the south, and sediment builds up against the upstream side of the northern jetty. This, in part, is what makes south Mission Beach so wide. On the downstream side of the jetties, however, the jetties cut off sediment supply to the south, and as a result the cliffs at Ocean Beach have receded considerably (Kuhn and Shepard 1984):



Mission Bay jetties, San Diego, CA
(image courtesy of Google Earth)

Kuhn and Shepard (1984) report that since the construction of the Mission Bay jetties in the late 1940s, Ocean Beach has had to be maintained by dredging because sediment supply from other sources (e.g., longshore drift and the San Diego River) is insufficient. Other causes of coastal erosion include surface water runoff, pedestrian traffic, overwatering of bluff tops, and animal burrowing.

Here's an animation that nicely demonstrates this: **Human Interference with Sand Drift—Figure 20.12a.**

A controversial topic here in California has to do with the benefits vs. risks of seawalls—coastal structures designed to prevent sea cliffs from eroding. If you own a $2 million dollar home atop a steep sea cliff that overlooks a wonderful stretch of beach, you'd probably be inclined to build a sea wall to armor the unstable cliff below your home, so coastal erosion, mass wasting, and cliff retreat don't threaten your investment.

The problem arises, however, in the effects of the seawall, which include (1) passive erosion, (2) down-current sediment loss, (3) placement loss, and (4) access loss (Griggs et. al 2005).

Passive erosion results as the coast retreats (erodes landward) on either side of the seawall. This results in the creation of a sort of headland in the vicinity of the seawall, which focuses wave energy, and results in the shrinking of the beach in front of and along the sides of the seawall. The beach shrinks in front of the seawall not so much because waves bounce off the seawall and erode the beach on their way back out to sea, but because the seawall itself blocks new sediment from being added to the beach, while longshore currents remove existing beach sand.

Down-current sediment loss results because sediment from the sea cliff that might otherwise be added to the longshore drift is trapped behind the seawall.

The construction of a seawall can also result in placement loss, which simply means that the seawall itself takes up some of the area formerly occupied by the beach.

Finally, some seawalls are constructed in such a way as to restrict public access to the beach below. California beaches are public property (up to the high tide line), and a valuable public resource. Is it really fair, many would ask, to protect private property at the expense of public resources?

One way to protect the coastline is to establish a coastal setback zone where no new construction is allowed. That way, the sea cliffs can erode as they're supposed to, and the eroded material keeps the beach below from disappearing.

Some towns and cities in California and in other states are already doing this. In much of California, however, cliff-top development is so extensive that establishing coastal setback zones is impractical.

Another option is for coastal conservation organizations to purchase clifftop property when it comes up for sale; however, this is very costly. A third option is to replenish beaches by artificially adding sand (very costly and may not work). Damming virtually all of the river systems in California only contributes to the problem of beach disappearance, because dams not only block water, they also block the sediment that would otherwise be flushed into the coastal zone at river mouths. In summary, beach erosion is a big problem, and solutions are controversial and costly.

References Cited

Abbott, Patrick L., 1999. The Rise and Fall of San Diego: 150 Million Years of History Recorded in Sedimentary Rocks. San Diego, Sunbelt Publications, 231 p (http://www.sunbeltbook.com/BookDetails.asp?id=116).

Griggs, Gary, Patsch, Kiki, and Savoy, Lauret. 2005. *Living with the Changing California Coast.* University of California Press. November. 551 pages (http://www.ucpress.edu/books/pages/10203.php).

Kuhn, Gerald D. and Sherpard, Francis P. 1984. *Sea Cliffs, Beaches, and Coastal Valleys of San Diego County: Some Amazing Histories and Some Horrifying Implications. University of California Press (http://www.escholarship.org/editions/view?docId=ft0h4nb01z;brand=eschol).*