Wilson Cycle Description

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Layout of this Guide

For each stage represented in this Wilson Cycle, several pieces of information are provided:

Description of Process: a description of the forces acting on a particular location, and the results of those forces;

Composition: the particular rock types that result from these forces, including chemical make up, major minerals present, and texture (grain size and orientation), expressed in photographs, verbal descriptions, and categorical graphs;

Specific Locations, Present and Past: the map locations of examples of the environment, with current and past locations represented.

Stage A: Stable Continental Craton

Description of Process: This stage represents the basis of all existing continents, showing in a simplified fashion both continental crust and adjacent oceanic crust, both of which lie over the mantle. You will notice that the continental crust is considerably thicker than the ocean crust. Continental crust is less dense than oceanic crust (approximately 2.7 g/cm^3 vs. 3.3 g/cm^3), and so "rides" higher over the mantle. Continental crust is also more rigid that oceanic crust. Furthermore, the development of continental crust involves considerable thickening.

Composition: This diagram represents considerable simplification, in that the internal structure has not been shown, but is represented as a crystalline "basement" for the continent. This basement material represents a variety of rock types, including granites and diorites, gneisses and schists, and various sedimentary rocks. Taken as a whole, the average composition of the basement rock is the same as a granodiorite – enriched in sodium, potassium and silica-rich minerals, such as alkali feldspar, sodium plagioclase, and quartz, and depleted in iron/magnesium, silica poor minerals, such as olivine, pyroxene, and amphibole.

Locations: The majority of the continent shows very little relief, with the high spots having been worn down by erosion, creating a peneplain. But while the vertical dimension is shown proportionally to ocean crustal thickness, there is relatively little information on the horizontal dimension. The continents as we know them today are actually collections of smaller continental crust blocks, all of which are undergoing erosional processes. Current examples of large, stable continental masses that are eroded down to basement include much of Africa and nearly all of Australia.

Stage B: Initial rifting

Description of Process, Stage B begins when plumes of hot, low density material rise up through the mantle toward the surface. Mantle plumes appear randomly scattered around the Earth and produce features such as the Hawaiian Islands and the Yellowstone hot spot. It is not clear why, but sometimes such plumes initiate a rifting; Stages B, C, and D together represent one of these complete rifting events and the opening half of the Wilson cycle.

When a mantle plume rises under a continent, large pools of hot magma accumulate at the base of the continent. The continent acts like a heavy blanket holding the magma down - but only for a short time. Heat from the magma pool transfers to the continent, causing it to expand and swell upwards several kilometers to form a hot spot - a circular bulge about a thousand kilometers in diameter. At depth the continental rocks are hot and plastic, and stretch and thin, much silly putty being pulled apart. At the continental surface, however, the rocks are cold and brittle and crack to form normal faults as the continent stretches during its upward expansion. The result is a jumble of crustal blocks that fall and rotate downward into the spaces created by the stretching, thinning continent. The blocks that fall the farthest form graben valleys, while adjacent blocks that do not fall as far form horst mountains. Typically, a large graben forms in the axis, the place where ocean lithosphere will begin to form in Stage C.

Meanwhile the faulting opens conduits along which magma works its way toward the surface. Much of this magma is mafic (basaltic) from the mantle plume, and its spills onto the surface through fissure volcanoes forming flood lavas, and a plethora of small cinder cones. However, other magmas result from the partial melting of the lower regions of the continent, and form pink granite batholiths which give rise to large rhyolitic volcanoes.

As soon as uplift begins, weathering and erosion attack the horst mountains. Coarse-grained breccias, conglomerates, and sands form large alluvial fans and braided rivers along the graben edges. Toward the center of the graben, however, lakes commonly form since the graben blocks are falling down, giving no outlet for the rivers to the sea; these fill with muds. In time the grabens will fill completely with sediment and have little surface expression, but at this stage the area looks like a broad valley a few 10s of km wide surrounded by high, rugged mountains.

Composition:

Igneous: One of the key ideas behind the Wilson cycle is that rocks gneous, sedimentary, and metamorphic volve. If they did not then all the great variety of rocks on Earth would not be possible. The basic idea is that when the Earth outer mantle and lithosphere originally formed over 4 billion years ago their composition was mostly a homogeneous chemical mix called the parent rock. In hand specimen, the parent rock looks like an iron-rich, silica-poor mafic or ultramafic igneous rock, such as an olivine basalt. The parent rock, however, has smaller quantities of all the other elements necessary to form all the other rocks; all we have to do is to separate and concentrate these elements.

Fractionation is the process by which elements or other constituents are separated from each other based on differences in size, weight, valence, reactivity, etc. What this means is that a rock of mixed composition is divided into one or more fractions that concentrate those compositions. For example, if an igneous rock is heated slowly, the cooler forming minerals closer to the bottom of Bowen Reaction Series will melt before the hotter forming ones higher up. Once approximately 20-30% of the rock is melted, the molten material, which is less dense than the unmelted rock, begins to rise through the continent along cracks and fractures, separating this fraction from the denser, unmelted portion of the rock, which remains behind. Alternatively, the first crystals to form as a magma cools can sink, being denser than the remaining magma. These then accumulate on the bottom of a magma chamber, separating this fraction from the rest of the magma. In igneous rocks, the molten fraction always has a composition lower in iron, magnesium, and calcium, and richer in silica, sodium, and potassium, lower in the reaction series than the original rock that was melted to began with. The unmelted residue always has a composition higher in the reaction series than the rock we started with.

In Stage B the magma that has ponded at the base of the continent is the mafic/ultramafic parent rock. It cools slowly resulting in a melt that is mafic in composition (iron and calcium rich; i.e. pyroxene and calcium plagioclase), and an unmelted residue (iron and magnesium rich; i.e. olivine and pyroxene) that is ultramafic in composition. The unmelted residue stays down and eventually crystallizes to form ultramafic rocks. This is the first fractionation processes taking place at Stage B.

A second fractionation process involves fractional melting of the lower portion of the continent. The average composition of a continent is a granodiorite, an intermediate rock in the middle of Bowen reaction series. The heat from the mantle plume conducts to the continental rock, fractionally melting it. The melted fraction is a silica and potassium-rich/iron-poor mixture that, because it is hot and of lower density, rises up through the crust as blobs to form granite batholiths. Some of it may rise all the way to the surface and form large, rhyolitic volcanoes.

Sedimentary: Sedimentary rocks, like igneous rocks, evolve via fractionation. But in the case of sediments and sedimentary rocks, the fractionation process is driven by weathering and sorting. When a parent rock, such as a granite, first begins to weather abundant feldspar and quartz grains are released, as these are the primary minerals in a granite. The resulting sediments are feldspar and quartz rich, called arkosic. These are immature sands that have been deposited before feldspars can weather into clay and grains rounded and sorted. Quartz and feldspar do not weather equally well, however. As the feldspar changes into clay, quartz is left behind, as it barely weathers at all. So, if we start off with an arkose with lots of feldspar and quartz, and continue to weather it, feldspar gradually disappears, until there is none left, while quartz, because it does not weather, becomes more and more what the sandstone is made of. A sandstone made only of quartz is said to be mature because weathering can no longer affect it.

The particles that compose a sandstone (or breccia or conglomerate) are of three kinds: quartz, feldspar, and lithic (rock) fragments. A lithic fragment is a general term for a rock fragment that is anything but quartz or feldspar, and may consist of minerals, or unweathered igneous, sedimentary or metamorphic rock fragments. Lithics, like feldspar, are unstable and eventually weather completely to clay (or dissolved minerals). In Stage B lithic particles are not that common, but they will be abundant in other stages.

An easy way to visualize sedimentary rock evolution is with a ternary diagram, which is a graph with 3 instead of 2 axes. Data are plotted on this diagram in relative percentages, with quartz at the top apex, and feldspar and lithics at the lower two apexes. Rocks with lots of feldspar and/or lithics are said to be immature, meaning they have barely begun their evolution. Near the top but not at the top, when feldspar and/or lithics are between 10 and 25 percent, the rock is submature; well evolved, but not completely. Rocks with over 90 percent quartz are mature. Sourcelands may produce sediments of many compositions, and they can all be plotted appropriately on the ternary diagram, but one thing is true about evolution on all ternary diagrams; no matter where you start, the evolution of sedimentary rocks is always in the direction of the quartz apex.

Finally, sediment maturity is also measured by particle size and shape; a mature sediment by definition must be composed of all sand sized particles, rounded and well sorted. Breccias and conglomerates, having much larger, more assorted-sized grains, are immature. Their maturity increases as the sediment travels down stream and the larger particles are left behind by sorting.

The relationships between sediment composition and the tectonic conditions under which they form are important clues to interpreting geologic events and history. Figure 2 summarizes the work of generations of geologists studying rocks from all over the world under many tectonic conditions. Each of the four fields in the ternary diagram is a part of one or more stages in the Wilson cycle.

Thus in Stage B, where there are high horst mountains adjacent to deep graben basins, sediment does not enough time to evolve, at all. Poorly sorted breccias and conglomerates are deposited rapidly at the base of the horst and the sorting and rounding process is cut short. Even further out where there is sand, the sediment is feldspar rich, poorly sorted and less angular. Because of the rapid burial, feldspars do not have time to weather out; the rock may have as much or more feldspar as quartz. Ancient deposits like this tend to show up in field B in the ternary diagram (Figure 3 - QFL diagram.). However, as the horsts erode down and less feldspar is supplied by weathering and erosion the composition of the sediment evolves toward the Quartz apex (arrow in Figure 3).

Stage C Rift to Drift: Early ocean basin formation

Description of Process: Stage C follows seamlessly from Stage B, in that the original continent stretches, thins, and eventually splits in two to form two continents (Eastcontinent and Westcontinent). Between the two continents a new ocean basin forms, at first only a few tens of kilometers wide, but eventually opening to a full ocean thousands of kilometers wide. This is called a ift to drift sequence in that the original continent rifts into two pieces, and the pieces then drift apart as the ocean basin widens (this is a carry over of Alfred Wegener ontinental drift terminology, even though we know continents do not rift in the way he described it).

The final splitting of the continent (rift-phase) occurs when rising magma forms a series of vertical dikes, sheeted within each other like reams of paper. Dike after dike after dike injects into the thinned continental crust until one cannot tell if they are looking at continental rocks overwhelmed by basaltic dikes, or oceanic crust contaminated by continental rocks. This zone is sometimes referred to as the transition zone. Transition refers to the fact that this area has seismic velocities between slower continental rocks and faster oceanic rocks, but can also be interpreted as transitional between pure continental crust and pure oceanic crust. With the final continental split, the active zone of rifting or axial graben, shifts from under the Stage B continent to the center of the new Stage C and D ocean basin. A limited amount of lateral tectonic force, called idge-push begins to operate, further driving the continental plates apart, allowing more new ocean crust to form at the ridge.

The new ocean crust forms by partial melting and rising of mantle material below the active rift. Below the surface, in the magma chambers, the slow cooling leads to the formation of coarse-grained gabbro. At the surface, the lava spills onto the ocean floor, into the cold ocean water, where it quickly chills to form large balls and blobs of lava pillow basalts. Below the region of fractional melting, about seven kilometers below the ocean floor, the crust ends and the mantle begins, at the seismic velocity boundary known as the Moho, named for the Croatian seismologist Andrija Mohorovićič. Seismic waves speed up noticeably at this boundary, indicating an increase in the density of the rocks; their composition is ultramafic, or their metamorphic equivalents (e.g. ecologite).

Meanwhile, the two new continental margins, one each for East- and Westcontinents, separate, move apart, and begin to subside (the drift-phase). During Stage B these regions were raised up by the heat from the mantle plume, but since that source of heat is now located in the middle of the expanding ocean basin the new continental margins cool, become denser, and sink. Within only a few million years they subside below sea level and are covered by water flooding the now low-lying area. The submerging areas, containing remnants of the rift-phase horsts and graben with their sediment and volcanic fills, are soon buried under a drift-phase wedge of sediments called the Divergent Continental Margin (DCM). The environments range from coastal, near-shore marine environments to shallow shelf sediments off shore. If the climate permits, limestones also deposit. The cumulative weight of these sediments accelerates the sinking of the continental margins; subsidence and shallow water deposition will continue for the next 100 million years.

Composition:

Igneous: The material directly beneath the rift, in the upper part of the mantle but below the Moho, is primarily one of three ultramafic rocks: dunite=olivine, pyroxenite=pyroxene, or peridotite=olivine+pyroxene. This rock is very rich in iron, magnesium, and calcium, but the silica and aluminum content is relatively low. Partial melting is facilitated by the lower confining pressure under the rift, which lowers the melting point of the rock. Molten material rising and then cooling without reaching the surface forms gabbros, rocks that are still rich in iron, magnesium, and calcium, but are also richer in silica and aluminum. Gabbros are composed of coarse grains of pyroxene and calcium plagioclase. Material that reaches the surface retains the same chemical composition, but the rapid cooling produces fine-grained basalts.

Sedimentary: Distinctive shifts take place in the nature of sediments deposited from a rift to drift transition. During the rift phase (Stage B), the active formation of horsts and graben, sediments are feldspar-rich (that is, they are arkoses). This is because the horsts are uplifted continental rock, and continents are granitic in composition and thus feldspar-rich. As long as the horsts are active, the eroding sediments are feldspar-rich. But, as erosion wears down the horsts, feldspar is chemically weathered out, and the bulk compositions shifts increasingly toward quartz, the one mineral that does not weather out.

Thus, by the time we reach the drift-phase margins of the continents, the sediments that are deposited on the shores of the new ocean basin are primarily quartz sands, forming a thick blanket that marks the rise of sea level along this coastline. Deeper shelf water along the DCM is characterized by siltstones and shales, formed from sediments carried further out into the ocean basin as a result of the finer grain size. In some cases the sediments are dominated by shallow water limestones, particularly in tropical areas.

Locations: Examples of margins at the Stage C rift-phase are found along the Red Sea and the Gulf of Aden between northeast Africa and Arabia. Active rifting is forming new ocean crust down the center of these basins, but the ocean basin has not begun to open yet. Alluvial fan and braided river environments are actively depositing along the sides of the horsts. The axial graben has subsided below sea level, where black shales are accumulating. These areas have not begun the drift-phase, and a DCM has not yet begun to form; they are probably only a few million years away, however.

Ancient examples of rift-phase deposits are often hard to find. They are still buried under tens of thousands of feet of DCM sediments, or have been deformed, metamorphosed, and more or less eroded by later events. However, examples of rift-phase sediments from the 650-700 Ma rifting of Rodinia and 270 Ma rifting of Pangaea are found all along the eastern North American margin. These include the Rodinia-aged Ocoee sands, Mt. Rogers volcanic, and Lynchburg graben sediments. On top of these sediments are early drift-phase deposits, including thick Cambrian-aged Antietam quartzite (formerly sandstone) and (meta) siltstones. These are found on the west flank of the Blue Ridge, running from Pennsylvania to Georgia.

Other ancient examples are the sediments of the upper layers of the Pangaea-aged Hartford, Newark, Culpepper, Danville, and Deep River basins, running from New England down through the Carolinas. These contain coarse, immature sands, lake muds, and volcanic ash and lava deposits.

Stage D: Full Ocean Basin

Description of Process: Stage D flows seamlessly from Stages B and C. The evolving DCM continues to subside while sediments eroding from the continent are deposited along a broad, slowly deepening shelf. The coastal regions are dominated by sandy environments (beaches, barrier islands), but they grade offshore into shelf muds, and (if the climate is tropical/ subtropical) limestones before transitioning into the continental slope and rise where water depth increases rapidly toward the ocean floor. Because crustal cooling and sediment accumulation continues along the DCM, subsidence also continues - although ever more slowly with time. But, because subsidence is greatest toward the most distant, rifted edge of the continent the sediments thicken here to as much as 10 km or more. Inland toward the continental interior the DCM sediments thin to a feather edge, and may even disappear completely. Overall, however, the DCM is a geologically quiet place, even boring, especially compared to rifting and mountain building events.

Meanwhile, new oceanic crust continues to form along the axial center of the widening ocean basin. Fractional crystallization of magma below the axial rift produces deep ultramafic plutons, gabbros and sheeted dikes, and finally pillow lavas on eruption at the sea floor. There is still upward bulging of the crust due to heat flow (creating the mid-oceanic ridge or rise). This bulge is mostly below sea level (with exceptions like Iceland), but the water here is shallower than most of the ocean basins. Within the bulge, there is ongoing graben formation along the ocean ridge axis.

Shallow focus earthquakes, fault movement, magma flow, and fractures are common along the ridge axis. Sea water infiltrates the crust along these fractures, comes in contact with heated rocks, and flows back upwards and outwards toward the surface. Water under pressure and at high temperatures is quite active chemically, dissolving metals, sulfur, and carbon dioxide. When this water flows outwards and upwards "black smoker" and "white smoker" geothermal vents (underwater geysers) form. These are tall, hollow chimney-like structures, spewing hot mineral laden water into colder, less acidic sea water leading to mineral precipitation that causes the chimney to grow even taller. Temperatures reach 300 degrees C, yet the water does not boil - it is under much higher pressure on the sea floor. Surrounding the geothermal vents is an exotic community of blood red tube worms, clams, fish, and a host of other animals, all sustained not by energy from the sun, but by bacterial communities living in the vents, deriving energy through chemosynthesis from breaking apart H2S.

When we view a modern ridge axis (via remote cameras or research submersibles) it resembles an underwater moonscape; tumbled lava blocks, pillow basalts, shear cliffs and steep chasms. Young crust, recently cooled from lava, has not been around long enough to accumulate sediments. Moving away from the ridge axis, however, this rugged landscape is buried under a progressively thicker layer of fine-grained mud or ooze—pelagic sediment. At the ridges this sediment is derived primarily from the skeletons of plankton or microscopic organisms, including foraminifera (calcium carbonate) and radiolaria (silica). Away from the ridges more and more clays compose this sediment. Finally at a thousand or so km away from the ridge we reach the abyssal plain, a seemingly endless, featureless, flat expanse of sediments.

Composition:

Igneous: Oceanic lithosphere is composed of 4 layers of rock about 10 km thick called the ophiolite suite, all–except for the sedimentary layer on top–generated from magma brought upward by convection cells.

The deepest ophiolite layer is in the asthenosphere just below the Moho. On the cross sections the base of the black ocean crust is the Moho so except for the red batholith at the rift center the deeper asthenosphere layer is not detailed. But, it is in that batholith that the mafic/ultramafic parent rock starts the fractionation process that generates the ophiolite suite.

The parent material coming up from deeper in the mantle is a very viscous plastic, which melts at the lower pressures near the surface. It then fractionally crystallizes during cooling. The first minerals to crystallize out are the heavy olivine and pyroxene, which settle downward and concentrate in layers resulting in the olivine/pyroxene rich, feldspar poor ultramafic rocks such as dunite, pyroxenite, and peridotite. The lower specific gravity melt floats upward above the Moho to form a layer of gabbro, and above that sheeted basaltic dikes (analogous with the sheeted dikes of Stage C.) Finally, feeder dikes take some of the remaining magma to the surface where they form the basaltic pillow lavas on the ocean floor. Associated with the pillow lavas are white and black smoker deposits generated when sea water percolates down, is superheated, and then returns to the surface as underwater geysers. These are rich in hydrothermal metal sulfide minerals such as pyrite, galena, and covellite; most of the important copper reserves in the world also come from ancient examples of these sites. Often the ophiolite rocks are hydrothermally metamorphosed to soapstone and serpentinite.

Sediments: Divergent Continental Margins are characterized by mature (quartz rich) sandstones and muds. For example, the east and gulf coasts of North America with their beach and barrier island systems, and shelf muds are a good example. However, the most distinctive DCM sedimentary records are vast, very thick accumulations of carbonate rocks (limestones and dolomites). These are climate sensitive and require tropical or subtropical conditions and a complete absence of clastic sediments (sandstones/shales). The necessary conditions are not that common on Earth today (one exception is the Great Barrier reef complex of Australia), but vast DCM carbonate deposits are quite common in the geologic record.

Ocean floor sediments are mostly pelagic (very fine grained suspended sediments and tiny skeletons), and have distinctive distributions. For example, ridge axes are dominated by calcareous (calcite) or siliceous (silica) bearing skeletons of single celled organisms; e.g. foraminifera, radiolarian, diatoms. Moving away from the ridge axis, calcareous skeletons decline (because they dissolve back into the cold, deep water below the carbonate compensation depth) and pelagic muds and siliceous skeletons increase. The deepest muds on the abyssal floor are derived from volcanic ash, wind blown sediments off of the world’s major deserts, and (strangely enough) the burned up remains of meteorites. It takes millions of years for these sediments, raining down into the ocean and sinking, to create the thick deposits we now see.

<u>Locations:</u> In the modern world, the Atlantic Ocean Basin and parts of the Indian Ocean Basin are surrounded by DCMs. Running down the center of these oceans are the axial rifts (e.g. mid-Atlantic ridge), the site of sea floor spreading. These rifts zones are also known as mid-oceanic rises, rifts, ridges, and divergent plate boundaries.

Ancient examples of DCM sediment wedges are the thick carbonate sediments that compose the Shenandoah Valley of Virginia and the Great Valleys of Maryland and Pennsylvania. These have been faulted and folded since then to form the Appalachian Valley and Ridge province. Many other DCM successions are known but have been more or less metamorphosed, fragmented and dismembered by younger tectonic events making it harder to pinpoint them easily on a map.

Stage E: Island Arc (a.k.a. Volcanic Arc) Development and Change

Description of Process: If we assume that the Earth remains the same size, then as new crust is created at divergent plate boundaries (Stages B, C, D), old crust must be consumed somewhere else at convergent plate boundaries. The convergence in Stage E is being driven by divergence some place, but the source of that divergence is not specified in this model. It could be within the ocean basin created in Stages B, C, and D, but, since that ocean basin is completely closed by Stages H and I, the driving force is more likely to be outside this model, someplace else on the Earth. One way or the other Stage E begins the closing half of the Wilson cycle. The model closes the basin in two distinct stages (E, F, and G, H).

Since the surface of the Earth is roughly 70% oceanic crust, one of the most common places convergence can occur is through the interaction of two oceanic lithospheric plates. Where this convergence occurs, older, colder crustal material tends to be of higher density, plus being weighed down by any sediments sitting on it. As a result, this material slides underneath younger or less dense oceanic material. The resulting feature is called a subduction zone, as one plate subducts (descends) beneath the other, dipping into the Earth at an angle of about 20-40 degrees. Where the one slab descends, a deep ocean trench forms, as the conveyor-like movement drags the surface down. At about 200 km beyond the trench the descending slab reaches 100-150 km deep, a zone where fractional melting begins to generate magma with an intermediate composition.

The processes of subduction and fractionation are facilitated by seawater, which along with the wet oceanic sediments, forms a type of lubrication. This lubrication is not perfect, however, as the overlying plate scrapes this sediment off the subducting plate, forming an accretionary wedge of folded and faulted sediments. Earthquakes occur all along the descending slab (which is how we know their structure and behavior). The foci for these earthquakes increase with depth as one travels across the subduction zone and toward the volcanic arc. Water also acts as a flux, lowering the melting point of the subducting slab and overlying mantle rocks resulting in the generation of intermediate magma. The volcanoes that result are typically the highly explosive composite type (a.k.a. stratocone). From a satellite view the volcanoes forming above the subducting slab string out in a curvilinear arc, hence “volcanic arc.”
The rising magma emplaces itself into the lower crust as batholiths, forming the core of the volcano. If enough batholiths accumulate, they merge to form a small block of relatively light weight rock that floats on the underlying mantle—a microcontinent. As the batholiths cool they conduct heat into the surrounding oceanic lithosphere, leading to regional (Barrovian) metamorphism.

The new island arc also produces sediments eroded from the arc and deposited on the flanks of the volcano; some ends up in the trench and gets subducted. The frequent earthquakes associated with this convergence triggers massive submarine turbidity flows (slurries of water and sediment that flow down slopes at high velocities), producing characteristic deposits of sediments. The sediments are “immature” meaning they contain abundant unweathered, mostly volcanic rock fragments; which tells us they were eroded and buried relatively quickly.

On the trench side of the volcanic arc sediment eroded from the volcano, as well as ocean floor sediments, tend to get dragged down into the Earth by the descending slab; the descending slab acting like a conveyor belt. As the oceanic slab descends some of the sediments get scraped off, and stacks and piles up as a disordered, folded, faulted, sheared mess of material called the accretionary wedge (or the French word melánge=mixture.) Accompanying this is a type of metamorphism called “blueschist,” in that the minerals that make up the rocks are flakey in appearance with a noticeable bluish cast.

Composition:

Igneous: Magmas produced by the fractional melting of subducted ocean crust and mantle are richer in silica, sodium, and aluminum and poorer in iron, magnesium, and calcium than the mafic/ultramafic rocks from which they were fractionated. The result is calcium-sodium plagioclase feldspars and amphibole minerals forming coarse grained diorites (plutonic) and fine-grained andesites (volcanic), although granodiorites, white granites (=plagiogranites) and rhyolites are also common. The magmas and lavas that produce these intermediate rocks are of lower temperature and higher viscosity than the mafic rocks from which they were derived. As a result their volcanoes tend to be explosively erupting. Well known historic eruptions such as Tambora, Santorini, Vesuvius, and Pinitubo are of this type.

Sedimentary: Sediments derived from the arc volcanoes consist of several materials, including fine-grained volcanic ash deposits, fine-grained volcanic muds, and often times, larger pyroclastic fragments resulting from explosive eruptions. Weathering of the volcano produces lithic-rich, quartz-poor, very immature sediments. Commonly they have 90% or more rock fragments. Once the volcano is extinct and erodes away it may expose some of the deeper plutonic rocks, which are feldspar and sometimes quartz rich. Thus, the sediments evolve in time toward sediments with more feldspars and quartz (Figure 4).

Metamorphic: Minerals and rocks are stable only under the conditions at which they form; change the conditions and they must change and adapt to the new conditions. That is, no rock is accidental. Weathering is a major class of such changes; igneous minerals formed at high temperatures are unstable at the Earth's surface and weather to clay which is stable at the Earth's surface. Metamorphism is a second class of adapting to new conditions, and is a rock's response to heat and/or increasing pressure. Heat increases either with increasing depth or with proximity to magma intrusions. Pressure increases with depth because of the weight of overlying rock (lithostatic pressure), or may increase when, for example, two continents collide, squeezing rocks in between (directed pressure).

The "classic" work on metamorphic rocks was done by George Barrow, working with thesequence of metamorphic rocks seen in Scotland. Five kinds of metamorphism exist depending on different combinations of temperature, pressure, and water abundance.

1. *Barrovian (a.k.a. "Regional"):* moderate pressure, intermediate to high temperature
2. *Blueschist:* high pressure, low temperature
3. *Eclogite:* high pressure, high temperature
4. *Hydrothermal:* low pressure, low temperature, high water
5. *Contact:* low pressure, high temperature, low water

All five of these occur in a volcanic arc system, but Barrovian, blueschist, and eclogite are the most widespread (regional in distribution).

In metamorphism minerals recrystallize without melting, resulting in new minerals that are stable at elevated temperatures and/or pressures. Metamorphic rocks are described by their grain orientation and size, and the extent to which minerals are segregated. In order to cope with the changing pressure-temperature conditions, the minerals in the rocks begin to recrystallize, while the mineral grains align themselves perpendicular to the pressure, creating foliation. Each kind of metamorphism occurs in discrete regions or zones that can be mapped, with low pressure, low temperature regions closer to the surface and higher pressure, higher temperature regions at greater depth. Rocks in each zone contain characteristic minerals that are stable and only form when temperature and pressure conditions reach specific levels, making them an "index" of the pressure-temperature conditions. At the highest temperatures, minerals begin to melt, forming migmatites that are fractionated from the metamorphic rocks.

The other important variable in metamorphism is the composition of the parent rock. Any kind of rock—igneous, sedimentary, or metamorphic—can undergo any kind of metamorphism, so the variety of metamorphic rocks is potentially great. The same temperature and pressure conditions often results in very different looking rocks, based on the original, or protolith, rock. For the Wilson cycle model we present here, we are interested in only a few parents, mostly sedimentary (sandstones, shales, limestones) and igneous (mafic–basalt/gabbro–and ultramafic–peridotite).

Parent Rock	**Resulting Metamorphic Rock Type**		
	Greenschist Grade	***Amphibolite Grade***	***Granulite Grade***
Mafic (basalt/gabbro)	Greenschist rock	Amphibolite rock	Granulite rock
Shale sediment	Slate/Phyllite	Schist	Gneiss

A word of caution: *greenschist, amphibolite,* and *granulite* grade or facies are terms used as relative measures of temperature and pressure. There are also rocks named *greenschist, amphibolite*, and *granulite,* that are used for rocks derived from the metamorphism of basalts and island arc sediments in Stage E. One should always be mindful of whether one is discussing the type of metamorphic rock, or the extent to which a metamorphic rock has been recrystallized. An overall framework metamorphic conditions, grades, and types is found in Figure 5.

In Stage E, Barrovian metamorphism takes place when batholiths emplace into oceanic mafic parent rocks (i.e. basalts and gabbros), and the resulting metamorphic rocks reflect that original composition. At the lowest–*greenschist*–grade of metamorphism the metamorphic rock generated is called a greenschist, because the rock takes on a greenish cast and shows a flaky segregation of chlorite (a greenish mica) and epidote (a greenish-yellow mineral). At intermediate–*amphibolite*–grade, chlorite and epidote are unstable and recrystalize to form hornblende, plagioclase, and often red garnet. The new rock is amphibolite, a rock that appears as though made of needles of jet-black grains of hornblenede. At the deepest, highest temperature and pressure–*granulite*–grade, the amphibole and garnet minerals recrystallize to form pyroxene (which is dry), creating the rock granulite. At sufficient temperature, pockets of partially melted rock called migmatite can also form. Metamorphic rocks that are formed in these regions are often not seen until exposed by much later erosion.

The distinctive character of blueschist metamorphism is high pressure and low temperature, and these conditions only occur at the leading edge of the overriding plate, at the accretionary wedge. The low temperature is because the sediments descending down the subduction zone are cold–from being on the ocean floor–and the high pressure because the sediments are dragged deep into the Earth much faster than they can heat up. Blueschist rocks can contain glaucophane, epidote, jadeite.

The most exotic metamorphic rocks at this stage are eclogites, exotic because they require very high pressures and temperatures that only occur very deep in the Earth. Their parent is the olivine/pyroxene-rich ultramafic rocks, the unmelted residual portion of the subducting ocean plate. The resulting eclogite is a stunning looking rock composed of a pale green pyroxene and pink garnet (http://csmres.jmu.edu/geollab/Fichter/MetaRx/Rocks/Eclogite1.html) No other rock looks like it and to the trained eye it is unique, and special. Because they form so deep it takes the Earth some effort to bring eclogites to the surface; it happens in conditions like Stage H.

Because the magma that is rising to the surface beyond the subduction zone is literally forcing its way upwards, it rises very slowly. At depth, the rocks are fairly ductile, and deform around the rising magma. Because rock is a poor conductor of heat, the minerals in the surrounding, or country rock are altered, forming new minerals in zones surrounding the magma body. This is called contact metamorphism. Where subsurface water comes in contact with this hot rock, it is heated rapidly and can dissolve minerals in the country rock. Finding the fractures formed by the emplacement of the magma the path of least resistance, these hydrothermal (literally "hot water") fluids radiate outwards. As they cool, new minerals crystallize in veins. These veins are often the source of important mineral deposits, such as gold, silver, and copper.

Locations: There are numerous modern-day locations of Stage E. Active subduction zones with island arc formation include the Aleutian Islands, the Marianas Islands, and the Philippine Islands. Ancient examples, now attached to continental plates (see Stage F below), include the Chopawomsic, Arvonia, Carolina Slate, and James Run Belts in North Carolina, Virginia and Maryland and the Sonoma Belt in northern California and Nevada. We find relatively young (Cenozoic) blueschist rocks today (they are common in parts of California), but they are easily altered so ancient accretionary wedges are not longer blueschist in composition.

Stage F: Island Arc-Continent Collision

Description of Process: As the subduction that started in Stage E progresses, oceanic crust is consumed; these disappearing ocean basins are called remnant ocean basins, since they are shrinking remnants of their former selves. Eventually the remnant ocean basin is completely consumed and the island arc collides with the DCM of Westcontinent. The collision deactivates the subduction zone, but before doing so the subduction zone works as a ramp causing the volcanic arc to ride up over the edge of Westcontinent. In all such collisions the overriding block is called the hinterland, and the overridden block the foreland.

The collision causes a tremendous amount of structural deformation. Sediments that were formerly part of Westcontinent's DCM are folded and thrust faulted up and back toward the center of the continent, while additional thrust faults develop in the volcanic hinterland. Mountain ranges that result from these upthrusts can be the size of the European Alps. Between the hinterland and foreland is the suture zone that "sutures" the two blocks together. Typically the suture zone contains accretionary wedge rocks—all that remains of the former ocean basin—but may exist only as a simple thrust fault plane. Here a person can straddle what were once two distinct and widely separated blocks, with one foot on the continent and one on the volcanic arc. The suture zone also separates distinctly different rocks – volcanic and metamorphic rocks making up the hinterland and sedimentary rocks the foreland.

Inland from the mountain range a foreland basin subsides, geologically rapidly at first, but then every more slowly with time. When sea level has been higher, as in the geologic past, the foreland basin environment is marine. Initially the sediments filling the foreland basin are deposited in deep water (400 to perhaps 600 feet deep), but as the mountain erodes more and more the foreland basin fills in with shelf, then shoreline, and then river environments. These river sediments are common when sea level is lower, as it is today, creating a terrestrial environment for the foreland basin. Total sediment accumulation is on the order of 10,000 to 20,000 feet. These are some of the more common and conspicuous deposits in the geologic record.

One of the more important processes taking place during these tectonic events is that rocks are evolving. We saw igneous rocks evolve by fractionation during Stage E (mafic rocks descend along the subduction zone, fractionally melt to produce intermediate magmas that rise up to form batholiths, leaving behind unmelted residue that forms ultramafic rocks that stay in the mantle.) Sedimentary rocks undergo similar fractionational evolution. When a parent rocks, such as a granite, first begins to weather, abundant feldspar and quartz are released, as these are the primary minerals in a granite. The resulting sediments are feldspar and quartz rich sediments, called arkosic, which are immature sands because they have been deposited before feldspars can weather into clay and grains rounded and sorted. Quartz and feldspar do not weather equally well, however. As the feldspar changes into clay, quartz is left behind.

For a sandstone to evolve from immature to mature takes many millions of years, and typically more than one cycle of uplift, weathering/erosion, deposition, and re-uplift. We know from Stage E that the sediments deposited around the volcanic arc are immature; there has just not been the time for them to evolve much. With the arc-Westcontinent collision and mountain formation, these old sandstones are uplifted, weathered, and eroded anew, as processes work to tear down the mountain range. As a result, the material making up the sedimentary rocks is re-worked as the rocks are weathered and eroded again. Chemical and mechanical weathering processes, erosion, and deposition work to round grains even further, sort grains to a higher degree, and remove even more minerals that are unstable, leaving behind the quartz that is more stable under Earth surface conditions. These mountain building events are described as "recycled orogens" meaning the uplifted rocks have already been through at least one mountain building-maturation cycle.

<u>Composition:</u> As the final convergence proceeds, former DCM sediments are compacted and lithified further, enough that greenschist grade metamorphism can occur. Expanses of sandstone and limestone become quartzites and marbles, while shale becomes slate. As the new mountains are raised and erosion occurs at a high rate, the reworking of rock material produces sediments that have much less feldspar and lithic fragments, but more clay and quartz. Submature sandstones (intermediate between immature and mature) are the result. The QFL composition of sediments in Stage F starts off with an already high quartz content, generally over 50 % (Figure 2, Field C).

Of course, not only older sedimentary rocks are exposed and eroded, but igneous and metamorphic rocks that were formed deep beneath the surface are exposed too, providing a window into the deep crust. This means that even though the ultimate outcome of weathering and sandstone formation is always the same—quartz sands—the details of that evolution of sediments does not always follow exactly the same path. The arrow in Figure 6 shows the general path of sediment evolution.

<u>Locations:</u> Modern locales where Stage F is occuring or imminent are limited, but the incipient collision of Australia with Indonesia/New Guinea is the clearest and best. The north dipping subduction zone has produced the volcanic mountains seen in Indonesia and New Guinea (e.g. the Owen Stanley Range). In the geologically near future (couple of tens of millions of years) the Arafura and Timor seas–which are remnant ocean basins–between Indonesia/New Guinea and Australia will disappear down the subdution zone, and Indonesia/New Guinea as the hinterland will collide with northern Australia as the foreland.

Some of the best exposed and long studied ancient examples of volcanic arc continent collisions are associated with the Ordovician Taconic and Devonian Acadian mountain building events, or orogenies, all up and down the North American Atlantic seaboard from Newfoundland to the Carolinas. A whole series of volcanic arcs are exposed in the piedmont, including the Chopowansic, Arvonia, James Run, Carolina Slate Belt, and Inner Piedmont regions of Georgia, South and North Carolina, Virginia and Maryland. The foreland basins with their thick sediment fills are preserved west of the Blue Ridge mountains in the Valley and Ridge province.

Stage G: Cordillera Mountain Building

Description of Process: Once a remnant ocean basin has closed, and the mountain range built, the only thing left for the mountains of Westcontinent to do is erode. Stage F built the arc-Westcontinent mountain, but by the time of Stage G the mountain is eroded down nearly to sea level—a peneplain. Walking across this eroded landscape is to peer deep inside the Earth. Nearly all the rocks exposed across this peneplain—plutonic igneous and metamorphic—can only form deep inside the Earth, miles below the surface. For them to be visible now at the surface means that all those miles of overlying rock have been removed by erosion. The only rock actively forming across the peneplain is a mature quartz sandstone, the very last unweatherable, scant remnants of the entire mountain range that used to be here.

Meanwhile, on the east side of the ocean basin, another convergent plate boundary has initiated along the edge of Eastcontinent. In Stages C, D, E, and F Eastcontinent has been a tectonically stable—even geologically boring—divergent continental margin. But, now the oceanic basin and continent decouple (break apart) near where they change over. Because the oceanic place is old, cold, and denser than the continent it begins to subduct beneath Westcontinent. As the ocean floor material wedges down into hotter, more ductile mantle material a slab-pull force begins to pull and drag the rest of the ocean plate into the subduction zone. The slab-pull force appears to be the dominant force in plate motions, accounting for up to 50% or more of the plates driving force. The subduction sets in motion a series of processes. First, a deep ocean trench is formed on the ocean floor where the ocean plate begins its descent. At the same time, powerful medium to deep focus earthquakes occur as the plates jerk and slide past one another due to the forces of convergence. Third, as the ocean plate is pulled into the mantle, the increased temperature and water saturated conditions cause the plate to partially melt; pods, or diapirs, of intermediate magma float up through the crust, emplacing as vast and expanding batholiths. In time the magma reaches the surface forming spectacular stratovolcanoes.

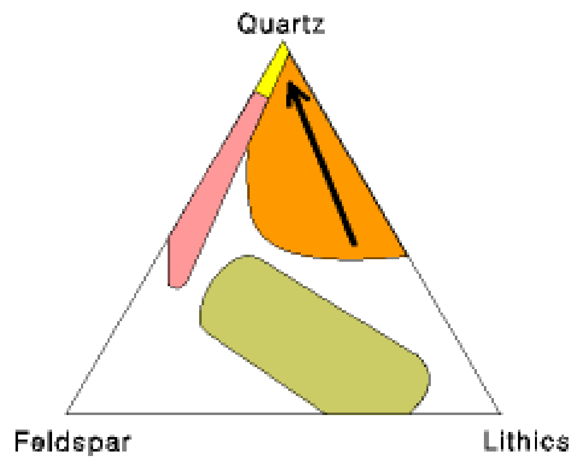
While the oceanic plate is subducting during convergence the continental plate margin begins to thicken. The wedge of DCM sediments that have been accumulating on the edge of Eastcontinent for the past 100 million years or more, are now subjected to increased heat and pressure. Because these rocks have become pliable from the heat and pressure they undergo ductile deformation: folding, stretching, contorting, and thinning like taffy or silly putty. As a part of this process, the rocks are upwardly thrust to build into a large mountain range. Simultaneously, the heat and pressure alters the old DCM sediments, resulting in widespread and expanding Barrovian metamorphism.

As the mountain builds it begins to weather and erode. For millions of years the up-thrust is more effective than the weathering and erosion and the mountain builds up and up many kilometers above sea level. Still, large volumes of sedimentary material are created. Some of these sediments are deposited on the hinterland side, but many pour into the ocean and down into the deep ocean trench just off the coast where are deposited. Almost as soon as the sediments are deposited in the trench though, they are wrenched off again by the descending oceanic conveyor slab to be simultaneous dragged into the Earth and scraped off and thrust up to form an accretionary wedge. As in Stage E these accrectionary wedge sediments are subject to rapidly increasing pressure, but very slow heating and undergo blueschist metamorphism.

Composition:

Igneous: One of the most impressive expressions of the convergence is the eruption of large composite (strato) volcanoes. The fractional melting of the ocean plate and the base of the continent, generates magmas and lavas enriched in silica, aluminum, potassium, and sodium, and depleted in iron, calcium, and magnesium. The resulting igneous rocks are andesites, dacites, and rhyolites, with diorites, white (plagio-) granites, and pink granites as their plutonic counterparts. Also noteworthy are the large volumes of volatile material in the lavas, such as water vapor and carbon dioxide. When combined with the more viscous, silica-rich lavas, these volatiles build up to produce violently explosive eruptions, large quantities of volcanic ash, and very slow-moving lavas. In the subsurface, these same volatile materials produce hydrothermal veins and coarse-grained pegmatites that penetrate into the DCM sediments and country rock. Minerals grains with sizes often much greater, than 5 cm are common. Contrary to the normal rule of coarse-grained materials cooling very slowly, pegmatites form quickly, with the volatile material facilitating the formation of large grain in a manner similar to frost forming on cold, damp nights. Rare elements that do not easily fit into other minerals, such as beryllium and lithium, are concentrated here to form emerald, aquamarine, and spodumene. These crystals are huge, with single grains of spodumene recorded at the size of utility poles.

Sedimentary: As the mountains build erosion work to tear the mountains back down. Many of the rocks composing the mountain are older DCM rocks that have already been through at least one cycle of weathering, erosion, and deposition. This next generation of weathering and erosion reworks the sediment again, increasing its maturity (orange field in the diagram are sediments derived from the erosion of these second generation rocks). As time goes by the remaining feldspar and lithic grains are diminished even more—chemically weathering to form clays—while quartz abundance correspondingly increases. The arrow shows the evolutionary path of the sediments.



Metamorphic: Rocks that form in the Cordillera represent the "classic" sequence of Barrovian metamorphic rocks first described by George Barrow in Scotland. When the shales that were part of the old DCM are subjected to heat and pressure, clay minerals reform into microscopic, green chlorite mica grains that align themselves perpendicular to the pressure, producing a slate. With increased temperature and pressure the chlorite crystals increase in size, forming phyllite, a slate-like rock with a distinct sheen. Slate and phyllite are said to be greenshist, or low grade metamorphic rocks. At higher temperatures and pressures, the chlorite grains are no longer stable, and change to form muscovite and biotite grains that are more segregated from feldspar and quartz grains. The new rock is called a schist, which is of medium or amphibolite grade. The highest grade metamorphic rocks in Stage G are gneisses, in which the grains of quartz, feldspar, mica, and garnet separated into distinct bands. These are the equivalent of granulite grade rocks in Stage E. Former quartz sandstones become quartzite, and limestones become marble.

Off the coast in the accretionary wedge is the high pressure/low temperature blueschist metamorphism, and deep in the mantle, under the continent is eclogite metamorphism, although these rocks are so deep they are not often brought up from depth and seen at the surface. Contact metamorphism is found in many places throughout the mountain where intrusives like dikes and sills cut through the country rock, but this will be mostly local in effect.

<u>Locations:</u> The classic present day location for a Cordilleran formation is along the west coasts of South America and North America (especially the Cascade Mountains of Washington and Oregon), forming the spine of the Andes and Rocky Mountains. The Alps of Europe and the Appenine mountains of Italy have a similar appearance.

Ancient examples are widespread on most continental land masses. A major example is the Sierra Nevada batholiths in the Sierra Nevada mountains of California. A little further west in the coast ranges and off the coast in the Channel Islands are well developed and well exposed ancient blueschist accretionary wedges. In the east many of the rocks in the Blue Ridge mountains (or more technically Blue Ridge province) in Pennsylvania, Virginia, and North Carolina are multiple subduction generated batholiths, almost all later metamorphosed by the South America-North America collision that created the Rodinia supercontinent (as in Stage H).

Stage H: Continent-Continent Collision

Description of Process: In many respects Stage H resembles Stage F, in that a remnant ocean basin has been completely consumed in a subduction zone, this time resulting from a collision between Westccontinent and Eastcontinent. This collision and mountain building finishes closing the ocean basin that began opening in Stage C. In our Wilson Cycle model, the ocean basin closes in just two stages. A real ocean may close in many stages. There can be two, three, or more volcanic arcs caught up in the closing. The subduction zones may dip in different directions than is depicted–east, west, some other way, or any combination of these. In many ancient examples, several volcanic arcs and microcontinents are trapped between the two continents, creating a complex history. In other words, there may be a series of Stage E-G events before the transition from Stage G to Stage H. But while these multiple events add complexity, the processes and rocks generated at each stage are the same in all these cases, and the same basic principles are used to unravel the events and geologic history.

As in the Stage F collision the subduction zone acts as a ramp and Eastcontinent slides up and over the edge of Westcontinent (with its sutured, extinct volcanic arc). With the collision the crust is thickened considerably at the point of contact. Westcontinent is pushed down, forming the foreland, while Eastcontinent is buoyed upwards, becoming the hinterland with an enormous mountain range on the front margin.

There is a considerable disruption of rock at the point of contact, producing folding and thrusting of material from the hinterland onto the foreland. In between the two blocks a suture zone forms, linking the two blocks together. Layers of rock caught between the plates sometimes squeeze out at the suture zone. The layers of rocks move along large thrust faults–and like dough being folded–form large, overlapping, stacked folds called nappes.

All the thrust faulting brings previously deeply buried rocks, such as the Stage G igneous plutons and the Barrovian metamorphic slates, phyllites, schists, and gneisses, up from the depths, while weathering and erosion strips off the overlying rock to leave them exposed at the surface. On the surface the metamorphic grade of the rocks can be mapped in discrete zones, with the highest grade material near the line of suture or near plutons. Slices of the subducted, perhaps partially melted ocean plate can also be thrust upward, interlayering eclogites, serpentinites, soapstones, amphibolites, and/or granulites within shallower forming rocks. In places, complete sequences of oceanic lithosphere—called the ophiolite suite—are brought to the surface and exposed.

Meanwhile, on the foreland side of the mountain rivers carry sediment toward the foreland basin, which begins filling with sediment. Initially the foreland basin is deep water, and the rivers feeding into it are relatively short. But, as the basin fills and shallows upward the river systems get longer and longer, and river deposits become a more significant part of the geologic record.

Composition:

Igneous: A collision always involves at least one active subduction zone to consume the remnant ocean basin. Thus, the Stage H collision is preceded by the Cordilleran orogeny under Eastcontinent that began in Stage G. Thus, all the igneous activity of stage G continues into Stage H, tapering off as the final suturing will bring this to an end.

Sedimentary: With such a complex tectonic history rocks of many different kinds and histories are brought to the surface for weathering and erosion. Older sedimentary rocks may be in their third or fourth generation of evolution, and are thus quartz rich. On the other hand, a wide diversity of new kinds of igneous and metamorphic rocks are exposed, weathered, and eroded. These sediments are often immature, containing large fractions of feldspars and lithic fragments. But, these also are clues to the mystery tale. After mountains are removed by erosion, the only evidence that remains of the nature and composition of the mountains are from the sediments that were eroded from them. By identifying the kinds of unweathered feldspars and lithic fragments, and the order and sequence in which they are deposited, we can reconstruct a large part of the history of the mountain range. The adage “no rock is accidental” takes on particular pertinence here. Even the tiny weathered fragments when studied through a microscope contain volumes of information about what kind of rock they weathered from, the conditions under which they formed, the transformations they have undergone, the different kinds of rocks that were exposed at the same time in the same place, and so on. This is some of the most interesting detective work in geology.

Metamorphic: Metamorphism is pervasive throughout this kind of mountain building, not only from heat emanating from the igneous activity, but also because bodies of rock are being compressed, stressed, brought up from depth, and driven down into the depths again by the structural movements. We expect Barrovian, contact, and hydrothermal processes to be common. Adding even more complexity is retrograde metamorphism, in which high grade rocks can be re-metamorphosed to a lower grade. For example, rocks formed at great depth are uplifted to shallow depths where they adjust by overprinting the high pressure/high temperature minerals and textures with low pressure/low temperature ones. Some rocks may be metamorphosed multiple times, with each event leaving something of its occurrence imprinted in the rocks. Interpreting the rock, and thus its tectonic history, becomes a detective story where faint clues, partially-destroyed evidence, distracting evidence, and a local-regional focus tension can lead to vigorous arguments on the origins of a landscape – the essence of the scientific process in the Geosciences.

Locations: Perhaps the best example of Stage H in the modern world is the Himalayan Mountains. The Indian subcontinent is the foreland, while Tibet is the hinterland. This same general line of mountains extends from Southeast Asia, through the Himalayan Mountains, the Caucuses and Anatolian Mountains, the Balkans and into the Alps. The Pyrenees and Urals Mountains also mark such margins, although of older age. The remains of ancient events include the Alleghenian orogenic features of the Appalachians and the Calodonian Mountains of Scandinavia. Suture zones are also evident in Nova Scotia, Newfoundland, Ireland, and Scotland. Evidence for even older such events is seen in the Grenville-age metamorphic rocks of the Blue Ridge. These rocks are so old as to have been through the rock cycle more than once.

Stage I: Erosion to Peneplain

Description of Process: The mountains that formed in Stage H are no more, ground down to their very roots. The processes of erosion through water, wind, and ice have taken the once proud mountains, broken them down to fragments of sand and clay, and transported this material to every low-lying area of the continent, sorting and rounding material to full maturity. Eventually, the entire continental surface is reduced to no more than a few hundred feet above sea level , with relatively little relief. This peneplain surface is so close to the lowest possible level of erosion that further reduction is almost imperceptible. This peneplain surface becomes an effective base level, the lowest level to which running water can erode the land. As the mountains are worn down, the removal of overburden causes deep seated rocks, such as high-grade metamorphic and coarse-grained igneous rocks, to be brought to the surface, acting as windows to the deep continental crust. Sometimes these rocks are exposed at the surface, but often they are buried under a thin veneer of younger sedimentary rocks.

Superficially, Stage I resembles Stage A in that we are dealing with a single continent again, but with two significant differences. First, even though Westcontinent and Eastcontinent are now rejoined to form one continent, the Stage I continent is larger than in Stage A, in part because of the volcanic arc that has been incorporated into its structure. In this fashion, Stage I demonstrates a basic principle; with each Wilson cycle: continents grow larger. If we extend this process back through the Earth's geologic history we discover that the first continents were just tiny blocks, no bigger than some states today, or the islands of Madagascar and New Zealand. Continents have not always been as big as they are today. They have had to grow with time, from tiny proto- and micro-continents to larger and larger continents. At times all the continents have even coalesced to form supercontinents (like Rodinia and Pangaea), but if we could have flown by the Earth 3.5 billion years ago, instead of the continents we see today we would be impressed by how huge and barren the ocean looks—and it would probably appear as just one large ocean. And, within that ocean, scattered here and there, are small islands—volcanic arcs, proto-continents, microcontinents—the first inklings of what a contemporary Earth would come to look like—several billion years in the future.

There is a second difference. In Stage A, the model was drawn very simply. Notice that the Stage A, the continent appears uniform throughout. But looking over all the other events recorded in the Wilson cycle, that simplicity cannot possibly be accurate. The continent in Stage A was formed by similar processes to those occurring in all the other stages of the Wilson cycle. Thus, if we were to draw Stage A accurately, it would have to contain a variety of igneous, sedimentary, and metamorphic rocks and provinces. However, for the sake of this model, and learning about the Wilson cycle, a simple Stage A is presented for the sake of clarity.

<u>Composition</u>: In contrast to the previous stages, there are few active processes producing new rock materials—the continent is tectonically asleep. If one were to walk across the exposed peneplained surface, they would be amazed at the diversity and complexity of rocks exposed. Locally, the composition reflects the history of the individual events that have invaded, been sutured onto, or altered the original continental block. There are bands of mafic rock, bodies of felsic rock, and zones of sedimentary rock, most of which have been more or less deformed and metamorphosed. Material has been through the rock cycle multiple times. Overall, however, the average continental composition is that of a granodiorite – enriched in silica, aluminum, sodium, and potassium. As in Stage A granodiorite has a lower specific gravity than the mafic rocks that compose the ocean basins the continents "float" higher on the underlying mantle. A peneplained continent in isostatic equilibrium, "floats" with its surface only a few hundred feet above average sea level.

The only active rock processes in Stage I are sedimentary, and they fall into two categories. The first involves quartz sandstones. The only thing present is the one thing that cannot weather, which is quartz. Feldspar and lithic fragments have been removed from the sediment by processes in the previous stages. The quartz is referred to as a "lag" deposit, because it lags behind after everything else is removed, and it forms a thin "veneer" (which may in fact be a couple of kilometers thick) blanketing the continent. The quartz is the only thing that remains of all the mountain building events that formed the continent. True, there might well be clay mixed in with it, but over millions of years of exposure rain washes the clay down to the sea, or wind blows it into the atmosphere and off the continent. Thus, as Figures 2 and 7 show, we are finally at the top apex of the QFL diagram.

The second category of sedimentary rocks to form in Stage I are limestones, but they dependent on sea level and climate. The climate must be tropical or subtropical for carbonate rocks to precipitate from seawater, so the location of the continent with respect to the Equator is critical. Sea level today is near one of its lowest stages in Earth history, and even peneplained continents are exposed. When sea level is near maximum height, however, continental cratons are largely under water and it is easy to have limestones deposited across most of them.

<u>Locations</u>: Current examples of Stage I include many of the same continents described in Stage A. The large, low elevation, low relief parts of Australia, Siberia, Canada, and Africa contain the roots of ancient volcanic arcs, mountains ranges, ocean basins, and sutures now long eroded. The core of all the modern continents—those areas referred to as shields (stable areas with exposed bedrock) or cratons (stable areas with a thin veneer of sedimentary cover)—are typically composed of several individual terranes, each with its own history. For example, ancient marine limestones are common in North America from the Gulf coast to the arctic circle, areas not commonly associated with tropical climates.

North America in particular exhibits 4 billion years of continental growth, made of dozens of individual fragments of a wide range of sizes. In the United States most of these ancient terranes are covered by younger sedimentary rocks, and hard to see. The broad, relatively flat Great Plains of the central region is underlain by the roots of multiple volcanic arcs and microcontinents, but because they are sediment covered we know very little about them. In the mountains of the western United States uplift and erosion has sometimes produced windows through the overlying younger sedimentary rocks allowing us to see these ancient remains. Examples include, among others, rocks at the bottom of the Grand Canyon, the central portion of the Black Hills, and the Front Range and Sangre de Cristo mountains in Colorado and New Mexico. In the east, the rocks in the Blue Ridge and Adirondack mountains are ancient terranes. Canada has the best exposures of ancient terranes since most of the central region all the way to the arctic has the exposed roots of ancient mountains and arcs.