Sedimentary Rock Formations of the Grand Canyon

Introduction

The sedimentary geology of the Grand Canyon region is extremely diverse and spans more than a billion years of earth's history. Two distinct age groups of sedimentary rock sequences are exposed within the canyon's depths (Figure 1). The older, Late Proterozoic sedimentary sequence is comprised of the Grand Canyon Supergroup which consists of the Chaur Group, the Nankoweap Formation, the Unkar Group, and the Sixtymile Formation, and is only found in isolated patches along the main Colorado River corridor and some of its major tributaries (Figure 2). Beginning about 1,250 million years ago and lasting about 500 million years (during the Late Proterozoic Era), approximately 13,000 feet of sediments and lava were deposited in coastal and shallow marine environments. Basin-and-Range style crustal deformation beginning about 750 million years ago lifted and tilted these rocks. Subsequent erosion removed these tilted layers from much of the Grand Canyon region leaving only wedge-shaped remnants preserved in large graben structures (Figure 2), mainly observed in the eastern parts of the canyon. The younger sedimentary sequence comprises much of the Paleozoic Era and forms the vast majority of rocks exposed in the canyon's walls (Figure 2). These mudstones, sandstones, and limestones are widely distributed in the canyon, but total a mere 2,400 and 5,000 feet thick by comparison with Proterozoic rocks. They offer a plethora of evidence interpreted as coastal and marine environments, including several significant marine incursions from the west, developed on a passive continental margin setting between about 550 and 250 million years ago. Rock formations from the Cambrian, Devonian, Mississippian, Pennsylvanian and Permian periods are present.

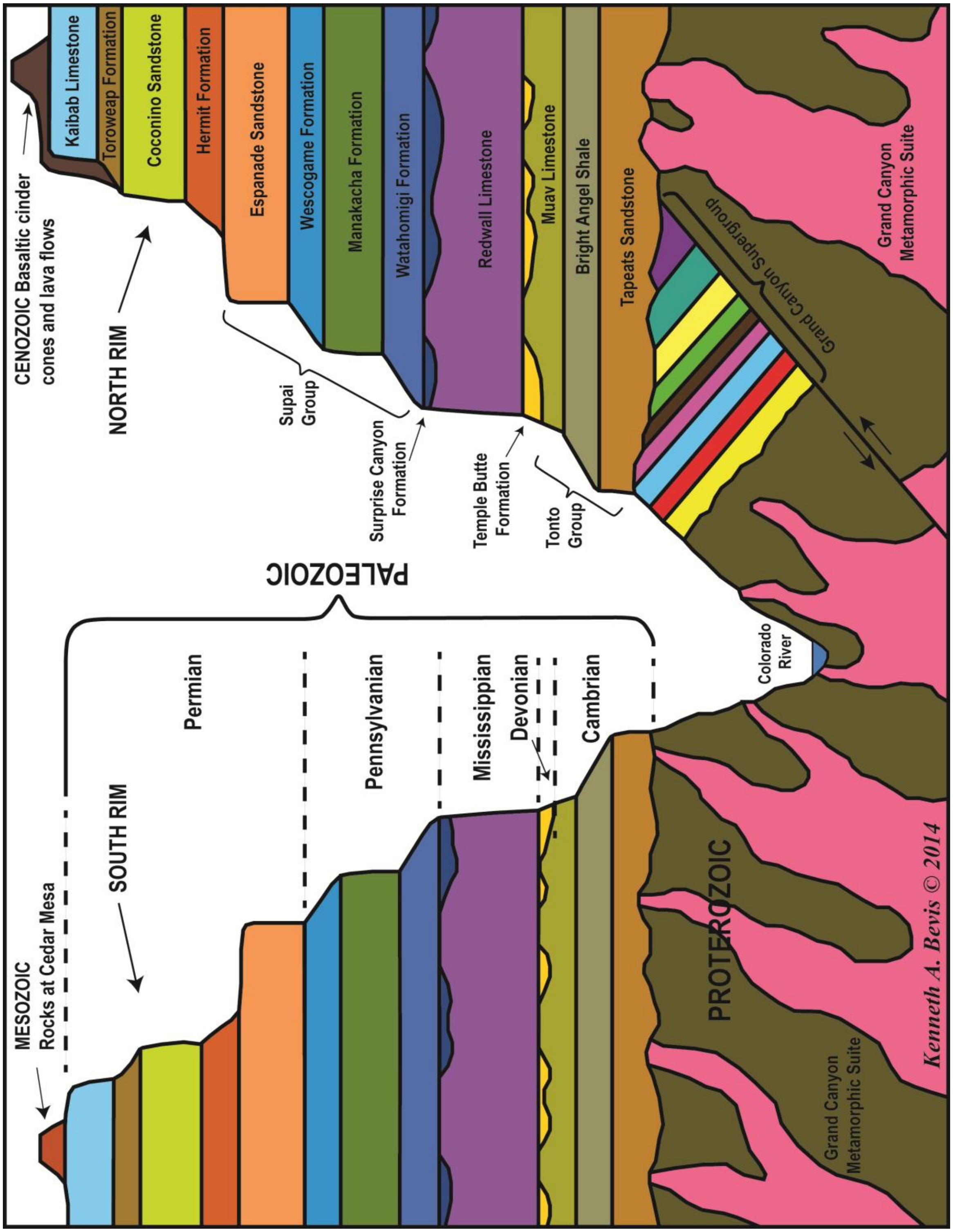


Figure 1. The suite of sedimentary rocks exposed by the downcutting of the Colorado River in Grand Canyon National Park includes an older Proterozoic sequence, and a younger Paleozoic sequence.

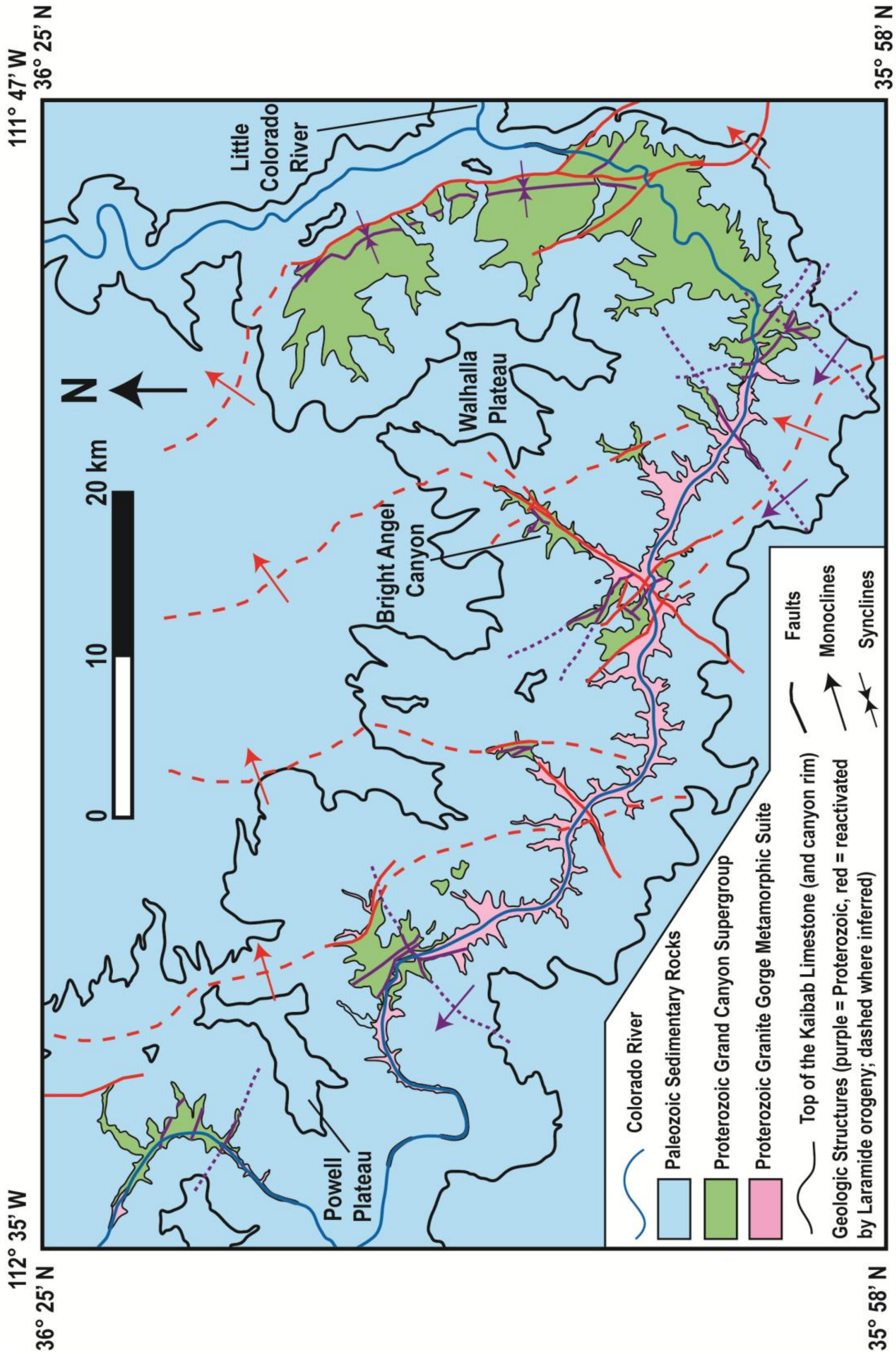


Figure 2. A geologic map of the eastern Grand Canyon area indicating the general outcrop locations of Proterozoic crystalline basement and Grand Canyon Super Group sedimentary rocks,

sedimentary rocks of the Paleozoic sequence, and geologic structures; note the general juxtaposition of Supergroup rocks against bounding normal faults.

Erosion has removed most Mesozoic Era sedimentary rocks from the region, although small remnants can be found, particularly in the western Grand Canyon. Nearby rock outcrops, particularly to the north in the Grand Staircase area, suggest 4,000 to 8,000 feet of Mesozoic sedimentary layers once covered the Grand Canyon region, but were removed by uplift and erosion in the early Teritary. Cenozoic Era sediments and sedimentary rocks are limited to the western Grand Canyon and to stream terraces and travertine deposits found superimposed on older rocks near the Colorado River itself. Lava flows and associated cinder cones comprise the majority of Cenozoic deposits. Volcanic activity began between about nine and six million years ago and is still ongoing in some areas. One major volcanic field formed on the Shivwits and Uinkaret Plateaus in the northwestern Grand Canyon region, including spectacular lava cascades that poured down the canyon walls like so much candle wax; while similar volcanics formed south of the Grand Canyon in the San Francisco Volcanic Field.

The Late Proterozoic Grand Canyon Supergroup

History and Geologic Setting

The Grand Canyon of the Colorado River exposes nearly two billion years of earth's history, including Paleozoic sedimentary rocks, Late Proterozoic sedimentary rocks, and Middle Proterozoic crystalline basement rocks, plus a multitude of faults and folds related to ancient and ongoing regional tectonic upheaval. Blakey and Ranney (2008) report that the basement rock of the Colorado Plateau, formed approximately 1,750 million years ago, is composed of metamorphic rock laced by igneous intrusions. These rocks have been referred to as the Grand Canyon Metamorphic Suite or "Precambrian crystalline rocks," and most recently, Timmons et al. (2012) denotes them as the Granite Gorge Metamorphic Suite and Granitoids. The metamorphic rocks formed from the sandstones and mudrocks, volcaniclastic material, and volcanic rocks accumulated within elongate basins formed between successive volcanic arcs and the continental mainland as the island arcs collided with the proto-North American craton. The volcanic arcs travelled northwest to merge with the continental edge, which had only extended to the current location of Utah and southern Wyoming. These collisions folded the basin sediments accordion-style, forced them to a great depth where they underwent metamorphism, and sutured them to the continent to become part of its basement. Co-generational igneous rocks were formed as the deepest of the subducted material melted into magma and proceeded to rise buoyantly to the surface, forcing its way into the fractures and foliations of the overlying metamorphic rocks as it rose. This process is best inferred from the observed ribbon-like bodies of intrusive, light-colored, felsic (granitic) rocks of the Zoroaster Granite imbedded within the darker-colored, vertically-layered metamorphic rocks of the Vishnu Schist. The solidified rocks were slowly uplifted, exhumed at the surface, and removed by erosion in orogenic events caused by the collisions and then aided by isostatic uplift (Blakey and Ranney, 2008). The final exhumation from depths averaging 33,000 feet below the surface occurred between 1.3 and 1.25

billion years ago, determined by the cooling age of the feldspars within the granitic intrusions (Timmons et al., 2012), as much as 500 million years after their formation.

Separated by a significant unconformity, scattered wedges of down-faulted, northeasterly-tilted, Late Proterozoic Grand Canyon Supergroup overlie the crystalline basement in the central and eastern Grand Canyon (Figure 2). The Grand Canyon Supergroup are the oldest sedimentary rocks exposed in the Grand Canyon's walls, generated over an extended period of a time between 1255 and 742 million years ago (Timmons et al., 2012). At more than 12,000 feet thick, the group is three times thicker than the entirety of the younger Paleozoic age sedimentary rocks above it. The Late Proterozoic sedimentary package occurs as wedge-like bodies exposed in only a few areas (Figure 2); Timmons et al. (2012) reports that it outcrops best in the eastern Grand Canyon between river mile 63 and 79 and up the Colorado's side canyons between river miles 53 and 63 (although isolated wedges are found in down canyon tributaries as far as river mile 134). The Supergroup accumulated on top of crystalline basement rocks following more than 450 million years of subaerial exposure, erosion, and penaplaination (Hendricks and Stevenson, 2003). Sediments were deposited in coastal and shallow marine environments throughout a shallow seaway that probably extended diagonally across Laurentia (the ancestral North American continent) from at least present-day Lake Superior to Glacier National Park in Montana to the Unita Mountains in Utah and the Grand Canyon of Arizona. In ascending order, it is divided into the Mesoproterozoic Unkar Group (formed between 1255-1100 million years ago), the Nankoweap Formation, and the Neoproterozoic Chuar Group and Sixtymile Formation (formed between 800-742 million years ago). The Nankoweap Formation lies sandwiched between the Unkar and Chuar Groups; its incompleteness makes its development and age difficult to interpret, but the rock unit is believed to have formed around 900 million years ago during a transitional period dominated by an erosional hiatus lasting about 300 million years between the end of Unkar and beginning of Chuar deposition.

The ubiquitous low-angle tilt and patchwork distribution of the Supergroup raises questions about its history since all sediments are considered to have originally been deposited as a blanketing of material on a generally flat surface. During Unkar Group deposition, Laurentia collided with fragments of continental material (now fixed to South America and Africa) along its southeastern margin (Blakey and Ranney, 2008 and Timmons et al., 2012). Collisional uplift induced erosion that shed copious amounts of detrital sediment westward. Back-arc extension thought to be associated with the culminating Grenville Orogeny, an extensive collisional mountain building event culminating between 1.2 and 1.0 billion years ago along the North American continent's northeastern margin likely thinned continental crust regionally, forming large rift basins that would ultimately fail to split the continent. However, thinning of the continental plate probably caused the Grand Canyon region to sink and aided flooding by a shallow seaway. The Cardenas Basalt and related diabase dikes and sills intruding older, underlying Unkar Group rock units formed at the close of Dox Formation deposition, mark outpourings of flood basalt lavas and their subterranean feeder system commonly produced during such rifting.

The Grenville Orogeny came to a close with the assembly of the supercontinent Rodinia, which was likely comprised of an amalgam of the North American, Antarctic, and Australian continents. Deposition of Supergroup rocks continued in the interior seaway from the rising

Grenville Mountains long after completion of Rodinia with the accumulation of the Nankoweap Formation (albeit not without significant periods of erosion), and Chuar Group by about 750 million years ago (Dehler et al., 2012). Periodic flooding of the seaway was tied to global climate fluctuations inducing alternate glaciation (with falling sea level) and interglaciation (with rising sea level). Subsequently, Rodinia began to break up as Antarctica and Australia rifted away, causing crustal extension and graben formation in the Grand Canyon region.

Blakey and Ranney (2008) and Timmons et al. (2012) point out that the remaining blocks of Supergroup rocks now found in the Grand Canyon are closely related to adjacent normal faults (Figure 2). The faults are suggestive of rifting, and indicate extension of the earth's crust by tectonic forces that are credited to the breakup of the supercontinent Rodinia during Late Proterozoic time. In brief, the sedimentary rocks of the Supergroup are believed to have accumulated in a NE-SW elongated basin within Rodinia created by back-arc Grenvillian extension, then became faulted and tilted in the Neoproterozoic as the western landmass of Rodinia (a combined Australia, Antarctica, and China) fragmented from the North American landmass. Although the Grand Canyon region lay to the east of the rift zone, continental crust in the area was stretched generally east-west and fractured along extensive NW-SE oriented normal faults (Figure 2).

Displacement on the Butte Fault was the most significant; Figure 3 presents a geologic map of this major extensional system exposed in the eastern Grand Canyon. According to Timmons et al. (2012), the rocks of the Unkar Group generally dip northeast at an angle of 10 to 30 degrees toward normal faults, themselves dipping at opposing angels of roughly 60 degrees to the southwest. In some fault-basins, deposition continued and synclinal folding of younger Supergroup layers occurred in conjunction with continued regional extensional faulting. Cogenerational deposition and deformation is well-exhibited within the upper Chuar Group and Sixtymile Formation (Dehler et al., 2012), where their juxtaposition against the Late Proterozoic Butte Fault is combined with development of the Chuar Syncline (Figure 3) and synsedimentary landslide-deposited coarse breccias and gravelly beds comprising the 740 million year old Sixtymile Formation. Stated another way, most of the Supergroup rocks had accumulated prior to initiation of rifting-induced normal faulting, but sediments continued to accumulate during faulting and were gradually being folded into synclines as deposition progressed.

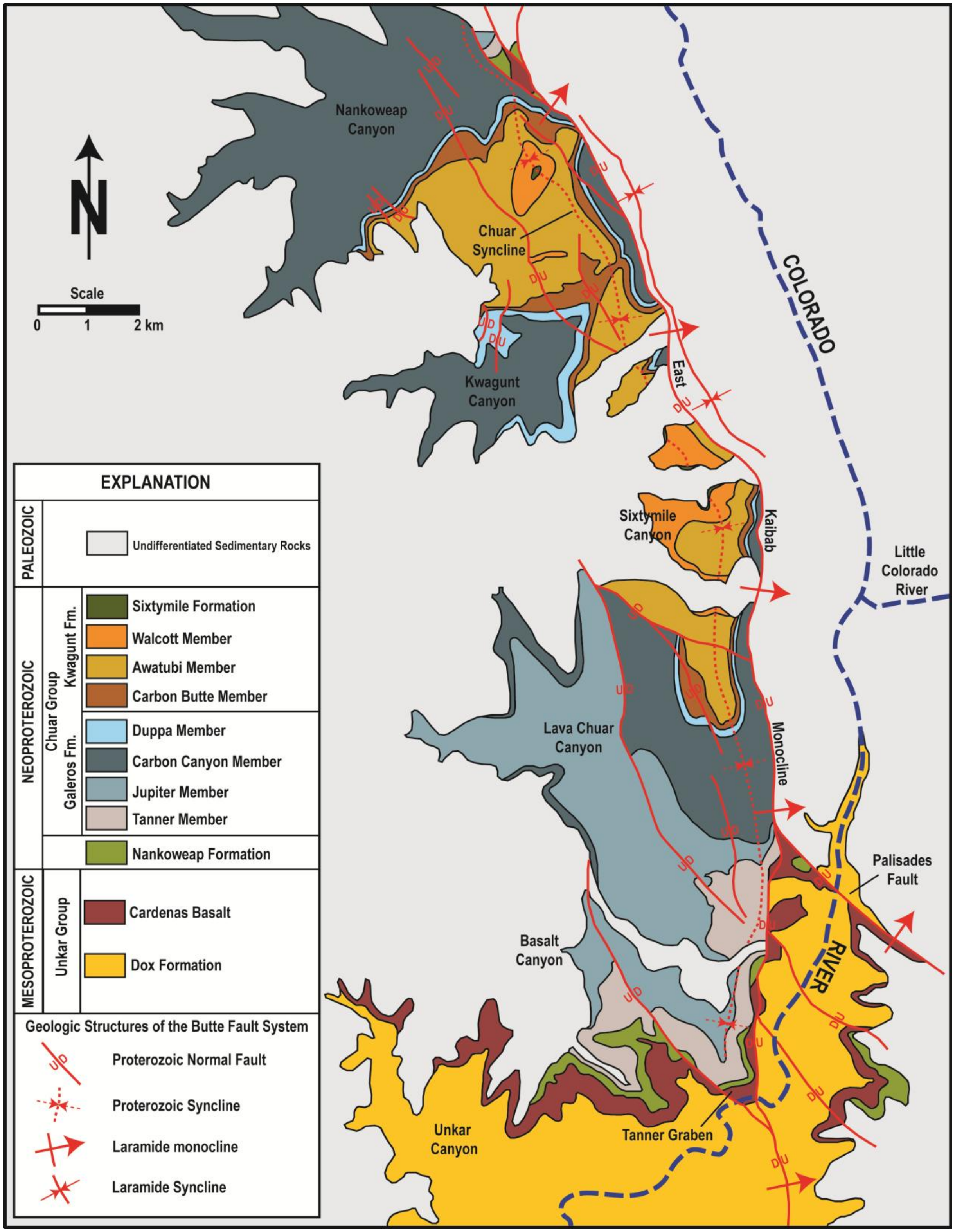


Figure 3. A simplified geologic map of the Butte Fault system in the eastern Grand Canyon (modified from Timmons et al., 2001).

Continued Neoproterozoic normal faulting eventually offset crustal blocks by as much as two vertical miles to form a series of parallel basins and ranges; initially ranges were capped by Supergroup rocks, while basins preserved Supergroup rocks titled backward into one-sided grabens. The majority of sedimentary rocks were forced upward and subsequent erosion from about 740 million to 545 million years ago removed the Grand Canyon Supergroup and more of the underlying crystalline basement rocks from much of the Grand Canyon region, leaving only a patchwork distribution of tilted, wedge-shaped

fragments of the Supergroup preserved in large graben structures (Figure 3), now mainly observed in isolated pockets along the main Colorado River corridor and some of its major tributaries (Blakey and Ranney, 2008). Displacement associated with the the Butte Fault system generated a particularly immense graben (Figure 3), preserving a thick package of sedimentary rocks that includes all known rock units comprising the Grand Canyon Supergroup; it is the only graben exposed in the Grand Canyon that reveals the Nankoweap Formation, the Chuar Group, and the Sixtymile Formation (Figure 3), comprising the upper half of the Late Proterozoic Supergroup sequence.

Erosion once again reduced the mountainous terrain to a peneplain lying near sea level, marked by small hills of resistant Zoroaster Granite and Shinomo Sandstone a few tens to hundreds of feet high. By 545 million years ago, western North America formed a mature passive continental margin, with the waters of the proto-Pacific Ocean lapping at its feet. A slight rise in sea level inundated this nearly flat-lying landscape, eventually to deposit the Tapeats Sandstone, first in a thick sequence of Paleozoic sedimentary rock units. This erosional gap in the geologic record, as much a 1.2 billion years in some areas, and as little as 200 million years in others, has been recognized in other parts of North America and the wider world, and is called the Great Unconformity. The Great Unconformity as it is exposed in the Grand Canyon provides an excellent example of the complex nature of most unconformities, consisting of a nonconformity where the Tapeats Sandstone overlies crystalline, igneous and/or metamorphic rocks of the Grand Canyon Metamorphic Suite, and an angular unconformity where the Tapeats Sandstone overlies the titled sedimentary rocks of the Grand Canyon Supergroup.

The Unkar Group (by Hannah Slover and Ken Bevis)

The Unkar Group, approximately 6,500 feet thick, records a major west to east transgression of the sea onto the western margin of Laurentia, or ancestral North America, with deposition occurring in mostly coastal riverine and shallow-marine environments (Hendricks and Stevenson, 2003; Timmons et al., 2012). The only significant disconformity within the group occurs between the Hakatai Shale and the Shinumo Sandstone, probably representing a temporary withdrawal of marine environments. The Unkar Group likely accumulated in a tectonic basin where the rate of subsidence was similar to the rate of deposition (Hendricks and Stevenson, 2003). Blakey and Ranney (2008) and Timmons et al. (2012) believe it possible that these deposits are related to the Grenville Orogeny, the continental collision in eastern North America that aided in the creation of the supercontinent Rodinia. Their inferred scenario has uplift and erosion induced by orogenic mountain building during the assembly of Rodinia providing sediment to a shallow marine seaway, a flooded back-arc basin or basins on the landward side of a subduction zone.

John W. Powell was the first, on his 19th century river expedition, to recognize the widespread unconformity between the Grand Canyon Supergroup and the Granite Gorge Metamorphic Suite in the southwestern United States, based on significant, observable differences in their geology. During 1882 and 1883, Charles D. Walcott conducted a thorough field study in the eastern Grand Canyon and divided the rocks that Powell had noticed into what he called the Unkar and Chuar Terranes, today referred to as the Unkar and Chuar Groups (Timmons et al., 2012). Together, he referred to them as the Grand Canyon Series and measured a total thickness of 12,000 feet, with his Unkar Terrane being slightly thicker at 6,800 feet. In 1914, Noble classified the Unkar and Chuar divisions as groups. He divided the Unkar Group into five formations. In ascending order, those formations were the Hotauta Conglomerate, Bass Limestone, Hakatai Shale, Shinumo Quartzite, and Dox Sandstone (Hendricks and Stevenson, 2003).

Following Noble's divisions, minor changes in grouping and nomenclature occurred over the years. Van Gundy noticed the unconformities on either side of what he named the Nankoweap Formation, resulting in its consideration as a separate rock unit from Walcott's division. In 1938, Keyes named the basaltic flow at the top of the Unkar Group the "Cardenas lava series," and included it within the Group, but from then until 1987, the name was altered to Rama Formation to Cardenas Lavas to Cardenas Lava (Hendricks and Stevenson, 2003). However, the term "lava" is not favored because it refers to a flow and not a rock, and Timmons et al., 2012, refers to this formation as the Cardenas Basalt.

Further changes to the divisions of the Unkar Group continued. Dalton (1972), suggested the Bass be considered a formation and the Hotauta a member of the Bass Formation. The Dox Sandstone was also changed to the Dox Formation by Stevenson and Beus (1982). Both alterations in classification to a formation were suggested due to the varying lithology. Another modification seen in the work of Timmons et al. (2012), is the Shinumo Quartzite being referred to as the Shinumo Sandstone. The result of over 100 years of study has resulted in an accepted five-fold subdivision of the Unkar Group; in ascending order, the Bass Formation, Hakatai Shale, Shinumo Sandstone, Dox Formation, and Cardenas Basalt (Figure 4).

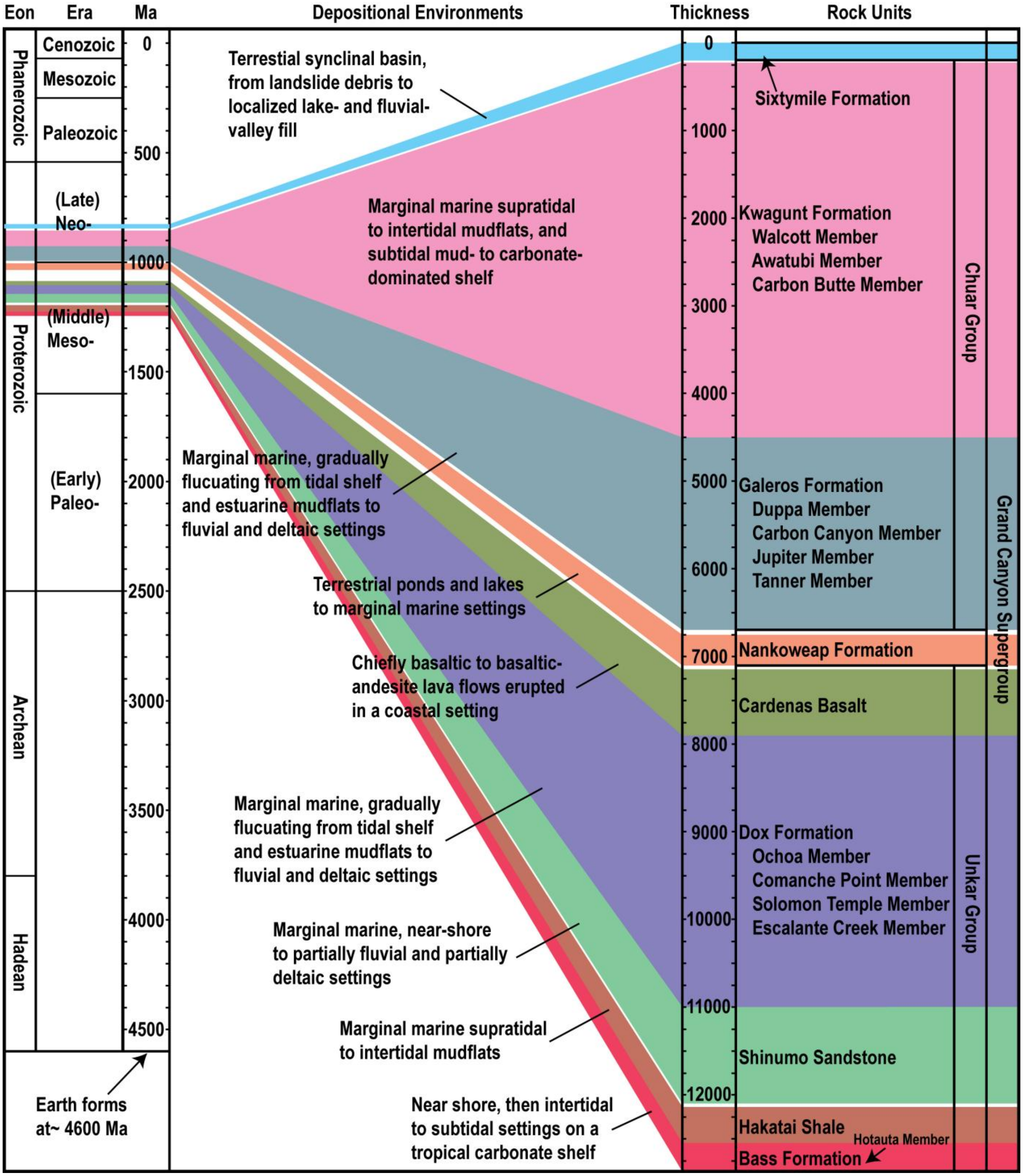


Figure 4. The rock formations of the Grand Canyon Supergroup.

Accumulation of the Grand Canyon Supergroup spans a period of approximately 1150 million years, although deposition was not continuous over this vast expanse of time. The underlying crystalline basement rocks are generally as young as 1700 m.y.a., and the overlying Tapeats Sandstone of the Cambrian Period was deposited about 550 million years ago (Hendricks and Stevenson, 2003). Hendricks and Stevenson (2003) report a Rb-Sr radiometric age determination for the Cardenas Basalt of 1100 m.y.a. Elston (1986) used paleomagnetic pole positions and the polar wondering paths to date the accretion of the Mesoproterozoic Unkar Group from approximately 1250 to 1070 million years ago; in agreement with the more precise Rb-Sr method used on the Cardenas Basalt. More recently, Timmons et al. (2012) has provided more refined dates using techniques unavailable to earlier researchers. Using the new data produced by these techniques, they inferred the maximum age of the Bass Formation to be 1254 Ma, and the eruption of the Cardenas Basalt to be ca. 1104 m.y. ago, indicating that the Unkar Group accumulated over a period of roughly 150 million years. These dates also allow us to determine the time span of the unconformity separating the Unkar Group from the crystalline basement to represent at least 450 million years (Blakey and Ranney, 2008; Hendricks and Stevenson, 2003).

Bass Formation

The Bass Formation, the lowermost unit of the Unkar Group (Figure 4), everywhere lies upon eroded Middle Proterozoic crystalline basement (the Vishnu Schist and Zoroaster Granite), and is observed where it forms the base of eastward, back-tilted wedges of Supergroup rock preserved in Late Proterozoic extensional grabens of the north-central and eastern Grand Canyon (Figure 3). The immense graben formed by extension on the Butte Fault system provides the most pervasive exposures of the Supergroup. Within the thick wedge of Supergroup rocks preserved in this graben, the Bass Formation crops out at river level just below Hance Rapids (river mile 77 at the mouth of Red Canyon), and begins to climb the walls of the Grand Canyon's inner gorge (Figure 5). Exposures here at the entrance to Granite Gorge and immediately south in Mineral Canyon offer an excellent opportunity for observation of this unit and for understanding the entire Supergroup's relationship to graben preservation. The Bass has a relief of 150 feet or less and occurs as a cliff or stair-stepped cliff, where the more resistant riser(s) is (are) composed of dolomite and the steep treads of shale and clay-rich sandstones. The thickness of the Bass Formation is greatest in the northwest at 330 feet at Phantom Creek. It thins toward the east to 187 feet at Crystal Creek, possibly due to a topographic high on the paleo-Vishnu Schist surface. Within the Bass Formation is the basal Hotauta Conglomerate Member. Deposition of this conglomerate occurs in low areas of the ancestral terrane. Hendricks and Stevenson (2003) describe the conglomerate occurring in the eastern Grand Canyon as composed of gravel-sized clasts containing chert, granite, quartz, plagioclase crystals, and micropegmatites in a quartz-sand matrix. In the west, the conglomerate transitions to an array of intraformational breccias and small pebbles, indicating an eastern source for the clasts (Hendricks and Stevenson, 2003). Timmons et al. (2012) point out that the quartzite occurring in the Hotauta Member is not of the Grand Canyon; supporting their contention that a river system carried the conglomerates from a distal source, likely from the east. The Hotauta Member is also well exposed along Bright Angel Creek, just as the North Kaibab Trail descends into The Box.

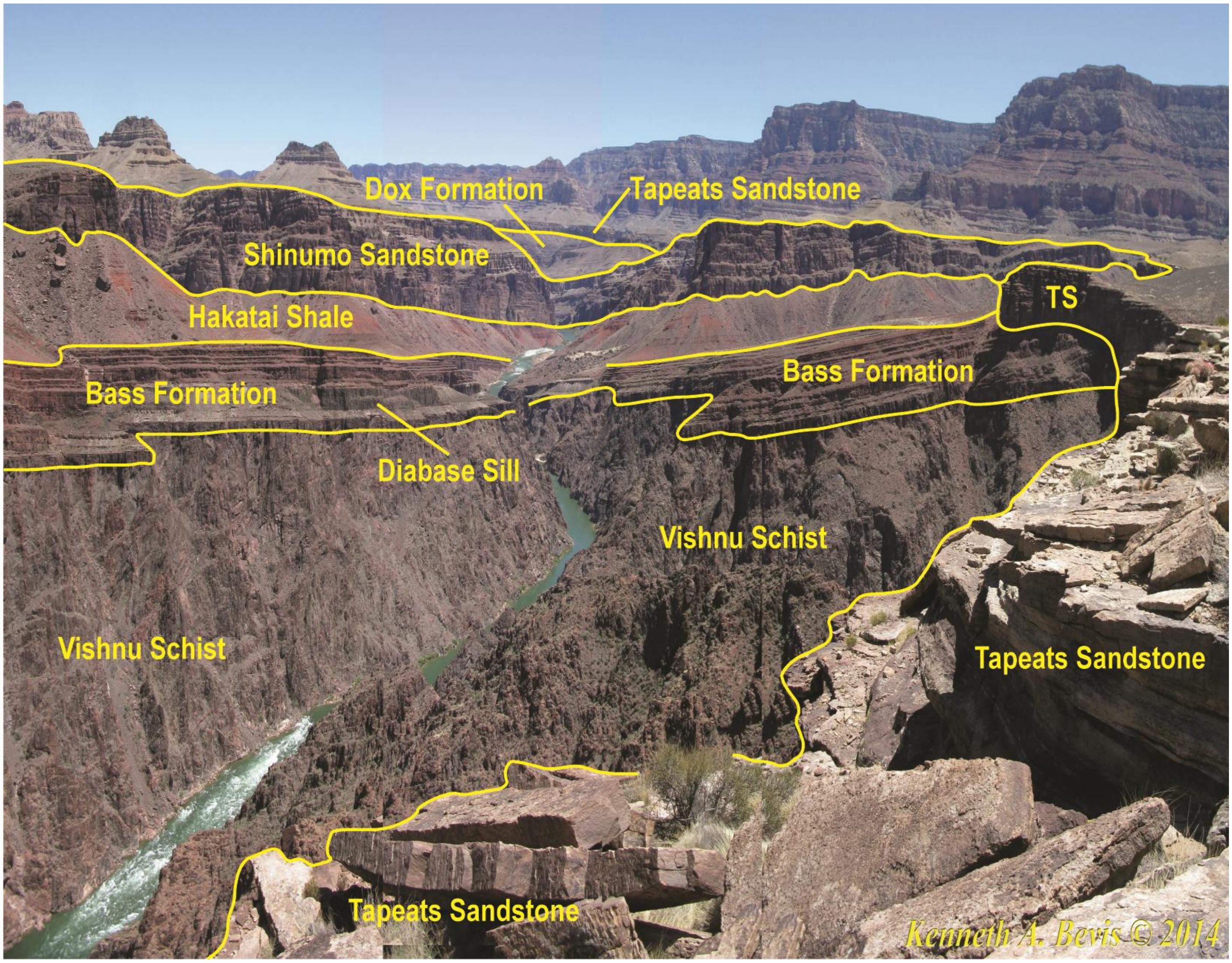


Figure 5. Eastward dipping layers of the lower Unkar Group, backtilted into the Butte Fault graben, rest on Vishnu Schist at the entrance to Granite Gorge below Hance Rapids.

The remaining Bass Formation, most of its thickness actually, is a complex unit dominated by dolomite. According to Timmons et al. (2012), it is likely the dolomite was originally limestone that underwent diagenesis, and was likely deposited in warm, shallow water. Within the dolomite are smaller amounts of arkose and sandy dolomite, characterized by intercalated shale and argillite (Hendricks and Stevenson, 2003). Other lithologies include intraformational breccias and conglomerates, stromatolites exhibiting an obviously laminated structure, and interbedded mudstones and sandstones (Timmons et al., 2012). The sedimentary features within the Bass Formation, including symmetrical ripple marks, desiccation cracks, intraformational breccias and conglomerates, normal and reversed small-scale, and graded beds involving the stromatolites, indicate a relative low-energy inter- to supratidal environment. The eastern transgression of the sea is inferred from the western accumulation of the carbonates and deep-water mudstones, and the eastern formation of stromatolites and accumulation of the shallow-water mudstones.

Mudcracks, ripple marks, and oxidized shales suggest subaerial exposure associated with periodic marine regression (Hendricks and Stevenson, 2003). Timmons et al. (2012) describes a relatively low-energy, tidal-dominated environment during the later time of the Bass Formation that resulted in an influx of mudstone and sandstone that transitioned into Hakatai Shale. Hendricks and Stevenson (2003) indicate a deltaic environment as the transition into the Hakatai Shale, with gradational contact in the east and a sharp, but conformable, contact in the west.

Hakatai Shale

The second rock unit of the Unkar Group is the Hakatai Shale (Figure 4). Possibly the most colorful formation in the Grand Canyon, the pervasive oxidation of the iron-bearing minerals of its mudstones present a mixture of purple to red to rich orange colors which are vibrantly displayed in Red Canyon (Hendricks and Stevenson, 2003). Figure 5 shows the Hakatai outcropping as a slope-forming unit just above the Bass Formation; its down-to-the east strata are dipping upriver in the direction of the Butte Fault which bounds the eastern edge of the graben in which the unit is preserved. The best exposures of the Hakatai Shale are along the Colorado River just above Hance Rapids (between river miles 76 and 77), in lower Red Canyon on the New Hance Trail, and along the lower part of the South Kaibab Trail. The thickness of the formation varies from 445 feet at Hance Rapids, but increases to 985 feet at Hakatai Canyon on the Shinumo Creek drainage. The Hakatai Shale is informally divided into three members. The lithology of the lower two are fractured clay-rich mudstones and shales of gentle-to-moderate, granular slopes, indicating a low-energy, mud flat environment. The upper member is composed of medium-grained quartz sandstone of ledgy, cliff-forming beds, suggestive of a higher energy, shallow-marine environment (Hendricks and Stevenson, 2003).

Sedimentary structures including mud cracks, ripple marks, tabular-planar cross bedding, salt casts and tool marks suggest a dominantly shallow-water depositional setting, likely a marginal marine environment. Another unusual feature preserved within eastern exposures of the Hakatai Shale are resistant sandstone columns within the lower two members. A comparison of mudstone-sandstone couplets within the formation indicates that while the contact between sandstone and mudstone beds is sharp, it is often irregular, suggesting that the sediments were not fully lithified but still soft and pliable when disturbed by tectonic activity associated with a series of northwest-trending, high-angle, reverse faults during the deposition of the Hakatai Shale (Hendricks and Stevenson, 2003; Timmons et al., 2012). The sandstone columns formed when fluidized sands were partially injected upward into overlying muds. An unconformable boundary occurs between the Hakatai Shale and the Shinumo Sandstone. The abrupt contact is very evident as the unconformity truncates the cross beds and channel deposits of the Hakatai Shale, with a relief less than 35 feet; and it has been interpreted to represent subaerial exposure and erosion during a significant marine regression (Hendricks and Stevenson, 2003; Timmons et al., 2012).

Shinumo Sandstone

The Shinumo Sandstone forms the third formation of the Unkar Group (Figure 4). It is predominantly composed of quartz arenite (having few impurities), but subarkose (quartz intermixed with feldspar) increases in the lower parts of the formation (Timmons et al., 2012).

This rock unit forms massive cliffs of red to brown to purple, but predominantly white to tan color, nicely observed in Figure 5 where they follow the same upcanyon dip of other Unkar Group strata related to their preservation in the Butte Fault graben. The thickness of the cliffs is fairly uniform, but increases to the west to 1328 feet at Shinumo Creek from its first appearance at 1132 feet near Papago Creek in the east. Hendricks and Stevenson (2003) describe four informal members for the Shinumo Sandstone. The lowest member is composed of a subarkosic conglomerate and submature quartz sandstone. The purity of the quartzite increases upward in the formation as the second member is a mature quartz sandstone. The lithology of the third member is a brown quartz sandstone with an abundance of cross bed, clay gall, and mudcrack structures. The final, and thickest member, is comprised of fine-grained, well-sorted, and rounded quartz grains held together by a siliceous cement (Hendricks and Stevenson, 2003). The deposition of the sandstone is inferred to have occurred in a very shallow, near-shore, marginal marine to partially fluvial and partially deltaic environment likely associated with a widespread though gently flucuating marine transgression. Its contact with the Dox Formation is conformable and marked by interbedding with the mud-rich sediments of the lower member of the Dox Formation.

Two unusual features, one sedimentological and the other geomorphological, mark the Shinumo Sandstone and are worth mention. The abundance of thick beds contorted by fluid evulsion in the upper part of its final member, representative of the mobilization of water in saturated sandstone by earthquake tremors, is suggestive of tectonic activity, but the specific faults relating to these features remain unidentified. However, this speculation is supported by similar contorted beds occurring in the Apache Group of Central Arizona, and even led to the credence of more widespread seismic activity (Timmons et al., 2012). The Shinumo Sandstone is a very resistant rock unit owning to its quartz purity, resulting in the formation of hills during pre-Tapeats subaerial exposure and erosional events from roughly 740 million to 545 million years ago. Subsequent marine invasion during the Middle Cambrian created temporary “islands” as sea level rose to eventually inundate them, a feature best observed where the Tapeats Sandstone and Bright Angel Shale onlap their margins, only to cover them with time and enough sea level rise.

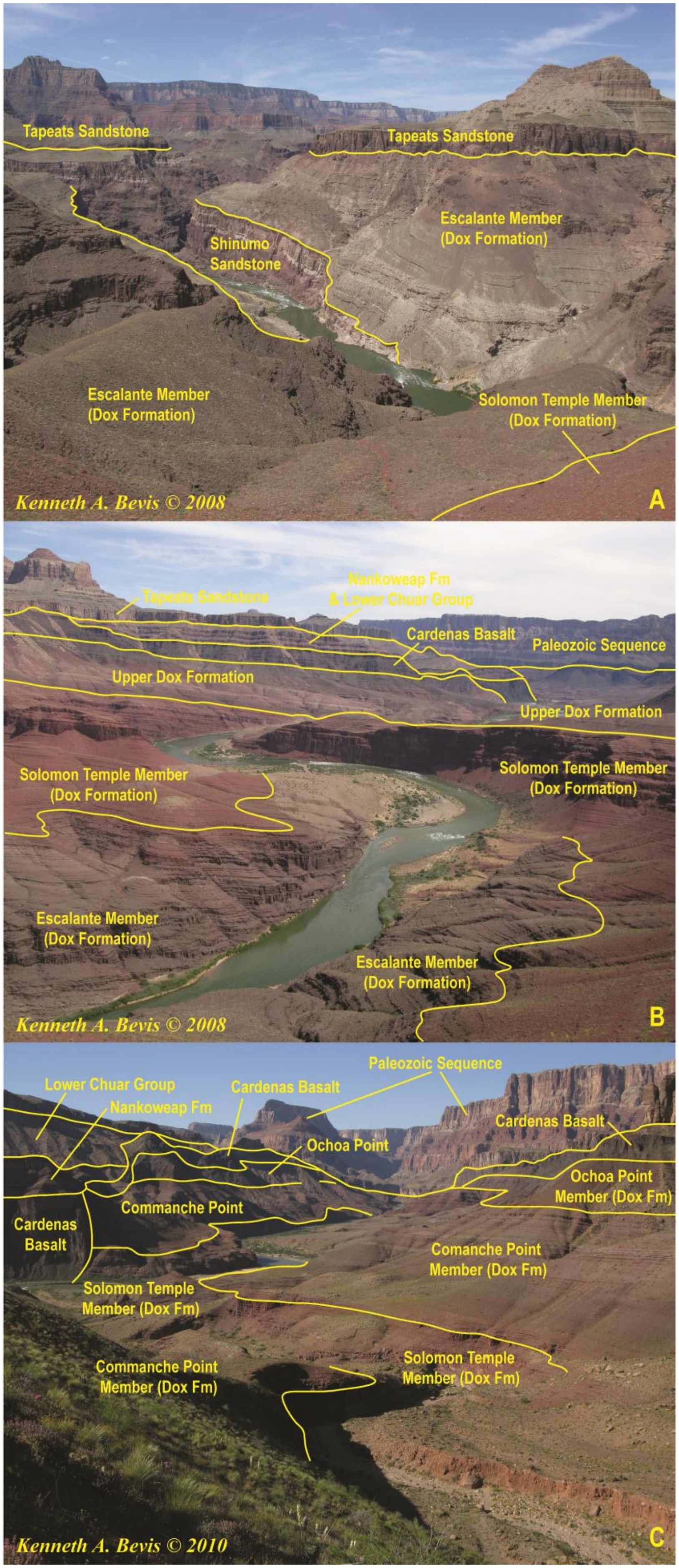
Dox Formation

The Dox Formation is the fourth, and by far the thickest unit of the Unkar Group (Figure 4). It is composed of four members: in ascending order, the Escalante Creek Member, the Solomon Temple Member, the Comanche Point Member, and the Ochoa Point Member. The lower Escalante Creek and Solomon Temple Members are preserved within each outcropping wedge of the Supergroup; however, the complete formation is exposed only in the easternmost and largest of the Late Proterozoic grabens, the one associated with the Butte Fault (Hendricks and Stevenson, 2003). Its impressive thickness is best exposed along the Colorado River between Palisades Canyon and Escalante Canyon (between river miles 65 and 75) within the Butte Fault graben. Contacts between the members are gradational and only recognized by topographic expression, depositional environment, and color variation.

The basal layer of the Dox Formation, the Escalante Creek Member, represents a dramatic break in depositional environment, containing a greater amount of feldspar and mica than any other layer in the Unkar Group, and it is the most immature of the upper sandstone units (Timmons et al., 2012). In the eastern Grand Canyon, this member reaches a thickness of 1280 feet and forms a cliff-slope topography in the vicinity of Escalante Creek at river mile 75 (Figure 6a). It has a light-tan to greenish-brown to grayish color, contrasting the red and red-brown color of the remaining members of the Dox Formation. The lower 800 feet of the member is a siliceous quartz sandstone and calcareous lithic and arkosic sandstone, while the upper 400 feet is a dark-brown-to-green-gray shale and mudstone. Sedimentary structures within the member include contorted bedding within the lower 100 feet, small-scale, tabular-planar cross beds, and graded beds with shale interclasts at the base (Hendricks and Stevenson, 2003).

The Solomon Temple Member stands out from the Escalante Creek due to its red-orange color and rounded-hill and slope-forming topography, both of which are more characteristic of the remaining members of the Dox Formation . The member is well exposed in the vicinity of Unkar Creek at river mile 73 (Figure 6b). Its lithology repeats in a cyclical pattern of red mudstone, siltstone, and quartz sandstone. It is 920 feet thick in the eastern Grand Canyon. The lower 700 feet is composed of shaley siltstone and mudstone with subordinate quartz sandstone, exhibiting a red-to-maroon color, and forming slopes. The oxidized maroon quartz sandstone combined with multiple channel features and low-angle, tabular cross beds, dominates the upper 220 feet of the member and suggests its accumulation in a floodplain environment (Hendricks and Stevenson, 2003).

The third and fourth members of the Dox Formation are the Comanche Point Member and the Ochoa Point Member, respectively. These two members do not exist west of 75-mile Creek due to pre-Tapeats erosion of preserved fault-bounded wedges. Both units crop out quite nicely in the vicinity of Commanche and Tanner Creek near river mile 68, and are observed especially well on the southeast side of the Colorado River where deformation form the Butte Fault is less significant (Figure 6c). The lithology of the Comanche Point Member is dominated by shaley siltstone and mudstone, but sandstone also occurs. The unit ranges from 425 feet to 617 feet in eastern Grand Canyon and is distinguished by its slope forming stratigraphy and variegated color. Pale green-to-white, leached red beds associated with stromatolitic dolomite layers occur up to 50 feet thick within it. Sedimentary features include ripple marks, mudcracks and curls, salt casts, and wavy, irregular bedding (Hendricks and Stevenson, 2003). The final unit of the Dox Formation, the Ochoa Point Member, underlies the Cardenas Basalt, ranges from 175 feet to 300 feet thick, and forms steep slopes and cliffs. Micaceous mudstone dominates the lower portion of the member, but grades upward to chiefly red quartzose and silty sandstone. Salt crystal casts are found in the mudstone, while asymmetrical ripple marks and small-scale cross beds occur in the sandstones (Hendricks and Stevenson, 2003).


Tapeats Sandstone
Tapeats Sandstone
Escalante Member
(Dox Formation)
Shinumo
Sandstone
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(Dox Formation)
Solomon Temple Member
(Dox Formation)
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A
Tapeats Sandstone
Nankoweap Fm
& Lower Chuar Group
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Solomon Temple Member
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Escalante Member
(Dox Formation)
Escalante Member
(Dox Formation)
Kenneth A. Bevis © 2008
B
Paleozoic Sequence
Lower Chuar Group
Cardenas Basalt
Nankoweap Fm
Cardenas Basalt
Ochoa Point
Ochoa Point
Member (Dox Fm)
Commanche Point
Cardenas
Basalt
Comanche Point
Member (Dox Fm)
Solomon Temple
Member (Dox Fm)
Commanche Point
Member (Dox Fm)
Solomon Temple
Member (Dox Fm)
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C

Figure 6. The massive Dox Formation of the Unkar Group crops out as a series of eastward dipping mudstones and sandstones thousands of feet thick between river miles 65 and 75 along the Colorado; (A) highlights the lowest Escalante Creek Member, (B) highlights the Solomon Temple Member, and (C) highlights the Comanche Point and Ochoa Point Members.

Overall, the variations in the strata of the Dox Formation members are very subtle and suggestive of gradually fluctuating sea levels along a low relief coastline. Timmons et al. (2012) describes wide, shallow river channels at the base of the Dox Formation. Best exposed in the lower two members, younger, mud-filled channels are cut through stacked, fine-grained sandstone channels to create cut-and-fill channel structures at Carbon Creek and at Unkar Rapids (river mile 73), indicating an estuarine environment where sea level fell and subsequently rose. The presence of sandstone steadily decreases as the formation develops. The Escalante Creek Member is characterized by large sandstone channels and accounts for the rapid transgression of the sea and filling in of the basin by the end of this member's time. The remaining members were probably deposited at or near sea level. The size of the channels decrease, as a braided stream or sheet-flow deposit would, during the time of the Solomon Temple Member, suggesting a fandelta as the depositional environment. A floodplain is credited for the mudstone and thin-bedded wave and current-rippled sandstones of the Comanche Point Member. The evidence of oscillating currents, mud drapes, and mud cracks indicate a tidal environment for the deposition of the final member of the Dox Formation, the Ochoa Point Member (Hendricks and Stevenson, 2003; Timmons et al., 2012).

The contact between the Dox Formation and the overlying Cardenas Basalt is conformable and even interfingering in some locations. Evidence proves that the sediments of the Dox Formation were still depositing when the first eruption occurred. In places, Dox sediments are mildly baked at contact with Cardenas lava flows; and there are thin, discontinuous deposits of basaltic lavas in the upper Dox Formation, including small folds and convolutions suggestive of soft sediment deformation. And quite uniquely, there is even a rounded mass of igneous rock (less than 3.3 feet in diameter) entirely covered in a thin layer of siltstone of Dox lithology and set in the lowest basaltic flow (Hendricks and Stevenson, 2003).

Cardenas Basalt

The final rock unit of the Unkar Group is the Cardenas Basalt (Figure 4). It is exposed only in the eastern Grand Canyon, composed of basaltic and basaltic andesite lava flows with sandstone interbeds, and with a varying thickness of 785 feet to 985 feet. The basalts are well exposed in the Tanner Canyon area and crop out quite nicely in the vicinity of Comanche and Tanner Creek near river mile 68, and can be readily observed making up the main cliff face of the Tanner Graben where it has been bisected by the Colorado River at Tanner Rapids (Figure 7). This formation is divided into two informal members. The lower unit, often referred to as the "bottle-green member," forms granular slopes and its thickness varies from 245 feet to 295 feet. Thin, discontinuous flows, sandstone interbeds, and broken basalt weathered to nodules are preserved within this unit despite its extensive weathering. Roughly two-hundred and thirty feet above the base of the Cardenas Lava, the basalt becomes more massive and less altered. A high sodium and magnesium content combined with depletion of potassium, all indicate a spilitic alteration which could have occurred with rapid quenching in the sea or brackish water. The sandstone interbeds likely occurred during periods of volcanic inactivity involving flowing or ponded water on the surface of the lava (Hendricks and Stevenson, 2003).

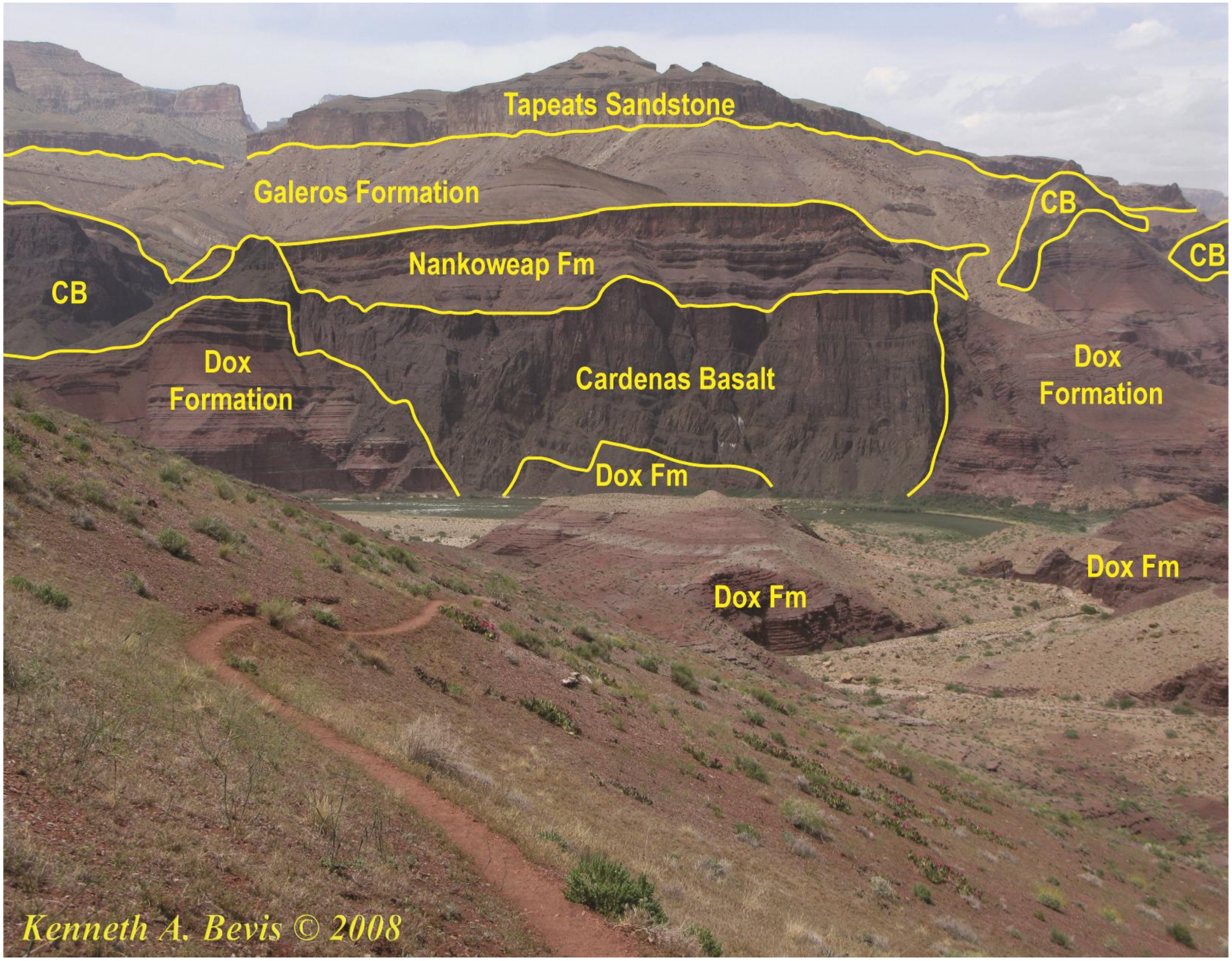

Figure 7. The Tanner Graben was formed in association with extensional movement the Butte Fault and can be observed in the cliffs along the Colorado River opposite the mouth of Tanner Canyon; Supergroup rocks exposed within the graben include the basal Dox Formation, the Cardenas Basalt, the Nankoweap Formation, and the capping Galeros Formation.

Approximately 328 feet above the base, a 16-foot-thick bed of continuous sandstone displaying lamination and forming very steep cliffs sits atop the bottle-green member of the Cardenas Basalt. In some areas, it occupies channels (most likely lava channels) that were carved into the lower member. This sandstone layer would likely have formed under the same conceptual model as the sandstone interbeds, but simply during a greater period of inactivity that provided sufficient time for the basin to subside until the lava surface dropped below sea level (Hendricks and Stevenson, 2003).

The upper member of the Cardenas Basalt is a cliff-forming basaltic and basaltic andesite lava flow sequence, with infrequent sandstone interbeds. The succession of features in individual flow units have led Hendricks and Stevenson (2003) to believe that the volcanic pile accumulated more quickly than the basin could subside to accommodate it. The evidence begins with an autoclastic breccia directly above the continuous layer of sandstone and is followed by a fan-jointed unit, ropy lava, and finally a lapillite unit at the 754-foot level. On top of the lapillite lava, a continuous sandstone layer sits upon a planar surface. This was interpreted as the result of volcanic activity ceasing temporarily following generation of the lapillite event, smoothing of the surface by erosion, and subsequent subsidence of the igneous rocks. This process prompted by the temporary cessation of volcanic activity is believed to have been repeated at least two more times within the upper member. Overall, the Cardenas Basalt was erupted in phases, allowing time for the deposition of interbedded sandstones. Eventually, volcanic activity concluded, the Unkar Group was tilted gently to the northeast (possibly due to tectonic movement along the Butte Fault), and an unknown amount of the Cardenas Basalt was eroded during its subaerial exposure before the deposition of the Nankoweap Formation commenced (Hendricks and Stevenson, 2003).

Igneous Intrusions

Inclusive to all of the formations of the Unkar Group below the Cardenas Basalt are igneous intrusions, specifically diabase sills and dikes. Figure 8 shows the spectacular Hance Dike intruding Hakatai Shale at Hance Rapids (river mile 77), one of many such intrusions exposed in and near Red Canyon. The diabase sills occur only in the lower two formations, the Bass Formation and Hakatai Shale, while the dikes occur above the sills both within the Hakatai Shale, and in the overlying Shinumo Sandstone and Dox Formation. The sills have been measured at thicknesses ranging from 75 feet in Hance Rapids to 985 feet in Hakatai Canyon. The fine-grained, chilled margins of the sills indicate that the magma was highly fluid and very hot (upwards of 2200 °C) when it intruded into the sedimentary rocks. Other alterations to the sedimentary rocks also occurred. Above the sills in the Bass Formation, contact metamorphism of the dolomite formed chrysotile asbestos, and adjacent to the sills in the Hakatai Shale, hornfels containing porphyroblasts of adalusite and cordiereite were altered to muscovite and green chlorite (Hendricks and Stevenson, 2003).

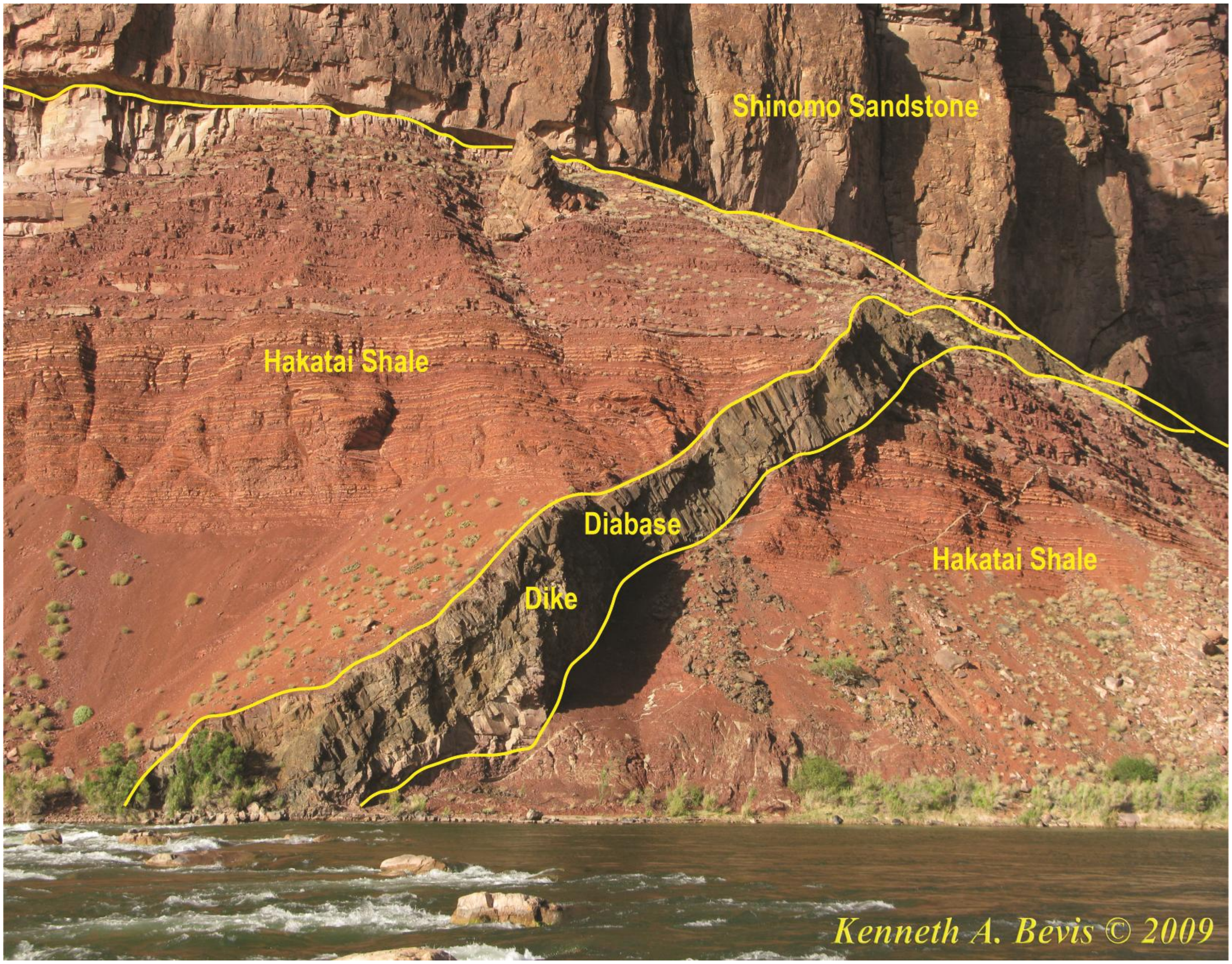

Figure 8. The Hance Dike intrudes the Hakatai Shale across the river from the mouth of Red Canyon; this gorgeous dike and its many companions intruded rocks of the Unkar Group and probably served as the conduit system feeding into the flood basalts that produced the overlying Cardenas Basalt.

Hendricks and Stevenson (2003) believe that the relationships between the sills, dikes, and Cardenas lava flows reflect a single, albeit prolonged, volcanic event. Mineralogy of the sills is uniform throughout the area and identical to the unaltered flow of the bottle-green member of the Cardenas. While sills do tend to be less felsic than the flows, this factor does not eliminate the possibly of a common parentage. Paloemagnetic evidence dates the intrusion of the majority of the sills to the same time as the deposition of the Comanche Point Member of the Dox Formation. However, the easternmost Unkar Group sill, the Hance Sill, could potentially be of a greater age. This information tells us that as much as 330 feet of sediment was deposited from the time the sills intruded into the lower formations of the Unkar Group and the initial eruption of lava flows associated with the Cardenas Basalt.

Geologic History

Hendricks and Stevenson (2003) credit the beginning of the Unkar time and the deposition of the Bass Formation to an initial marine transgression. The clast size variation of the Bass sediments and composition foreign to local crystalline basement rocks point to an easterly source beyond the Grand Canyon region. The ocean, which was located to the west in Unkar time, transgressed eastward across a gently undulating terrain as far as the eastern Grand Canyon, but likely much further. The eastern deposits indicate an intertidal-to-supratidal environment while the sediments in the middle of the Bass Formation in the western Grand Canyon indicate the terrain was below wave base. A gradual regression followed as a result of the Bass accumulation, and sediments from the upper Bass Formation and Hakatai Shale indicate a variation between subaerial and subaqueous environments. Tectonic activity along northeast-trending, high-angle reverse faults mark the end of Hakatai Shale deposition and ensuing uplift and/or marine regression resulted in a period of erosion prior to the deposition of the next formation (Hendricks and Stevenson, 2003).

Near-shore, fluvial, and deltaic conditions denote continued subsidence and a second major marine transgression during the accumulation of the Shinumo Sandstone and the early member of the Dox Formation. After 165 feet of Dox Formation sediment was deposited, the sea advanced eastward as the basin rapidly subsided and resulted in a sustained period of basin sedimentation and infilling. Minor erosion suggests that subaerial conditions had returned by the close of the Solomon Temple Member, and its contact with the Comanche Point Member indicates a transition to marine conditions once again. This fluctuation between marine and nonmarine environments continued throughout the remaining members of the Dox Formation. Basalts of the Cardenas Lava were initially deposited onto wet, shallow-water Dox sediments. The remainder of its deposition altered between marine and nonmarine environments with the sporadic accumulation of the lava and continued subsidence of the land, though the lava flows eventually accumulated more rapidly. Following the extrusion of more than 985 feet of lava, the area experienced tectonic uplift, and the Unkar Group was gently tilted toward the northeast, subaerially exposed and eroded an unknown amount, and new sediments of the Nankoweap Formation were deposited (Hendricks and Stevenson, 2003).

The Unkar Group records two types of faulting in response to plate tectonic forces (Figure 2). The first is reverse faulting, where one side of the fault plane is pushed upward relative to the other. Reverse faults typically record the horizontal shortening or contracting of the Earth's crust, suggesting it was squeezed perpendicular to the fault trace. Oftentimes, as the one side of the fault plane is pushed upward over the other, the strata on the opposite side of the fault plane is warped and folded into a monocline. Timmons et al. (2012) indicate that monoclinal folding in the lower Unkar is associated with deformation of rock layers over northeast trending reverse faults. Monoclines are of a small scale and usually die out or are eroded and that no northeast trending monoclines occur above the Shinumo Sandstone (or in the Chuar Group), indicating formation endured only in early Unkar time. Of the Proterozoic monoclines, all are northeast trending and record northwest-directed shortening of the crust, most likely in response to the forces of plate tectonics acting on Laurentia (Timmons et al., 2012). The second type of faulting is more pervasive within the Supergroup; normal faulting that records regional crustal extension perpendicular to the trace of the faults. Normal faults occur within all layers of the Unkar Group and continue into younger Supergroup rocks above. They are generally northwest-southeast trending and could possibly dip northeast or southwest in larger structures. Synclinal folding higher in the section suggests that extension of the crust and the deposition of the sediments were simultaneous (Timmons et al., 2012). These structures will be discussed in greater detail later.

In the past, geologists have had the difficult task of basing regional correlations solely on the lithology of units. Using this method, Hendricks and Stevenson (2003) suspected a correlation between the Unkar Group and the Apache Group of central Arizona. Both groups were unmetamorphosed except when in contact with igneous intrusions and had an age younger than the basement rocks that they rested on, and yet were older than Cambrian sediments. Based on paleomagnetic data, they suspected that the Mescal Limestone of the Apache Group was correlative with the middle members of the Dox Formation (Hendricks and Stevenson, 2003). However, the recent work of Timmons et al. (2012) with detrital zircon has proven that the Apache Group is too old to have any relation with the Unkar Group. Using detrital zircon-derived sediment ages matched with the inferred age of potential sourcelands, Timmons et al. (2012) was able to suggest that the Grenville Orogeny served as a strong source for sediments within the Unkar Group. Chemical analysis of the Unkar Group shows that its sediments have been moderately weathered, indicating a temperate climate and rapid transportation, information used by Timmons et al. (2012) to suggest that these sediments likely travelled westward from mountainous highlands in the southeast via a large river system. Timmons et al. (2012) believe that the Hazel Formation in west Texas, a coarse apron of sediments that records an impressive mountain building event, was likely transported from the Grenville Mountains, and thus correlates to the Dox Formation. The Hazel Formation records an impressive mountain building event referred to as the Grenville Orogeny. Evidence stretches from the southwest to the northeast of the United States, as well as on every current continent. This impressive continental plate collision resulting in the assemblage of the supercontinent Rodinia and the Grenville Orogeny occurred between 1250 to 1000 Ma, and its deconstruction from about 750 to 550 Ma (Timmons et al., 2012).

Summary

The Grenville Orogeny and the assemblage of the supercontinent Rodinia played a substantial role on the deposition of the Unkar Group. Evidence for this impressive continental to continental plate collision stretches from the southwest to the northeast of the United States; its formation occurred between 1250 to 1000 Ma, and its deconstruction from about 750 to 550 Ma (Timmons et al., 2012). The Unkar sediments accumulated in a basin on the western edge of the North American continent far from the epicenter of mountain building. A high percentage of sediments were transported from the continental scale Grenville Mountains and deposited in this basin as the sea experienced an overall eastern transgression punctuated by minor variations in sea level due to subsidence and basin filling. Fifty eight-hundred feet of sediment accumulated before the Cardenas lavas erupted onto the wet surface of the Dox Formation, adding more than 985 feet of igneous rock. After the eruptions ceased, the Unkar rocks were tilted slightly northeast, eroded, and the deposition of the overlying Nankoweap Formation commenced (Hendricks and Stevenson, 2003; Timmons et al., 2012).

The Nankoweap Formation (by Hannah Slover and Ken Bevis)

The Neoproterozoic Nankoweap Formation is located in the middle of the Supergroup, between the Unkar and the Chuar Group (Figure 4), and its contact with the slightly tilted strata of the Cardenas Basalt is unconformable. It was initially included as part of the upper Unkar Group and lower Chuar Group by Walcott (1894), but was later separated into its own formation by Van Grundy (1937). While the formation is named for its small exposure in Nankoweap Canyon, more extensive outcroppings occur in Basalt Canyon, Comanche Creek, and Tanner Canyon. A total of 370 feet thick, the Nankoweap Formation is composed of red-brown and tan sandstones with a subordinate amount of siltstones and mudstones (Hendricks and Stevenson, 2003); its strata overlie the Cardenas Basalt within the Tanner Graben exposed in the riverside cliffs at Tanner Rapids (Figure 7). The formation's base and top are erosional disconformities, and thus, it is incomplete; although its very incompleteness suggests that it formed during a period dominated by erosional forces. Its exact age is unknown though detrital zircon has suggested it is closer in age to the Chuar Group, having formed about 900 million years ago during a roughly 300 million year interval that separates the Unkar Group from the Chuar Group (Timmons et al., 2012). Two informally named members comprising the Nankoweap Formation are referred to by Timmons et al. (2012) as the lower red unit and the upper white unit. The lower red unit includes 40 feet of ferruginous, fine-grained quartzitic sandstones and siltstones. These strata are characterized by hematitic laminae and lenses of volcanic detritus from the underlying Cardenas Basalt. The 330 feet of the upper white unit rests disconformably upon the lower informal member. This upper unit is composed of fine-grained, thin-to-medium bedded sandstones, with an increasing presence of siltstone towards the top. Sedimentary structures include cross-beds, ripplemarks, mudcracks, soft-sediment deformation features, and rare salt pseudomorphs (Hendricks and Stevenson, 2003).

Timmons et al. (2012) report a hiatus between the two informal members that allowed lag deposits comprised of Cardenas Basalt clasts to accumulate during faulting and erosional activity, and a capping layer of white, fine-grained quartz-cemented quartz arenite. While it is doubted by subsequent research, Hendricks and Stevenson (2003) describe a trace fossil impression of what appears to be a stranded jellyfish. The structure, a total 5 inches in diameter, contains a series of lobes that are rounded at the extremities. Another explanation for this strange impression is a sand-blow or sand-volcano, which are formed by the upward expulsion of gas or fluids from sediments during seismic activity. Opinions remain conflicted on the origin of this specimen, but if it is a trace fossil, it would be the first record of complex life on earth. Overall, the lower formation is inferred to have been deposited in shallow water subject to periodic drying, and probably represents deposition within structurally controlled ponds or lakes. The upper member shows an increase in energy (moderate to low), but still in relatively shallow water, suggesting a possible marine incursion or deposition in a larger, deeper lake environment (Hendricks and Stevenson, 2003).

The Chuar Group (by Hannah Slover and Ken Bevis)

The Chuar Group is mid-Neoproterozoic, accumulated between 800 to 742 Ma as determined by U-Pb zircon dates. Exposed only in the eastern Grand Canyon, these deposits form the upper strata of an entire package of Supergroup rocks contained within a massive graben bounded in the east by the Butte Fault system (Figure 3) and truncated at the top by the Great Unconformity and overlying Tapeats Sandstone. The sequence displays Martian-like colors and the entirety of the group is approximately 6800 feet thick; although thickness varies east-west across the north-trending Chuar Syncline which parallels the Butte Fault since the sediments were deposited as the syncline developed (Dehler et al., 2012; Ford and Dehler, 2003). Dehler et al. (2012) reports this group to be nearly 85 percent mudrock, with interruptions of meter-thick sandstone and dolomite beds. The strata are fossiliferous, unmetamorphosed, and the contacts between formations are gradational and determined by the presence or absence of the carbonate beds (Ford and Dehler, 2003). Sedimentological evidence indicates that Chuar deposition occurred near the equator in a seismically active basin that experienced a pattern of slow sea level rise and fall as the supercontinent Rodinia began to separate (Dehler et al., 2012).

The Chuar Group is composed of two formations: the Galeros Formation and the Kwagunt Formation (Figure 4). The Galeros Formation includes, in ascending order, the Tanner Member, Jupiter Member, Carbon Canyon Member, and Duppa Member. The Kwagunt Formation overlies the Galeros Formation and includes, also in ascending order, the Carbon Butte Member, Awatubi Member, and Walcott Member. The Chuar Group is overlain by the final formation of the Supergroup, the Sixtymile Formation. This portion of the Grand Canyon Supergroup has not experienced extensive alterations to nomenclature suffered by the Unkar Group. However, the Chuar Group did originally include part of the Nankoweap Formation, the Tanner Member, and the Sixtymile Formation.

Galeros Formation

The lowermost member of the Galeros Formation is the Tanner Member. In Basalt Canyon, it forms a massive sloping ledge to cap the Nankoweap cliffs below; and it forms the cap rock within the Tanner Graben at Tanner Rapids (Figure 7). The Tanner Member is comprised of two different lithologies. The basal layer of this member fills in depressions carved into the upper Nankoweap Formation with 20 to 50 feet of thickly bedded, coarsely to finely crystalline dolomite. Parallel horizontal laminations and intraclast horizons are present within the dolomite. The upper 580 feet of the Tanner Member are predominately shales with subordinate siltstones, sandstones, and dolomites. The shales are predominately black, but weather to multihued shades of ocherous yellow, orange, red-purple, and pale green and gray. Finely laminated to massive, the shales also contain very thin to thin lenses and tabular beds of white to green siltstone and fine-grained sandstone. Hematitic cements that may weather to goethite are common, and within the upper 160 feet of shale, the small, circular fossil, *Chuaria circularis*, thought to be an algal-like organism is often present. Subordinate sandstone and dolomite beds increase in thickness towards the top of the member. The sandstones are green, fine-grained, and thin to thickly bedded. They reveal rare ripple marks and mudcrack casts. The dolomite beds are massive, but only occur in the upper two meters of the Tanner Member (Ford and Dehler, 2003).

The second member of the Galeros Formation, the Jupiter Member, repeats the cycle of the Tanner Member: carbonates below and shales above. The Jupiter member forms roughly the upper half of the Galeros rocks visible in Figure 7, and directly underlies the Tapeats Sandstone. The carbonates of the Jupiter Member occur in the basal 40 feet as stromatolitic limestones and dolomites. The upper part of these carbonates are also layered; they have an abundance of gypsum crystal casts, and some poorly defined and solitary stromatolite columns similar to the forms *Inzeria* and *Stratifera*. The upper, predominantly shale layer is 1516 feet thick and varies in color from red-purple to ocherous yellow to pale green to blue-black. The shales are often micaceous and within the black shales occur the rare *Chuaria circularis* fossils. Thin beds of sandstone and siltstone are also common. These are rarely more than a few inches thick, have an abundance of symmetric ripple marks and mudcrack casts, display soft-sediment deformation features, ripple lamination, rare raindrop prints and salt pseuodomorphs (Ford and Dehler, 2003).

The third member, the Carbon Canyon Member, is well exposed in Nankoweap Canyon in the northeast corner of the Grand Canyon (reach by the Nankoweap Trail). It forms the base of Nankoweap Butte on the south side of Nankoweap Canyon (Figure 9). The unit differs from the lower two members in that it is characterized by a significant component of sandstone beds in addition to the presence of the carbonate and shale beds of the lower members. Sandstone layers are typically not more than a few feet thick, but form a grand total thickness of 1546 feet within the unit. The carbonate beds are 3 to 6 feet thick and are made almost entirely of dolomicrite to dolosiltite with local chert nodules and laminations of quartz siltstone common. In some locations, the carbonates grade into calcareous siltstones and may display irregular, or crinkly, laminations of a possible algal origin. Symmetric ripplemarks and mudcracks are often present on the tops and bottoms of beds. The transition from yellowish-tan crinkly laminated dolomites to reddish-orange massive dolomites in the carbonate beds indicates a shallowing-upward cycle. The interbedded shale beds vary from blue-black micaceous shale to red and green mudstone and suggest temporary shallowing events with greater input of clastic sediments.

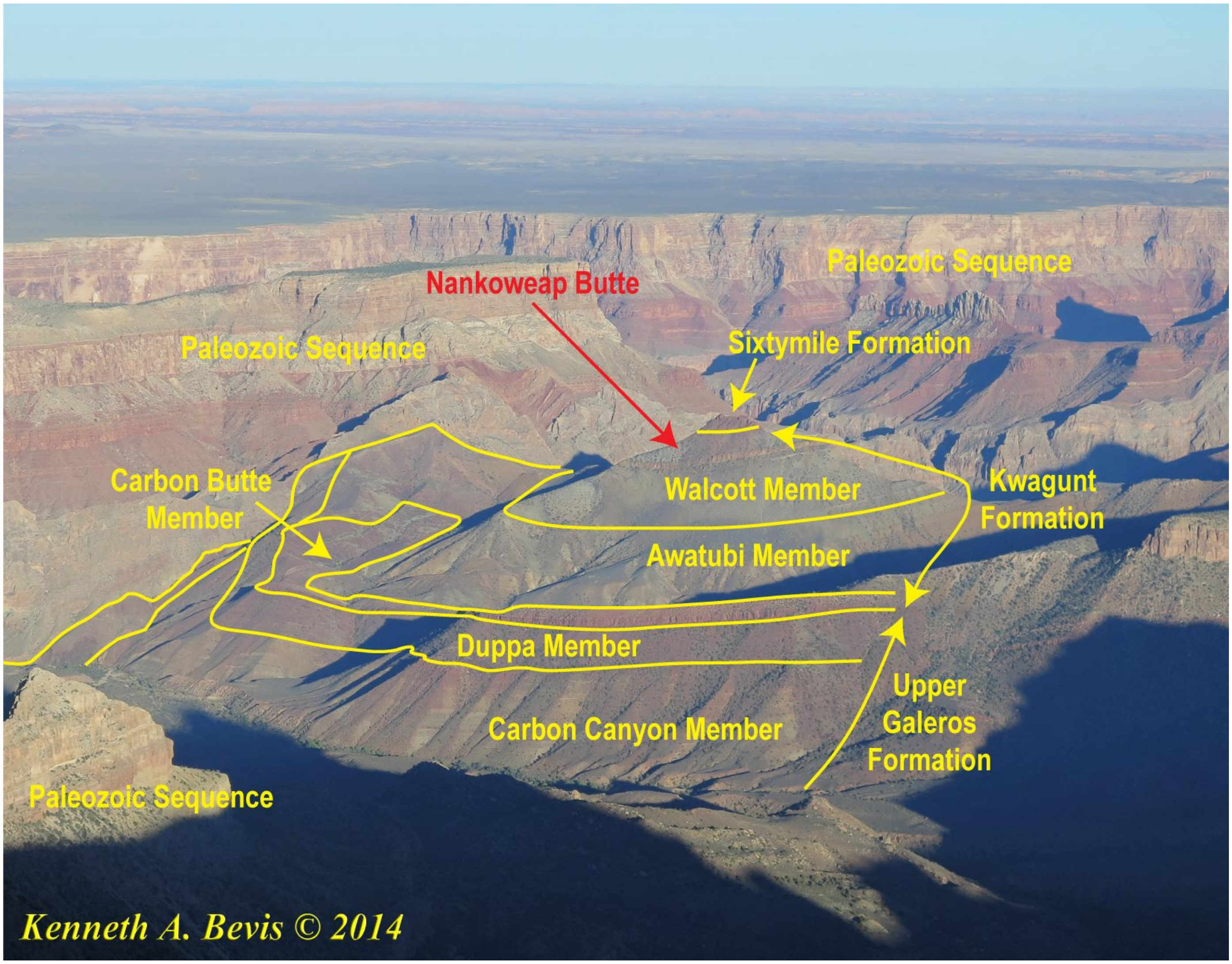

Figure 9. The north flank of Nankoweap Butte on the south side of Nankoweap Canyon exposes the upper Galeros Formation and the Kwagunt Formation of the Chuar Group, as well as the lower portion of the Sixtymile Formation; the view is from Point Imperial on the Grand Canyon's North Rim.

Rare sandstone beds of no more than 2 feet in thickness form the final lithology of this member. The sandstones have a green to gray to tan color. The subangular to rounded quartz grains are cemented by carbonates, silica, hematite, chlorite cement, or a clay matrix. Medium- to coarse-grained, well-rounded quartz sand is often randomly oriented within the fine-grained matrices. Mudcrack casts are a common sedimentary structure, but some laminae display truncated, insipient cracks that resemble worm tracks. Other features include symmetrical ripplemarks, interference ripples, ripple laminations, soft-sediment deformation features, low-angle planar crossbeds, and trough crossbeds. Stromatolites occur towards the top of this member and display strongly convex laminae, are sharply widening, and have irregularly branching columns. These columns are clustered into domes, 1 foot tall and usually under 2 feet feet in diameter but can reach 6 feet in diameter, taper at the base, usually closely spaced, and can grow into one another. The stromatolitic horizon displays little lateral variation and is often interbedded with black shales and intraclastic dolomite. While dolomization destroyed much of the detail, they appear to be of the form *Baicalia Semikhatov*. In the upper section of the member, a dolomite marker bed forms a distinctive horizon of potentially large-scale mudcracks forming polygonal patterns on the bedding planes and having undergone soft sediment deformation, indicating prolonged periods of subaerial exposure (Ford and Dehler, 2003).

The final and uppermost member of the Galeros Fromation is the shaley Duppa Member. This unit outcrops on the south side of Nankoweap Canyon above the Carbon Canyon Member in the flanks of Nankoweap Butte (Figure 9). More than 570 feet thick, shale is the dominant lithology with minimal thin siltstone beds. The siltstone beds occur throughout the member at a maximum of 3 feet thick and are composed of well-rounded silt grains. The shale beds are generally micaceous. Towards the top of the member, they grade into red mudstone and thinly bedded sandstones and siltstones. The Duppa Member has a gradational contact with the overlying Carbon Butte Member of the Kwagunt Formation (Ford and Dehler, 2003).

Ford and Dehler (2003) explain that the basal unit of the Galeros Formation, the Tanner Member, represents a sediment-starved basin that was rich in organic matter. Its laminated dolomite represents a shallow subtidal or intertidal environment, and the upper black shales indicate a deeper water environment with the occasional input of silts and sands during storms. The Jupiter Member appears to transition to a coastal or alluvial plain. A shallow subtidal to intertidal depositional environment is suggested by the lower *Inzeria* bed and the overlying variegated shales characterized by mudcracks and raindrop impressions indicate an intertidal to supratidal paleoenvironment. Sediment deposited in a mixed coastal or paludal swamp is likely responsible for the Carbon Canyon Member. Characteristics similar to the Tanner and Jupiter Members support a fluctuating subtidal to intertidal to supratidal depositional setting. Mudcracked, wave ripplemarks support fluctuation in currents in a tide- and wave-affected zone. The Duppa Member of the Galeros Formation, contains sedimentological evidence indicative of an alluvial plane. It is very similar to the Carbon Canyon Member except that the Duppa Member has less carbonate and sandstone beds, suggesting a deeper subtidal to supratidal environment (Ford and Dehler, 2003).

Kwagunt Formation

The Kwagunt Formation is the upper formation of the Chuar Group and its lowermost member is the Carbon Butte Member. The base of this member can be identified by the distinctive marker of the only thick sandstone layer of the entire Chuar Group. In Figure 9, the basal sandstone forms a ridge encircling the north and east side of Nankoweap Butte where it is folded upward as part of the syndepositional Chuar Syncline. At its type locality, the Carbon Butte Member is 252 feet thick, but varies across the Chuar Syncline, being thicker toward basin center. This thickness pattern is repeated by overlying layers of the Supergroup, indicating initial and protracted activity on the nearby Butte Fault. The lower 80 feet of the member is the thickly bedded sandstone which has weathered into a cliff. The strata is tan to red in color, fine- to medium-grained, and interbedded with shales and siltstones. Sedimentary structures include 3- to 6-feet thick cross-bed sets, symmetrical ripple marks, trough cross beds, ripple laminations, soft sediment deformation features, and mudcrack casts. Overlying the sandstones are 170 feet of mostly mudstones and shales, red to purple in color, interbedded with a smaller amount of thin to medium beds of fine- to medium-grained sandstone and siltstone. The uppermost part of the Carbon Butte Member is a 9-foot-thick unit of white sandstone. Within the sandstone are extremely well preserved symmetric ripplemarks, interference ripplemarks, soft sediment deformation features, and trough cross-beds (Ford and Dehler, 2003).

The middle member of the Kwagunt Formation, the Awatubi Member, is exposed in the central flanks of Nankoweap Butte, inside of the ringing ridge formed by the lower Carbon Butte member's sandstone bed (Figure 9). The common cycle of a carbonate base and overlying shales common to the Chuar Group returns for this middle member. The Awatubi Member is composed of 1128 feet of a stromatolitic carbonate base and then dominantly shales and mudstones. The carbonate unit is only 12 feet thick, and is characterized by biohermal domes (8 to 10 feet in diameter) that are made of complex columns (2 to 3 inches in diameter) and interbedded with confluent domes. These columns display almost perfectly flat laminae, and while dolomitization destroyed most of the detail, the form *Boxonia Koroljuk* is considered to be present. The matrix between the columns is usually crystalline dolomite while the matrix between the bioherms in coarsely granular dolomite. At the base of some of the bioherms, a flat-pebble conglomerate occurs. The upper shale unit is varying in color and is interbedded with thin to very thin beds of sandstone and siltstone. Common sedimentary structures include ripple laminations, symmetric ripplemarks, interference ripplemarks, horizontal and low-angle planar laminations, and mudcrack casts. Black, finely fissile shales, yielding an abundance of *Chuaria circularis*, occur on the eastern and western slopes of Nankoweap Butte beginning 30 feet from the top (Ford and Dehler, 2003).

The final member of the Kwagunt Formation and of the Chuar Group is the 838-foot-thick Walcott Member. This unit forms the upper flanks of Nankoweap Butte (Figure 9). Repeating a common depositional theme by now, a flaky dolomite layer occurs in the lower 12 to 32 feet of the Walcott Member, and is composed of oolitic and interclastic dolomite. Silicified and dolomitic laminations are folded, broken, and crinkly. Silty, intraclastic dolomicrite also characterizes the unit and is wavy to horizontally laminated. Following the dolomite layer, units occur, in ascending order, as black shales, silicified oolite and pistolite beds, and 3 distinctive carbonate beds. The black shales contain the fossil form *Chuaria circularis* and are interbedded with the oolite and jet black pistolite beds, usually 6 to 12 inches thick, with the exception of a 4- to 5-foot-thick white silicified oolitic bed. The pistolite beds often contain chert in place of the ooids and pisoids, and the outer surfaces may contain a mat of algal filaments or spheroidal bodies. Of the three dolomite layers, the two lowermost units are referred to as the "dolomite couplet." The lower couplet is an 8.5 to 12 foot thick package of sediment comprised of wavy to horizontally dominated dolomite, characterized by trough-cross-bedded oolitic and intraclastic dolomite, crinkly laminated, "cornflaky" algal stromatolite, and pink med-grained quartz sandstone. The upper couplet and middle dolomite layer is 31 to 38 feet thick of massive micrite to dolomicrite with rare mud chips, interbeds of black shale, and carbonate breccia zones. The final uppermost dolomite layer is referred to as the "karsted dolomite," and only occurs in Sixtymile Canyon. It is 40 feet thick and composed of crystalline dolomite with vugs, cavities, dissolution features, and brecciated dolomite and sandstone clasts in some cavities (Ford and Dehler, 2003).

In the Kwagunt Formation, the sedimentary structures of the Carbon Butte Member suggest a tide- and wave-affected shoreline, but also influxes of fluvial conditions (Ford and Dehler, 2003). The cross-bedding present in the unit records the opposing paleoflows of underwater dunes, a diagnostic tidal feature (Dehler et al., 2012). The sandstones of this member are dominantly coarse, clastic sediments that indicate an increase in supply of sediments or a significant increase in energy conditions. The Awatubi Member suggests an intertidal to shallow subtidal environment that deepened during its deposition. The basal bioherm unit represents a shallow subtidal to intertidal environment, the mudcrack casts of the symmetric ripple marked sandstones indicate a tide- and wave-affected environment, and the uppermost black shale (*Chuaria*-bearing) suggest a deeper water environment with an input of sediments through storms. The final Walcott Member represents a carbonate ramp. The black shales were deposited in a deep water environment, the water shallowed to a subtidal environment to deposit the oolite and pisolite beds, and finally, the upper carbonate unit was deposited during a transition to shallow subtidal, intertidal, and supratidal depositional environments. In general, the Unkar Group represents a time of quiet, nonturbulent embayment on a marine platform that fringed the paleocontinental west coast of North America. The coastal zone was influenced by both tidal and wave processes, affected by infrequent large storms, and permitted the deposition of mud and organic matter in quieter waters (Ford and Dehler, 2003; Dehler et al., 2012).

Geologic History

The paleontology recorded in the sediments of the Chuar Group help to indicate the variety of ancient depositional environments that these organisms once thrived in. Stromatolites are a very common feature, specifically in the Chuar dolomites. Today, stromatolite-forming algae are not as common, but examples exist in Shark Bay, Western Australia, and off the Baja Peninsula in the Gulf of California (Dehler et al., 2012). According to Ford and Dehler (2003), stromatolites grew in low-energy, shallow waters, with a gentle current and occasional periods of relatively higher energy. Intermittent desiccation also occurred and periods of low terrigenous sediment input allowed for a more successful growth. Dehler et al. (2012) concur, stating that stromatolites have to be submerged in clear water to grow, but not too deep because they need light for the microbes to photosynthesize. This knowledge can be used to predict water depth. One type of stromatolite is usually around 6.5 feet tall, so water depth had to be at least that deep, but no more than 328 feet at a maximum (Dehler et al., 2012).

Dehler et al. (2012) reports at least 6 different types of stromatolites in the Chuar Group. Their shape is strongly affected by the physical condition of their environment and the lamination reflects episodic growth. *Boxonia*, easily observed in the low hillsides past the Butte Fault in Kwagunt Canyon, looks like a giant brain. Another stromatolite form, *Baicalia*, appears in the Carbon Canyon Member and is less than 0.5 m all around. From the top, it appears to be a forest of large broccoli heads. When these broccoli heads are chaotically broken up, they indicate an environment in which they were reworked by storms or waves. Broken pockets within *Inzeria* and *Stratifera* also indicate the possibility of infrequent large storms. These stromatolites have a complex assemblage, from marble-sized to river-raft-sized, and are usually formed within one another, dome within dome. These particular forms are found in the base of the Jupiter Member (Dehler et al., 2012).

Also occurring in the Chuar Group are small, smooth, disc-like, organic-walled carbonaceous fossils called *Chuaria circularis*. A giant to the rest, sizes range from 70 µm to 5 mm and their fossils occur alone or in a group, never overlapping. They are hollow with a narrow marginal thickening and a wrinkled center (Ford and Dehler, 2003). Alive, they are thought to have been a smooth, featureless, plaktonic sphere (Dehler et al., 2012). Ford and Dehler (2003) state that this specimen is clearly an acritarch, related with other late Riphean to early Vendian acritarchs. However, other geologists have alternate interpretations of this fossil, including the possibility that it may be a brachiopod, gastropod, algal, a trilobite egg, or even inorganic (Ford and Dehler, 2003).

Found in the "flaky" dolomite and shale layers and the bacterial mats on the surface of stromatolites in the Walcott and Awatubi Members is a single-celled organism referred to as a vase-shaped microfossil or an amoebae (Dehler et al., 2012). Approximately 10,000 of these organisms occur in one cubic centimeter of shale (Ford and Dehler, 2003). This microscopic species moves about on a pseudopod (a finger-like extension of their cell) that extends from a hole in their test (protective house or shell). The tests, which look like tiny vases or bags with a small hole on one end, are abundantly preserved, while the cells decay. The amoebae deposited after the shale, likely promoting carbonate precipitation. Dehler et al. (2012) claims there are at least 11 speicies of amoebae, differing in shape of test, test opening, and presence of indentations or scales. The amoebae are best preserved by the billions at the top of the Walcott Member on Nankoweap Butte (Dehler et al., 2012).

The climate during Chuar time can be extrapolated from a variety of different clues within the strata, which are cyclic, indicating repetitions of environmental change. The majority, if not all, of the Chuar Group is commonly accepted to have formed as a nearshore marine environment populated primarily by single-celled organisms. Chuar Group rock units are laterally continuous and thus, facies changes occurred due to changes in water depth. The Chuar strata typically have carbonates at the base, originally deposited in shallow-water subtidal to intertidal, and even supratidal environments; and are capped by shales, often containing subordinate sandstones or dolomite. Shales are deposited in coastal zones below wave base, in quiet-water settings offshore as mud. Sandstones within the shales are interpreted as storm-wave reworking and winnowing of shales or an influx of sediments, both requiring higher energy conditions. However, the dolomite is precipitated as carbonates and requires a further shallowing of sea level. Repetition of this cycle of carbonates first, and shales next (with minor sandstones and/or dolomites) indicates significant sea level fluctuations corresponding to global-wide glaciation (falling sea level) and deglaciation (rising sea level) (Dehler et al., 2012).

Carbon-isotope signatures and shale composition paired with the stratigraphic data are able to provide a climate story for the Chuar Group. By dividing the sediments into four stratigraphic sequences of sandstone-rich and dolomite-rich cycles, geologists are able to track the changes in the carbon cycle and weathering rates of the shales. A carbon cycle ratio is approximated by two stable carbon isotopes preserved in the carbonate rocks and organic material of the shales. A more positive curve indicates an increase in primary productivity and organic carbon value. The Chuar Group exhibits four major excursions, and the lower Awatubi member records one of the largest fluctuations in the carbon-isotope curve ever recorded. Shale-weathering rates use the mineral compositions (kaolinite and feldspar) to determine the weathering rate and relative humidity of the source sediments. Examination of Chuar shale indicates less intense weathering during dolomite-rich times, and a lower sea level, and greater weathering during sandstone-rich times, and a higher sea level (Dehler et al., 2012).

By combining the information derived from the stratigraphic data, carbon-isotope curve, and shale-weathering rates, a climate scenario can be inferred. During the sandstone-rich intervals, a higher carbon-isotope value indicates an increase in organic-carbon burial and kaolinite-rich shale suggests intense weathering and a higher sea-level. Deposition of the sandstone cycles occurred during locally wetter and globally warmer times, when clastic sediments were delivered at a rapid rate, sea level was high, and glacial ice levels were low. On the contrary, during dolomite-rich intervals, lower carbon-isotope values indicate a decrease in the burial of organic-carbon and feldspar-rich shales propose less intense weathering and a lower sea level. This data describes a locally drier and globally cooler climate, with a decreased input of sediments, a lower sea-level, and more glacial ice than during carbonate-rich intervals (Dehler et al., 2012).

These conclusions are important in the study of the Neoproterozoic glacial deposits and their relation to the carbon-isotope excursions. Neoproterozoic glacial deposits are fascinating because they were deposited at sea level in equatorial regions, and are possibly associated with the significant excursions. A well-known hypothesis is the “snowball Earth,” the idea that all of the Earth’s oceans were entirely frozen for at least 10 million years. It is a possibility that all the Neoproterozoic carbon-isotope variations are related to glaciation, even if glacial deposits are absent. The Chuar Group has the excursions and no glacial deposits, but also independently suggests glaciation in its stratigraphic data sequences. While there are conflicting views on the “snowball Earth” hypothesis, the Chuar Group at least provides a timeline of glaciations, at least in the poles, between 800 and 742 Ma (Dehler et al., 2012).

While the Unkar Group was only marginally affected by post-depositional faulting and slight, northeastward tilting, the rocks of the Chuar Group were heavily influenced by syndepositional faulting related to growth of the Butte Fault system and accompanying synclinal folding (Figure 3). The Butte Fault is a major north- to northwest-trending normal fault recording a large west-side-down Neoproterozoic displacement (Dehler et al., 2012). This fault and its various splays are exposed for 18 km within the confines of the Grand Canyon, although the structure extends in the subsurface all the way across the Utah border and is expressed at the surface by associated monoclinal folding of Paleozoic and Mesozoic rocks. The majority of the subordinate normal faults have lesser displacements of only meters to tens of meters within the Chuar Group and are west-dipping and parallel the Butte Fault, but there are a few that dip to the east to form opposing sides of symmetrical grabens such as the Tanner Graben (Timmons et al., 2003). The Chuar syncline is a broad, asymmetric, trough-shaped fold of Chuar strata with a steeper dip on the eastern limb nearest the Butte Fault. The parallel trace of its axis to that of the Butte fault suggests a genetic relationship. The Tapeats Sandstone truncates the syncline, indicating that it is Neoproterozoic in age (Dehler et al., 2012).

An important relationship occurs between the Butte Fault, Chuar syncline, and the deposition of Chuar strata. It is believed that the upper strata of the Chuar Group experienced deposition synchronous with fault movement and synclinal development. Evidence for this resides within the sediments. The fact that there is a greater displacement of strata in the lower beds of the group indicates that the fault activated in the middle of Chuar deposition and remained active thereafter. The Butte Fault is also responsible for the absence of Chuar strata to the east of the fault. When it activated the eastern deposits were warped upward and subjected to erosion while the western deposits were lowered, thickened, and preserved. Finally, geologists believe the syncline formed during sediment deposition because the carbonate beds decrease in thickness on the eastern limb as it reaches the Butte Fault (Dehler et al., 2012). Water on this side of the synclinal basin would have been too shallow or absent to accumulate carbonates.

The Chuar Group began deposition just before or during the onset of low-latitude glaciation and during the early stages of rifting of the Rodinian supercontinent. Ford and Dehler (2003) suggest that this rifting could be the second recorded attempt to breakup Rodinia, their first "record" having been deleted by the erosion surrounding the Nankoweap Formation. Evidence of the rifting of Rodinia and similar syntensional deposits to the Chuar Group are found in British Colombia, Utah, and California. A possible scenario is an intracratonic rift, where the basin would trap the sediment. Paired with changing sea level and rainfall patterns, enough carbon could be buried, causing a radical shift to the carbon curve. Potentially, this may have been able to remove sufficient CO_2 from the atmosphere to bring glaciers to lower latitudes and elevations (Dehler et al., 2012).

Summary

The Chuar Group records cyclical changes of low-energy, coastal marine environments resulting in the deposition of shales, carbonates, and sandstones in a tectonically active basin during a Neoproterozoic rifting event associated with the breakup of the supercontinent Rodinia. The position of the marine shoreline fluctuated slightly from east to west, but generally stretched north to south. The shoreline was wave- and tide-dominated and occasionally experienced strong storms. The presence of multiple types of single-celled, individual and bioherm-building organisms found in the Chuar Group aids in determining the depositional environments and making regional and global correlations. The Chuar strata record synchronous deposition with the development of the Butte Fault and Chuar syncline, massive excursions in the carbon curve believed to be correlative with low-latitude and low-elevation, global-scale glaciation.

The Sixtymile Formation (by Hannah Slover and Ken Bevis)

The Sixtymile Formation, the final unit of the Grand Canyon Supergroup (Figure 4), lies between the uppermost member of the Chuar Group and the Cambrian Tapeats Sandstone. It is exposed only in four isolated patches along the axis of the Chuar Syncline in Sixtymile Canyon (its type locality), Awatubi Canyon, and in Nankoweap Canyon. The lower portion of the unit forms the cap rock on Nankoweap Butte (Figure 9). The formation reaches a maximum thickness of 200 feet and is mostly composed of breccias and sandstones with minor siltstones and mudstones. Slump folds and carbonate landslide blocks surrounded by finer-grained siltstones and mudstones are also present, though the origin of the blocks is uncertain (probably from a lower Chuar Group rock unit). The Sixtymile Formation is informally divided into three unnamed members. The basal member, only present in Sixtymile Canyon, is no more than 90 feet thick. It is composed of coarse breccias of multiple clast types and red sandstones. Among the clasts observed within the breccia are pebble- to cobble-sized chert and dolomite rock fragments and large, landslide-derived blocks, all presumably originating from erosion of the Chuar strata undergoing deformation at the time. Some of the sandstones are thinly bedded and laminated. The middle member is 80 feet thick and combines thinly bedded and laminated, fine-grained quartzitic sandstone and siltstone. Characterizing the strata are common chert lenses, parting lineations, and massive, thin, white beds. Slump folds occurring in the middle member parallel the axis of the Chuar syncline, suggesting that poorly consolidated, water-saturated sediment occasionally slide toward the interior of the synclinal basin as warping of the strata continued. The final upper member is 40 feet thick and fills in channels cut into the middle member with locally derived fine- to coarse-grained sandstone, siltstone, conglomerate, and breccia. The channels run parallel to the Butte Fault and are as deep as 16 feet, indicating the flow of water within a synclinal valley. Sedimentary structures include contorted bedding, adhesion ripples, and small cut-and-fill structures filled with trough crossbedded sandstone (Ford and Dehler, 2003).

The Sixtymile Formation represents a drastic change in the environment during Supergroup deposition, moving to one that was largely terrestrial. The coarse clastic sediments and landslide blocks represent a huge transition from the marine carbonates, shales, and sandstones of the Chuar Group. The large boulders, breccias, and contorted bedding of the lower member likely occurred due to mass wasting processes initiated by growth of the Chuar Syncline and the Butte Fault. The red sandstones suggest a subaerial environment at or near sea level. The thinly bedded sandstones and siltstones of the middle member indicate a low-energy, fluvial environment such as a floodplain, or deposition in a localized lake setting along the synclinal axis. The cut-and-fill sequence at the base of the upper member represents localized exhumation and deposition by debris flow and fluvial processes. The upper member seems to be of fluvial origin where the sediments were possibly eroded from the tall footwall of the Butte Fault on the east side of the synclinal valley (Ford and Dehler, 2003).

This final rock formation of the Supergroup follows the Chuar Group and continues to record the movement of the Butte Fault through syndepositional processes. The incisions cut into the middle member indicate a relative lowering of base level caused by uplift and/or a drop in sea level. Faulting likely penetrated to the surface to produce a fault scarp when sufficient erosion resulted in a lack of fine-grained sediments on top of the Butte Fault (Dehler et al., 2012).

Summary

The Grand Canyon Supergroup is unmetamorphosed, providing a remarkable record of changing depositional settings within the Unkar and Chuar basins associated with alterations in paleoclimatic and tectonic forces. The Unkar Group, generally composed of carbonates, sandstones, and igneous rocks, is 6500 feet thick and dates from 1254 to 1104 million years ago. Deposition of the Unkar Group began on a penaplained surface cut to the very crystalline core of proto-western North America as the supercontinent Rodinia assembled and sediments were eroded from the resulting Grenville Orogeny and transported to the Unkar basin. The basin probably underwent tectonically-induced subsidence brought on by crustal extension while experiencing an easterly-directed, marine transgression, stutter-stepped with minor sea-level fluctuations. The lower Bass Formation and Hakatai Shale are intruded by diabase sills and dikes related to the later eruption of the Cardenas Basalt that caps the Unkar Group; the basaltic lavas of the unit likely recording a failed continental rifting event. Finally, the Unkar strata were subjected to moderate tilting and erosion with deposition of the poorly understood Nankoweap Formation to follow. The Nankoweap Formation is a 370-foot-thick sandstone unit accumulated in shallow marine waters that is believed to represent a brief pulse of deposition during an overall erosional hiatus of about 300 million years.

The Chuar Group unconformably overlies the Nankoweap Formation with 6800 feet of sediment formed by a cyclic deposition of shales, carbonates, and sandstones between 800 and 742 million years ago. Chuar Basin deposition began at the onset of a rifting event that culminated in the breakup of the supercontinent Rodinia. Single-celled organisms preserved in the strata show the flowering of the Earth's biodiversity. The Chuar sediments record a synchronous deposition of sediments and formation of the Butte Fault and Chuar Syncline, global carbon curve excursions, and the onset of low-latitude, world-wide glaciation. The final deposition of the Grand Canyon Supergroup is the 200-foot-thick Sixtymile Formation, which represents a drastic change to high-energy, terrestrial environments as uplift associated with growth of the Butte Fault forced a westward withdrawal of the sea. It continues the record of the synchronous deposition and movement of the Butte Fault-Chuar Syncline structural coupling. Eventually, extension and normal faulting offset crustal blocks by as much as two vertical miles to form a series of parallel basins and ranges (similar the Great Basin region today); basins preserved Supergroup rocks tilted backward into one-sided grabens. Subsequent erosion from about 740 million to 545 million years ago removed the Grand Canyon Supergroup and more of the underlying crystalline basement rocks from much of the Grand Canyon region, leaving only wedge-shaped remnants of Supergroup rocks preserved in large graben structures, rocks which are now observed in isolated pockets along the main Colorado River corridor and some of its major tributaries. The Supergroup is the oldest sequence of sedimentary rock preserved in the Grand Canyon. While it is scattered in patches and deformed, its unmetamorphosed sediments are in pristine condition and provide a window into the history of Late Proterozoic life, depositional environments, and structural development of the early North American continent.

The Paleozoic Sedimentary Rock Sequence

The Tonto Group (by Hannah Slover and Ken Bevis)

The Tonto Group is considered "one of the classic sequences of sedimentary rocks exposed in North America" (Middleton and Elliot, 2003). Spanning the Middle Cambrian period from about 525 to 505 million years ago, the Tonto Group represents the lowermost sedimentary rocks of the Paleozoic sequence exposed in the Grand Canyon, and is subdivided into three formations, the basal Tapeats Sandstone, middle Bright Angel Shale, and upper Muav Limestone (Figure 1). The rock units are best expressed in the Grand Canyon along a prominent, bench-like surface known as the Tonto Platform formed in the central part of the canyon were the weak Bright Angel Shale has eroded back from the inner gorge (Figure 10). The lower, cliff-forming Tapeats Sandstone is a tan, medium- to coarse-grained quartz-rich sandstone deposited in high-energy coastal plain braided streams, beaches, and intertidal to shallow subtidal wave- and tide-dominated environments (Middleton and Elliot, 2003). The middle Bright Angel Shale Formation is comprised of greenish-gray colored, silty to sandy textured, slope-forming mudrocks (Berthhault, 2004; Blakey and Middleton, 2012). These sediments were deposited on shallow near shelf environments and strongly influenced by tidal- and storm wave-generated currents (Middleton and Elliot, 2003). The upper Muav Limestone, tends to be a cliff-forming, yellowish-brown intermix of carbonates and fine siliciclastics. It was deposited in subtidal shelf areas dotted with carbonate islands farther offshore (Middleton and Elliot, 2003).

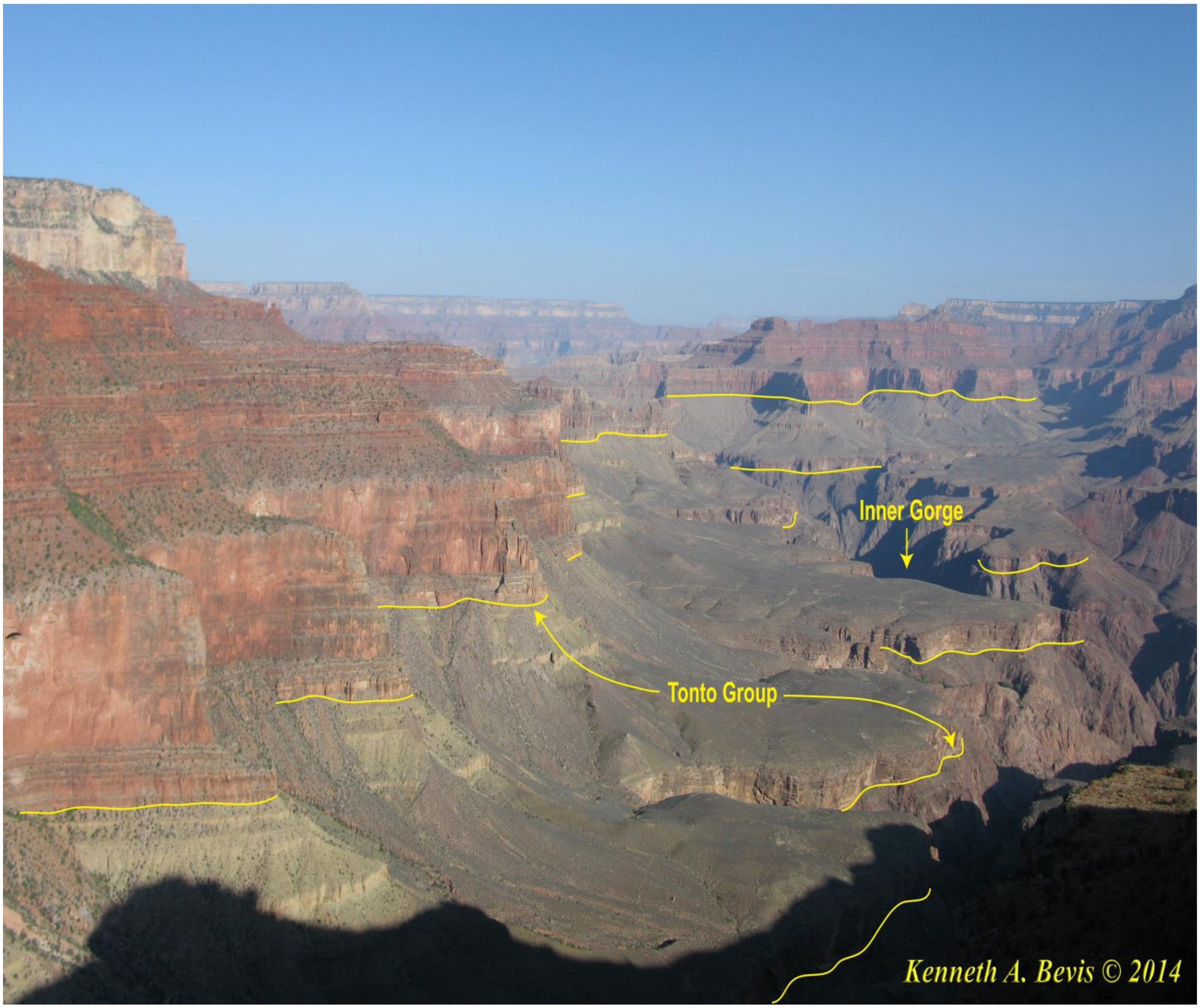

Figure 10. The Grand Canyon's Tonto Group, consisting of three formations, upper and lower cliff-formers and a middle slope-former, that crop out along a prominent bench or terrace above the Colorado River's inner gorge known as the Tonto Platform (that feature from which the group derives its name).

The sedimentary rocks of this group formed on the slowly subsiding, passive margin of western North America that formed after the breakup of the late Proterozoic supercontinent Rodinia (Blakey and Middleton, 2012). Its rock units mark an overall eastward-directed marine transgression of the proto-Pacific Ocean associated with a north-to-south oriented shoreline, but one characterized by multiple regressive phases that resulted in a complex intertonguing ofthe three formations. This process is nicely described by Blakey and Ranney (2008) as an oft-repeated cycle of "two steps in, one step out, two steps in, again and again" as the sea moved relentlessly eastward over millions of years. Figure 11 displays three paleogeographic maps developed by Ron Blakey that illustrate Early Middle Cambrian Tapeats Sandstone (Figure 11a), Middle Cambrian Bright Angel Shale (Figure 11b), and Late Middle Cambrian Muav Limestone (Figure 11c) depositional settings, respectively. Overall, Tonto Group deposition reflects the Simple Ideal Model (Figure 2 in THE GEOLOGY OF SEDIMENTARY ROCKS) of Fitcher and Poche (2001). Sediments accumulated during flooding of the passive continental margin of ancient North America during a major marine transgression towards the east, depositing a typical three-fold, onlapping, fining-upward sedimentary rock sequence comprised of shoreline quartz sandstones, near shelf mudstones, and far shelf carbonates (Figure 15 in THE GEOLOGY OF SEDIMENTARY ROCKS).

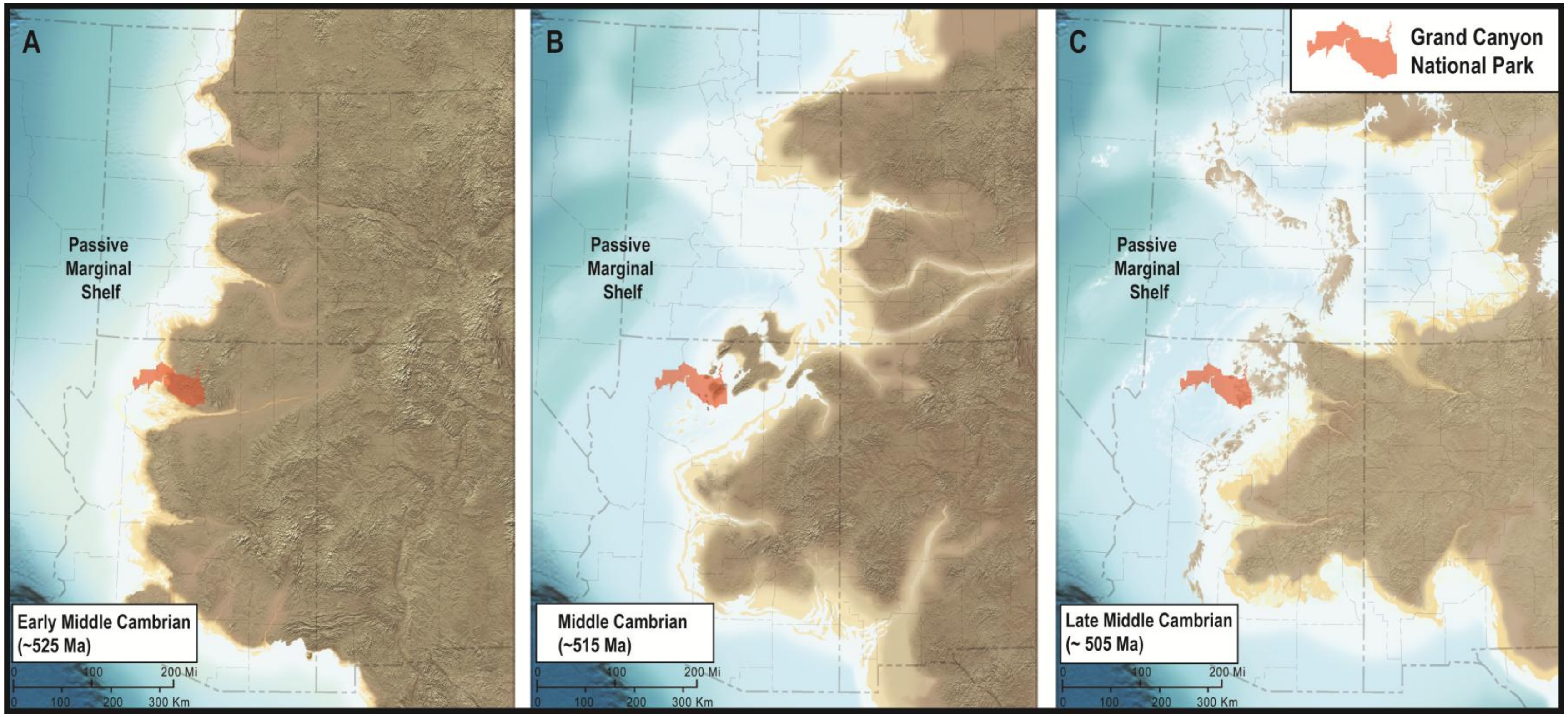
A
Passive
Marginal
Shelf
Early Middle Cambrian
(~525 Ma)
0 100 200 Mi
0 100 200 300 Km
B
Passive
Marginal
Shelf
Middle Cambrian
(~515 Ma)
0 100 200 Mi
0 100 200 300 Km
C
Grand Canyon
National Park
Passive
Marginal
Shelf
Late Middle Cambrian
(~ 505 Ma)
0 100 200 Mi
0 100 200 300 Km

Figure 31. Cambrian paleogeography of western North American during deposition of the Tonto Group's Tapeats Sandstone (a), Bright Angel Shale (b), and Muav Limestone (c); original maps are by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

The Cambrian Tonto Group directly overlies the Great Unconformity, with the 525 million year old Tapeats Sandstone deposited directly on tilted, 1.4 to 1.1 billion year old Grand Canyon Supergroup rocks in the eastern canyon, forming an angular unconformity; and 1.8 to 1.6 billion year old crystalline basement rocks of the Grand Canyon Metamorphic Suite in the west, forming a nonconformity (Middleton and Elliot, 2003; Blakey and Ranney, 2008). This lower contact is a major unconformity developed by erosion during extensive and prolonged subaerial exposure of western North America's passive margin that probably required substantial uplift brought on by mountain building. In some locations within the canyon, a thick paleosol occurs at the top of Precambrian surface suggesting significant chemical weathering of the basement rocks occurred prior to the beginning of Tapeats deposition (Middleton and Elliot, 2003). Wave erosion associated with initial Cambrian transgression may have removed the paleosol in areas where it is absent, and this weathered material provided a ready source of sediment as the sea moved eastward. The surface of the unconformity is very irregular. Lowlands, developed by weathering and erosion of weaker rocks, alternate with bedrock "hills" formed of more resistant material (Middleton and Elliot, 2003; Blakey and Ranney, 2008). The tips of the "hills" are often capped with the Bright Angel Shale rather than the basal Tapeats Sandstone which pinches out against the resistant knobs. Apparently, as sea level rose during the Cambrian transgression, these "hills" formed islands in the advancing sea, but were not inundated until deeper water conditions had ensued later during Bright Angel deposition.

The Tapeats Sandstone forms a conspicuous, pervasive, cliff-forming, medium- to coarse-grained sandstone weathered buff to brown in color deposited unconformably above Precambrian rocks; it is most easily recognized where it rests on vertically foliated rocks of the Grand Canyon Metamorphic Suite (Figure 12). The rock unit was named for exposures along Tapeats Creek in the west-central part of the Grand Canyon (Middleton and Elliot, 2003). The sandstone contains abundant quartz and lesser amounts of feldspar (Berthault, 2004), although as the formation progresses upward, the feldspar percentage, the bedding thickness, and the coarseness of the sandstone decrease (Middleton and Elliot, 2003). Overall, the formation ranges from 100 to 325 feet thick, reaching a maximum of 393 feet in Bass Canyon, although its thickness is clearly controlled by the relief of the unconformity it rests on. The base of the formation is most often composed of pebble conglomerates deposited in environments best characterized as "braided streams merging with nearshore marine areas" (Blakey and Middleton, 2012). The composition of the basal zone generally reflects the mineralogy of the underlying Precambrian rocks (Middleton and Elliot, 2003), with highest quartz and feldspar content near areas with abundant Zoroaster Granite.

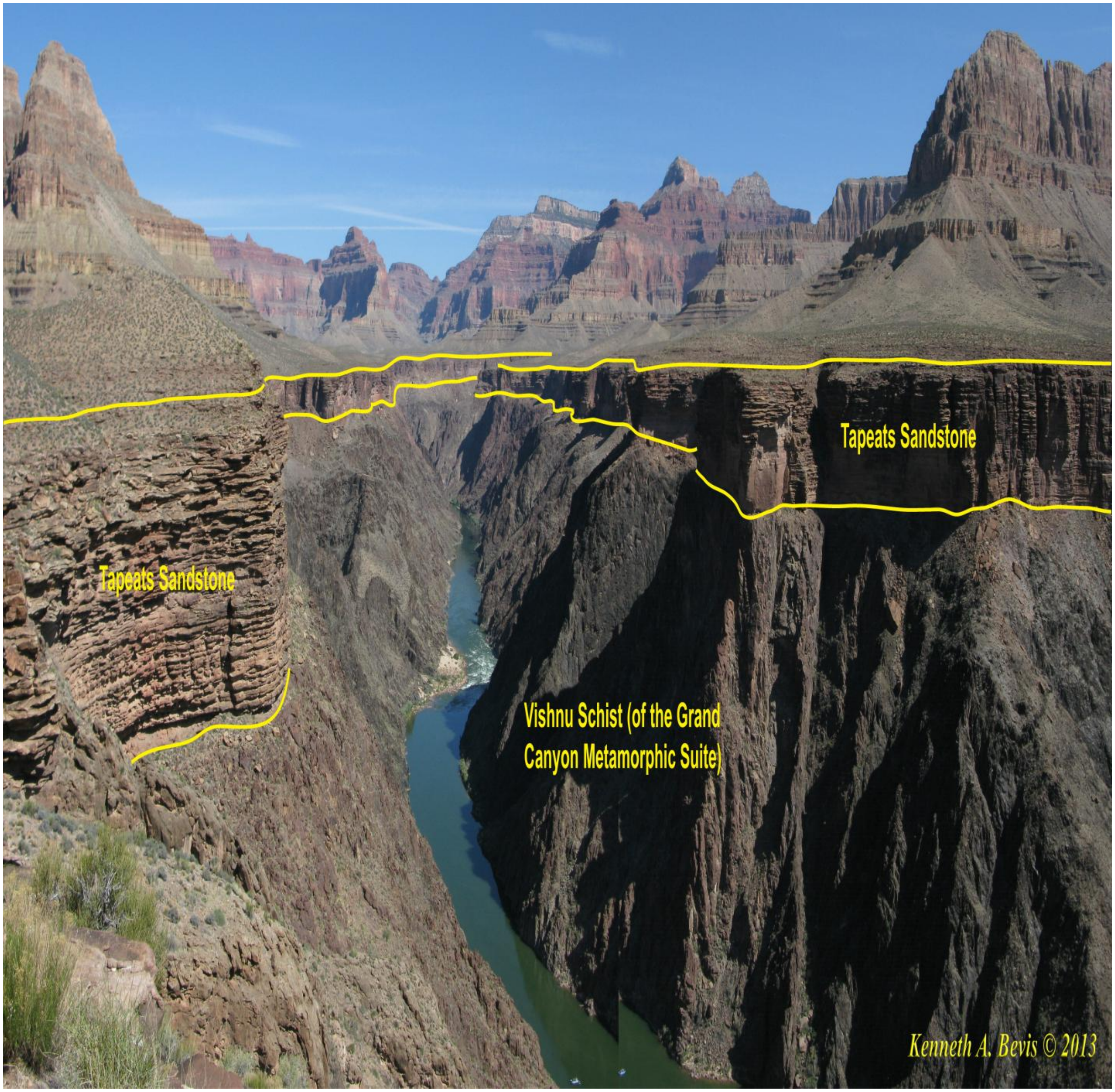

Figure 12. The Tapeats Sandstone overlies Precambrian basement, in this case crystalline rocks of the Grand Canyon Metamorphic Suite; here viewed along the inner gorge of the Colorado River from the Tonto Trail between Travertine and Boucher Canyons.

The majority of the Tapeats Sandstone forms a distinct cliff characterized by relatively thick bedding (individual beds are up to three feet thick) that exhibits planar and trough cross-stratification and poor horizontal stratification. The upper section tends to be weathered back into a stepped-ledge slope of thin interbeds of fine- to medium-grained lensoidal sandstone and mudstone interbeds, accompanied by trough and ripple cross laminations (Middleton and Elliot, 2003). The Tapeats' transition into the Bright Angel Shale forms an intercalated, gradational contact and marks a significant facies transition to lower energy conditions (Middleton and Elliot, 2003 and Berthault, 2004). Body fossils are very rare in the Tapeats Sandstone, but more common in the finer-grained sediments near the top of the unit (Blakey and Middleton, 2012). Fossils have been used to date the Tapeats as late Early Cambrian in the west and early Middle Cambrian in the east, indicating its time-transgressive nature. According to Blakey and Middleton (2012), the Tapeats was deposited along a shoreline characterized by numerous embayments where fluvial deposits derived from braided streams draining westward off the Transcontinental Arch merged with shallow, high-energy marine environment. It was likely formed in beach, intertidal barrier bar or shoaling environments, or shallow, subtidal zones influenced by tidal currents, although wave processes probably dominated in areas adjacent to resistant, island-forming "hills".

The Bright Angel Shale is fine-grained and dominated by greenish muds composed mostly of clay (Middleton and Elliot, 2003). The mudstone is accompanied by minor sandy dolomites and silty limestones, and somewhat more common green siltstones and sandstones containing glauconite and dark brown ironstone (Berthault, 2004). The mineral glauconite, common to clastic shelf environments, is responsible for the pervasive green coloration of the rock unit. The formation's abundance of muds provides a relatively high susceptibility to weathering and erosion and is responsible for its expansive, slope-forming topographic expression. Its name derives from the broad slope of interbedded shale, siltstone, and sandstone exposed just above the Tonto Platform near Bright Angel Creek. The view expressed in Figure 10 shows this slope east of Bright Angel Canyon quit well, although a closer view is offered by Figure 13, where the unit is exposed along Cottonwood Canyon.

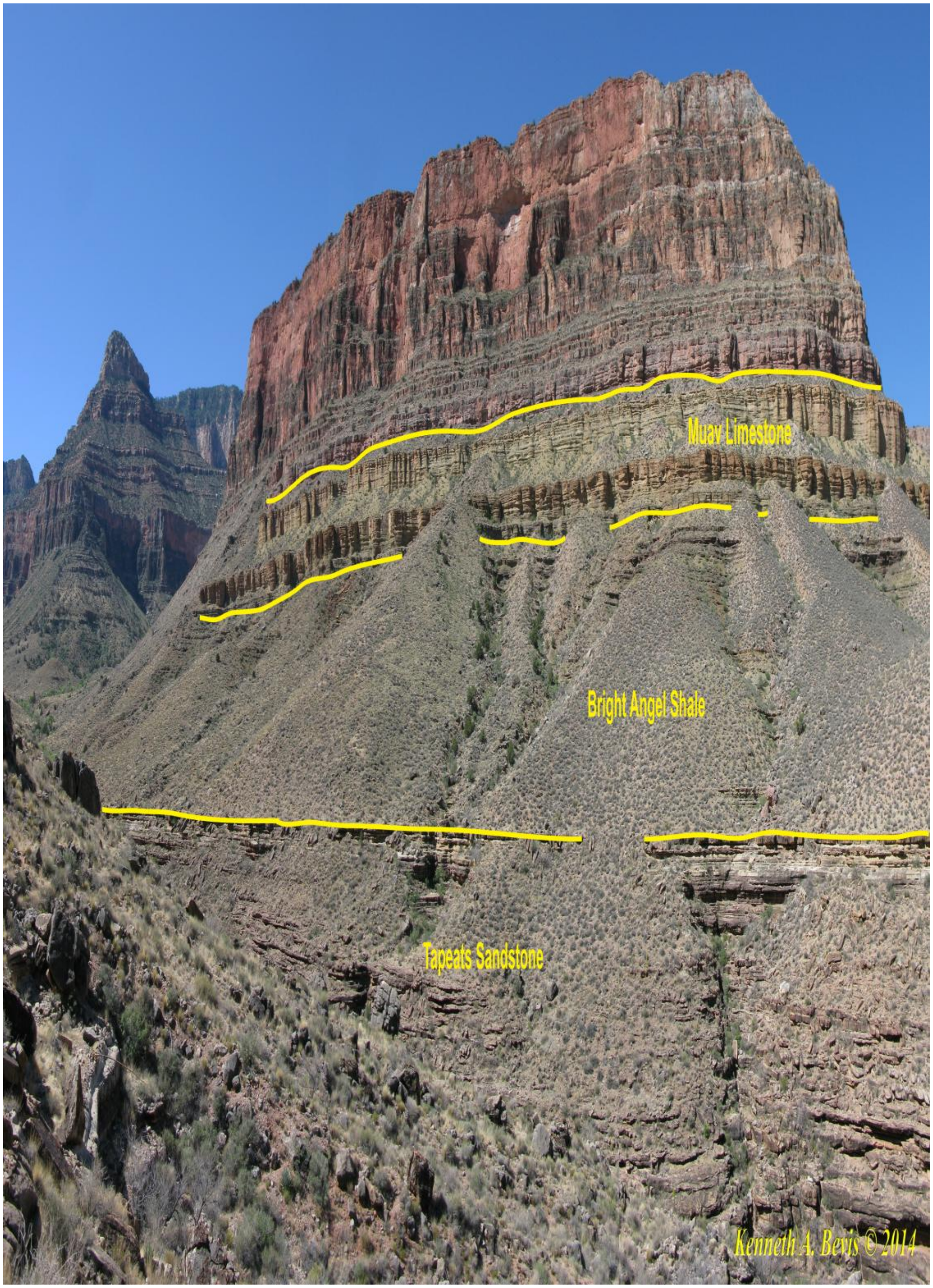

Figure 13. The weak, slope-forming middle unit, and the stronger, cliff-forming upper unit of the Tonto Group, the Bright Angel Shale and Muav Limestone respectively, are nicely exposed in the walls of Cottonwood Canyon as viewed from the Tonto Trail as in begins to round Horseshoe Mesa.

The average thickness of the formation is 320 to 400 feet; however, it thickens to over 450 feet in the western Grand Canyon, thinning to 325 feet at Bright Angel Creek in the central canyon, but reaching only 270 feet thick in the eastern part of the canyon. This complex variation in thickness is due to the time-transgressive nature of the unit; the near shelf conditions that caused its deposition persisted much longer to the west than to the east. Only one member is identified in the Bright Angel Shale formation. Located in the uppermost portion of the unit in its western canyon exposures, it is comprised of shale, siltstone, and limestone; although as the member progresses to the east, the limestone decreases and the member becomes entirely shale (consistent with an eastward-directed transgression). According to Middleton and Elliot (2003), the sedimentary structures identified in the Bright Angel Shale formation include "horizontal laminations, small- to large-scale planar tabular and trough cross-stratification, and wavy and lenticular bedding." They also identified coarsening- and fining-upward sequences in the central Grand Canyon. Trace fossils and body fossils are abundant in the Bright Angel Shale, and Blakey and Middleton (2012) report that the fossils record a "major proliferation of invertebrate fauna." The fossils place the western edge of the Bright Angel Shale during late Early Cambrian and the eastern side as Middle Cambrian (Middleton and Elliot, 2003), again attesting to its time-transgressive nature. The deposition of the Bright Angel Shale occurred in deeper water, under the lower energy currents of the subtidal shelf environment influenced by both calm weather and storms, and was likely deposited in lagoons or estuaries influenced by tidal action (Blakey and Middleton, 2012).

The Muav Limestone, also a cliff-forming unit, is the uppermost formation of the Tonto Group (Figure 10 and 13). Blakey and Middleton (2012) report that the formation "records the first widespread accumulation of carbonate sediments ….. in the Paleozoic." It is composed of yellowish-brown to yellowish-green, silty and sandy, dominantly impure limestone, though the purity improves to the west (Berthault, 2004). Its dominant color provides its namesake. It is mainly calcareous mudstone nearer its base, and its contact with the Bright Angel Shale is also intercalated and gradational, reflecting intertonguing of adjacent environments related to small-scale, short-duration, repeated transgressions and regressions probably during the highstand of the Cambrian sea. Its limestones, which become more abundant upward in the formation, are chiefly packestone, accompanied by intraformational, flat-pebble (intraclastic) conglomerates, and with minor interbeds of shale, siltstone, fine-grained sandstone, and silty limestone (Middleton and Elliot, 2003), suggesting deposition associated with clear, shallow-marine shoals and platforms (Blakey and Middleton, 2012) occasionally disturbed by storms.

In general, the siliciclastic content of the rock unit increases eastward as the carbonates decrease, in accordance with an eastward-directed transgression. Bedding alternates from thin to thick with small, irregular clay patches, and overall, bedding thickens to the west. The thin beds tend to be composed of micaceous shale and siltstone with a minor amount of sandstone and limestone. Beds are typically structureless or horizontally laminated. Also marking the bedding are small-scale trough, planar tabular, and low angle cross-stratification, as well as fenestral fabrics and desiccation cracks.

Mirroring the bedding is the overall thickness of the formation (Middleton and Elliot, 2003). It is 827 feet thick in the west at the Grand Wash Cliffs, 439 feet thick at the Tuweap Valley in the central canyon, and only 136 feet thick in eastern Grand Canyon. Within the Muav Limestone, seven members can be identified and separated on the basis of fauna, or prominent conglomerate, shale, or limestone layers. The lower four members are only found in the western canyon, but the upper three members spread throughout its entirety. Trace fossils are present in the Muav Limestone, but are not as abundant as they are in the Bright Angel Shale; however, according to Blakey and Middleton (2012), the Muav contains “the first major occurrences of invertebrate fossils”. The western section of the Muav has been assigned to the Middle Cambrian Period and the eastern section to the late Middle Cambrian Period (Middleton and Elliot, 2003), again paralleling the time-transgressive age trend of the two lower units in the Tonto Group.

In the western section of the Grand Canyon, dolostones overlie the Muav Limestone to a thickness of over 400 feet. It is thought that these rocks are Upper Cambrian, though there is no paleontological evidence to support this assumption. Geologists refer to it as the Cambrian Undifferentiated or the Grand Wash Dolomite. Three lithofacies are present within this dolomite sequence. Middleton and Elliot (2003) describe them as “white-to-buff massive dolomite,” “white-to-yellow, very fine-grained, thick-bedded dolomite,” and “gray, fine-grained, thick-bedded dolomite.” Oolitic grainstones and stromatolitic beds lie between these layers. The sedimentary structures include wavy and asymmetric ripple lamination and small-scale cross-stratification, and the dominant trace fossil is horizontal burrows and tracks (Middleton and Elliot, 2003).

Brachiopods and trilobites, though poorly preserved, are the most common invertebrates found in the Tonto Group (Middleton and Elliot, 2003). Brachiopods are most often found in the coarse-grained sandstone of the Bright Angel Shale and the mixed siliciclastics and carbonates of the Muav Limestone; while several dozen species of trilobites have been identified within the Tonto Group, congregating in the coarse-grained sandstone of the Bright Angel Shale and the mudstones of the Bright Angle Shale and the Muav Limestone. Fragmented sponges, primitive mollusks and echinoderms, and algae occur in the Bright Angel Shale and Muav Limestone, but they are not abundant. Trace fossils in the form of tracks, trails, and burrows are evident in all three formations of the Tonto Group, with the majority in the Bright Angel Shale and the least amount in the Muav Limestone (Middleton and Elliot, 2003). They become increasingly abundant in the upper half of the Tapeats Sandstone as it transitions into the Bright Angel Shale, with single and paired vertical tubes and horizontal traces quit common. The “unbranched, straight vertical burrows” created by Skolithos in the fine-to-coarse-grained and cross-bedded sandstone provide evidence of currents capable of active bedload transportation. U-shaped burrows, perpendicular to the bedding in fine- to coarse-grained sandstone of the Bright Angel Shale and Tapeats Sandstone have lead geologists to infer shallow-water deposition. Finally, within the Bright Angel Shale and the Tapeats Sandstone, on interbedded sandstone and mudstones are trilobite crawling tracks, referred to as Cruziana, likely created during a time of fair weather, while the resting tracks, referred to as Rusophycus, are suggestive of storm conditions (Middleton and Elliot, 2003)

The Tapeats Sandstone was deposited on sandy beaches, intertidal sand and mud flats, and shallow, subtidal sand wave complexes. Sedimentation below the beach zone, beyond the coast, occurred in water depths less than 100 feet. The consistent dip directions of the cross-stratification provide evidence that the islands dotting the coastline had minor influence on sedimentation, simply being a source of sediment themselves. Tidal flat mud deposition ranged from high to low and was intermixed with evidence of sandy tidal channels. At the base of the Tapeats, less mature, coarse-grained sandstone and conglomerates alternating with planar-tabular and trough cross-stratified sandstone fill shallow channels and imply braided stream deposits which grade into marine facies. Large channels occurring near the top of the Tapeats Sandstone exhibit internal structure suggestive of formation by subtidal channel complexes dominated by the ebb tidal phase.

The Bright Angel Shale was deposited on open shelf environments, in water depths between the Tapeats Sandstone and the Muav Limestone, but generally below fair-weather wave base. A combination of alternating upward-coarsening, upward-fining, cross-bedded sequences, and interbedded sandstone and mudstone indicate subtidal deposition manipulated by tidal currents and storm wave processes. Upward-coarsening sequences, up to 25 feet thick and several tens of kilometers wide, record deposition in a solely subtidal environment though the upper portion of many were in shallow, intertidal waters. The lower, finer sections contain laminated, bioturbated mudstones probably deposited during fair weather. The upper, coarser sections exhibit thick, planar tabular cross-stratification, often with abrupt changes in the dip of foresets that likely formed by migration of sand waves, dunes, and ripples associated with stronger tidal and storm-wave generated currents. Fining-upward sequences typically overlie an erosive base. They begin with pebble conglomerate or sandstone and grade into interbedded fine-grained sandstone and mudstone. The coarse basal layer is representative of high-energy deposition during storm-induced currents capable of transporting coarse material. Symmetric ripples and a high amount of laminated mudstone occur at the top of these sequences. They are marked by a diverse array of trace fossils representative of post-storm resettling of muds and sands.

The Muav Limestone, like the Bright Angel Shale, was also deposited in a subtidal environment. East of Bass Canyon, intraformational, often imbricated, flat-pebble conglomerates consisting of disc-like clasts of mostly micrite mark an extremely important facies within this rock unit. The intraformational conglomerates occur as both discontinuous lenses and large flat sheets extending for miles, and likely indicate subtidal deposits created during temporary regressions. In addition to a subtidal environment, some limestone and dolostone beds are thought to have been deposited in intertidal and supratidal flats. The laminated dolostones of this formation deposited in the western Grand Canyon area may have been limey sediments bound by tidal flat algal mats. The undifferentiated dolomites found in the western part of the canyon containing thick beds of oolitic grainstones and stromatolites interbedded with fine-grained carbonates are probably representative of shallow subtidal and perhaps intertidal environments, again related to regression.

There were many influences on the sedimentation of the Tonto Group and deposition occurred in a variety of interrelated environments associated with an overall eastward directed marine transgression of a north-south oriented coastline. Complex intertonguing of the rock units was generated in response to multiple regressive phases during dominant transgression onto North America's passive margin. This passive continental marginal suite records the classic onlapping, fining-upward, sedimentary sequence of the Simple Ideal Model (Figure 2 in THE GEOLOGY OF SEDIMENTARY ROCKS). The Cambrian Tonto Group contains the typical, stacked, beach and nearshore generated sandstone (Tapeats Sandstone), near shelf generated mudstone (Bright

Angel Shale), and far shelf generated carbonate facies (Muav Limestone) produced during a marine transgression.

The Tonto Group is geologically unique in several ways. Its rock units are well exposed and readily studied. Not only do they demonstrate a classic marine transgressive sequence, obeying Walther's Law of Facies and the Simple Ideal Model, but they also describe the time-progressive nature of many formations, that is, the sediments they contain are not the same age everywhere, but change in age as the depositional environment they are associated with migrates. In this case, each formation becomes younger toward the east as the marine transgression gradually inundated the passive continental margin. And of course, these rocks wouldn't have been preserved if not for the self-generating nature of passive continental marginal tectonic settings, creating accommodation space for the sediments to accumulate through subsidence (a mere rise in sea level could not account for the thickness of material preserved). Finally, these Middle Cambrian rock units also portray the amazing proliferation of life on planet earth at this time, located in the middle of a very short (geologically speaking), 150 million year time stream in which life went from nothing more advanced than single-celled organisms in the late Proterozoic, to evolve the most complex phyla found on earth today by the Late Ordovician (placing this is context, it took 3,000 million years of earth history to create single-celled organisms).

The Temple Butte Formation (by Hannah Slover and Ken Bevis)

The Temple Butte Formation is a westward thickening carbonate layer in the Grand Canyon deposited during the Late Middle to early Late Devonian Period roughly 385 million years ago (Beus, 2003). The rock unit is named for exposures on Temple Butte in the eastern Grand Canyon. It forms a relatively inconspicuous layer at the eastern end of the canyon where it crops out only as discontinuous lenses of paleo-stream channel fill (Figure 14), and at the western end where it forms a thicker, continuous layer, its relative resistance causes it to merge with the overlying Redwall Limestone cliffs overlying it. Extensive dolomization of these carbonates makes it difficult to interpret a depositional environment, but a few fossils are present to aid interpretations. In the east, it is likely that the formation was deposited in narrow tidal channels, appearing as discontinuous lenses, while the central and western outcrops were deposited in widespread, shallow, subtidal, marine conditions, resulting in a thicker and more continuous layer. Blakely and Ranney (2008) suggest that the channels were marine estuaries extending finger-like eastward onto the continental margin and coalescing westward into a contiguous shallow marine shelf environment.

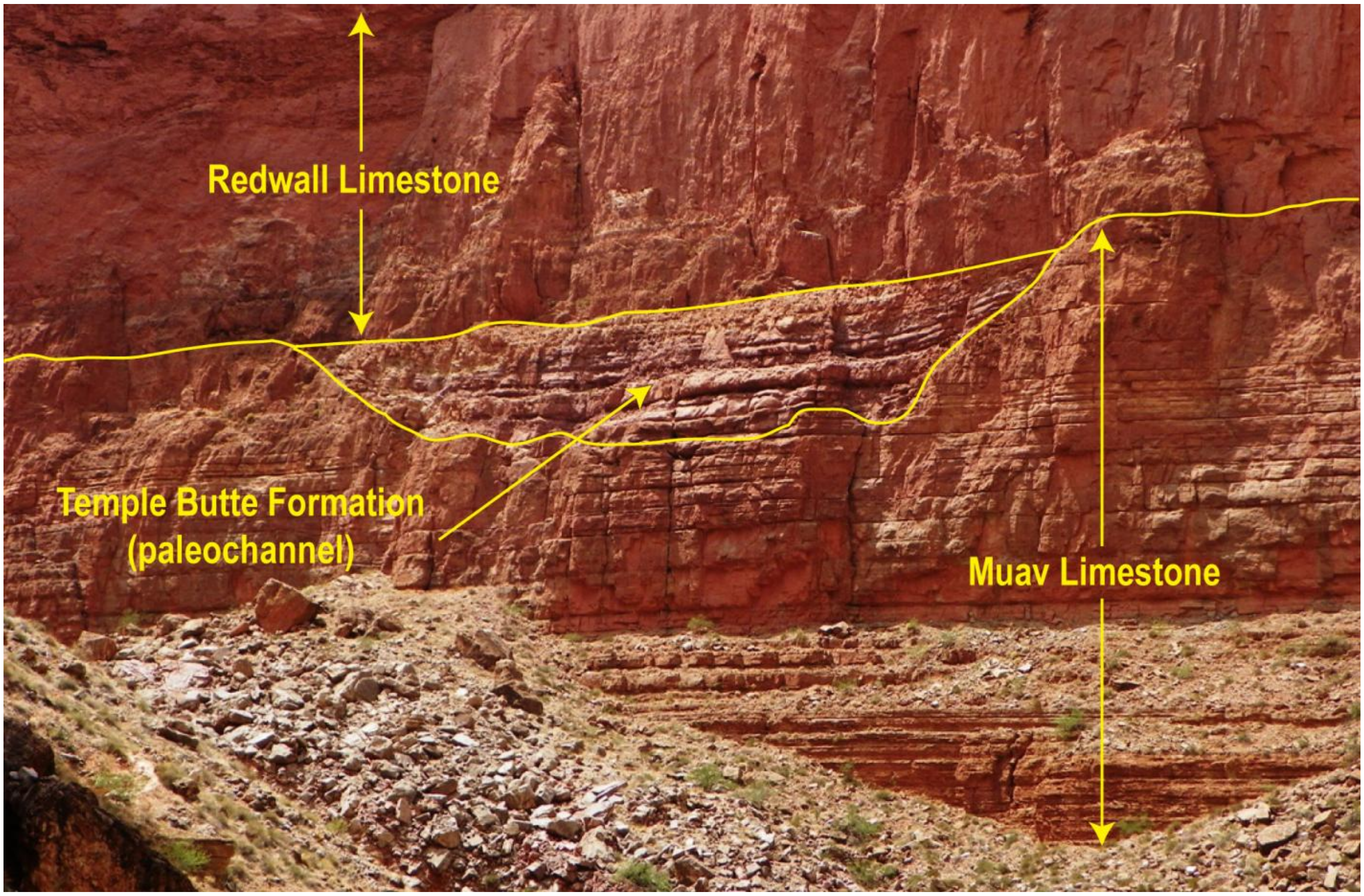

Figure 14. The Temple Butte Formation filling the paleochannel of an ancient stream cut into the Muav Limestone in Saddle Canyon (a side tributary of the Colorado River at mile 47 in Marble Canyon).; photo by Garry Hayes from his Geotripper blog, posted August 24th, 2013).

The Temple Butte's upper and lower contacts are unconformable with the Redwall Limestone and Muav Limestone, although the lower unconformity represents a much more significant time gap. Absent between the Muav Limestone and Temple Butte Formation is the Late Cambrian, all of the Ordovician and Silurian, and most of Early and Middle Devonian sediments, some 120 million years, but the gap between the Temple Butte and Redwall is only upper Devonian, a mere 45 million at most. Both gaps in the rock record are disconformities; that is, the rocks above and below the erosion surface are both sedimentary and are essentially undeformed. In the west (and also near Lava Canyon in the east) light-gray strata of undifferentiated Cambrian dolomite up to 500 feet thick conformably overlie the Muav Limestone (Beus, 2003). Overall, except for their appearance at Lava Canyon where they have been reported at a thickness of 163 feet, the strata are truncated eastward by the Temple Butte. These unnamed layers do appear to conform with deposition of the Muav Limestone, and Korolev and Rowland (1993) have correlated the undifferentiated dolomite layer in the Grand Canyon to the Banded Mountain Member of the Bonanza King Formation of Nevada, which is thought to be of late Middle Cambrian age. Where the Temple Butte is in direct contact with the Muav Limestone, the unconformable surface is identified by the considerable relief created by channels and depressions up to 100 feet deep. However, there are areas where the channels are absent. The result is a planar contact of gray dolomite from the Temple Butte and the Muav Limestone, making the unconformity difficult to locate even though it represents more than 100 million years of missing time (Beus, 2003).

The southwestern North American plate during the Late Devonian was tectonically quiet, essentially retaining all the features of earlier formed passive continental margin setting (Figure 15); although the Antler volcanic arc was approaching from the west, plowing into slope-rise deposits off the continent's edge (Blakely, 2014). As the arc encroached eastward, itbegan to collide with the passive margin of western North America**,** thrusting these deep water deposits onto the continental shelf and causing regional subsidence that may have aided deposition of the Temple Butte Formation. Overall, the westward thickening sediments of the Devonian Temple Butte are representative of a marine transgression to the east onto a submerged continental shelf (Beus, 2003; Blakey and Middleton, 2012).

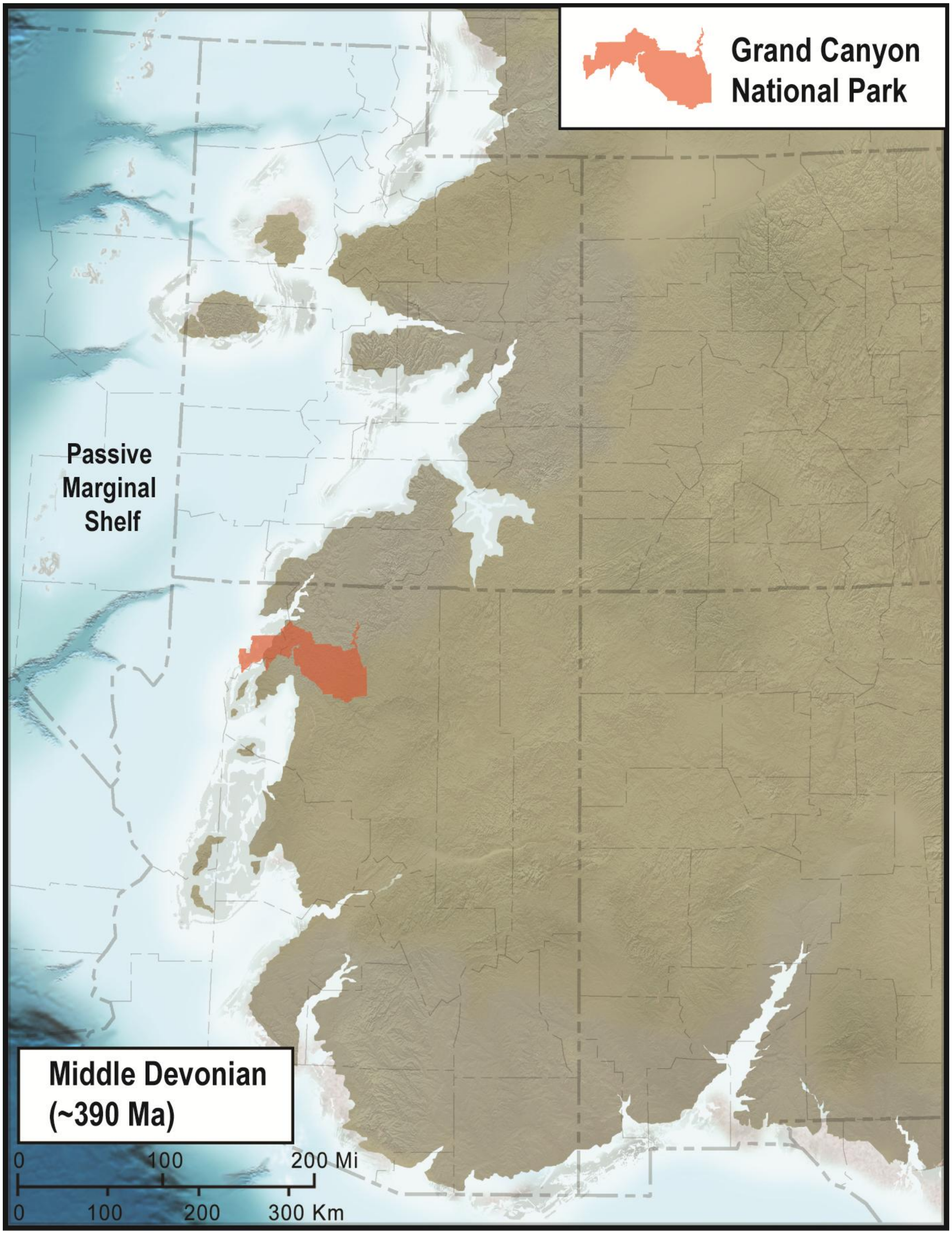

Figure 15. Middle Devonian paleogeography of western North America during deposition of the Temple Butte Formation; original map is by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

The Temple Butte is composed of gray and tan limestone and dolostone with minor traces of siliciclastics, but is often weathered to a purplish color. It occurs as thin, patchy layers, deposited in “deep, channelized surfaces” in the east but is “more continuous” in the west (Blakey and Middleton, 2012). The discontinuous lenses in the east are usually less than 100 feet in thickness, but may reach up to 400 feet in width and are “eroded into the upper surface of the Muav Limestone or the Cambrian undifferentiated dolomite” (Beus, 2003). To the west of Hermit Creek, the formation (composed of continuous dolomite) thickens to more than 450 feet (Beus, 2003).

The basal channel-fill sections of the Temple Butte are usually composed of dolomite or sandy dolomite of a distinct pale, reddish purple color, creating irregular, or even gnarly, beds (Beus, 2003). Basal conglomerates composed of subrounded dolomite pebbles occur, but are very rare. Variations occur between these beds; horizontal in some localities, or conforming to the walls of the channel in others, but when the strata do conform to the curved channel margins, the layers above often truncate them with angular discordance. The upper section of the channel fills are often composed of about twenty percent insoluble residue and an interesting columnar alternation pattern of pale gray and purple dolomite probably caused by weathering. Bedding above these channels is more continuous and can be described as thick and blocky ledges. It is a “fine to medium crystalline dolomite” with a “dark-to-light olive gray” color and often “weathers to a sugary texture” (Beus, 2003). Another, less common, but extensively deposited version of the carbonate in the Temple Butte formation occurs as very fine crystalline dolomite with a grayish-orange-pink color. It is thinly bedded, has evidence of concoidal fracture, often seems porcelaneous, and it weathers to an extremely light gray or yellow. Within this layer of aphanitic dolomite, laminae or thin lenses of rounded, frosted quartz sand occur; although locally, lenses may reach up to 7 feet thick. Most of the dolomite beds form steep, receding ledges with the exception of the central Grand Canyon where they conform to the unbroken, 1550 foot carbonate wall comprised of the Redwall and Muav Limestone. While fossils are abundant in other Devonian formations, few are identifiable in the Temple Butte. Brachiopods, gastropods, corals, fish plates, and trace fossils are all present, but conodont microfossils are the most helpful in determining the age to be of latest Middle Devonian to early Late Devonian (Beus, 2003).

Temple Butte strata truncate older rocks from west to east across the Grand Canyon region. The disconformity at the base of this formation marks a significant stratigraphic break in the Grand Canyon’s Paleozoic sedimentary sequence, representing the latest Cambrian, the entirety of Ordovician and Silurian, and Early through Middle Devonian time. Sediments of the upper Muav Limestone were deposited during eastward-directed marine transgression of the Middle Cambrian period while Temple Butte sediments mark a much later transgression of the Middle Devonian sea, making the unconformity greater in the east because subaerial exposure would have dominated this region longer

In general, the deposition of the Temple Butte Formation occurred on a passive continental margin. Extensive dolomization and the paucity of recognizable fossils make it difficult to pinpoint exact depositional environments. The Temple Butte of the central and western Grand Canyon seems to have accumulated in shallow, subtidal, open marine conditions which remained more or less static over time despite overall transgression. The aphanitic limestone is more specifically considered to have been deposited in supratidal conditions. Discontinuous lensoidal patches of Temple Butte in the eastern Grand Canyon were formed in tidal channels associated with an intertidal environment. These deposits thicken and coalesce to the west where marine conditions became more contiguous. A comprehensive paleoenvironmental interpretation of Temple Butte deposition is a system of estuary channels in the eastern Grand Canyon area that merged with a more pervasive, gradually subsiding, shallow marine shelf to the west.

The Redwall and Surprise Canyon Formations (by Hannah Slover and Ken Bevis)

The Mississippian Period produced two carbonate dominated formations, the Redwall Limestone and Surprise Canyon Formation. Its basal unit, the Redwall Limestone, forms massive, steep cliffs and is stained red by the iron oxides weathered from the overlying Hermit Formation and Supai Group rocks, and is probably the most easily identified rock unit in the Grand Canyon. Wedged between the Redwall Limestone and the overlying rocks of the Supai Group is found the inconspicuous, discontinuous patches of limestones, siltstones, sandstones, and conglomerates high in fossil content that are now recognized as belonging to the Surprise Canyon Formation (Blakey and Middleton, 2012). Both formations accumulated in shallow-marine passive-marginal environments (Beus, 2003), although of a very different nature and extent (Figure 16). The Redwall Limestone accumulated during the Early to early Late Mississippian associated with two major transgressive-regressive cycles on open-marine continental shelf carbonate platforms (Figure 16a). It is broken down into four member subunits (Beus, 2003). The basal layer is referred to as the Whitmore Wash member and was deposited during the first transgression. This layer is followed by the Thunder Springs member, formed during the subsequent regression. The latter two layers, the Mooney Falls and Horseshoe Mesa members, accumulated during the second transgressive-regressive cycle. The Surprise Canyon Formation was deposited during the Late Mississippian (and Early Pennsylvanian) Periods, in an array of discontinuous karst-generated sinkholes and caves and in the paleovalleys of ancient dendritic drainage systems eroded into the limestone of the Redwall after marine regression and subaerial exposure of much of the carbonate-dominated shelf (Figure 16b). The Surprise Canyon's limited exposure makes it undeniably the least easily identified Paleozoic rock unit in the Grand Canyon. The variety and distribution of this unit's deposits suggest a transitional environment from fluvial to estuarine conditions associated with marine transgression onto a karstified coastal plain dissected by stream valleys.

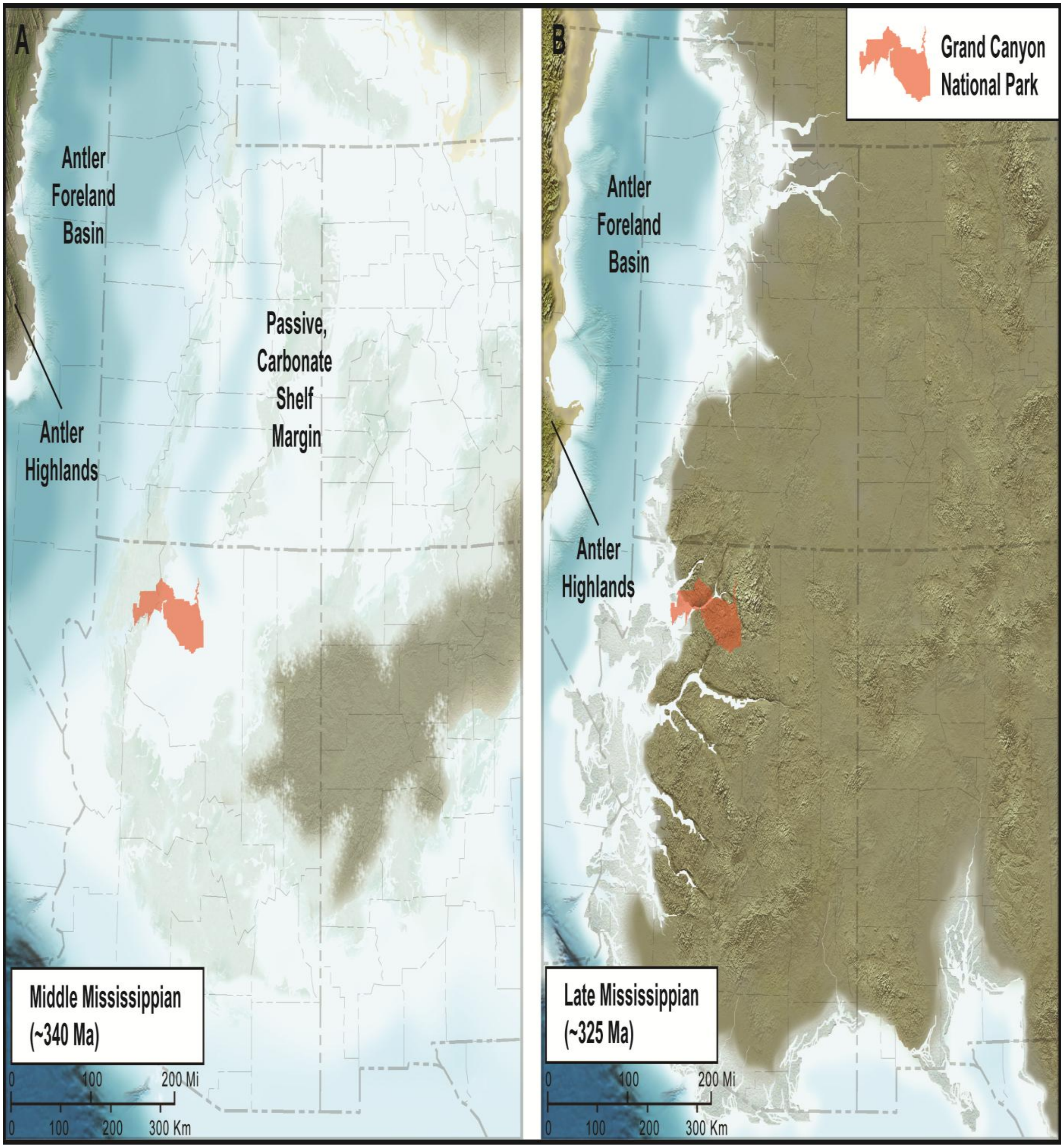

Figure 16. Middle Mississippian (A) and Late Mississippian (B) paleogeography of western North America during deposition of the Redwall Limestone and Surprise Canyon Formation, respectively; original maps are by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

The passive continental margin of the Mississippian North American west coast was generally unaffected by tectonism. However, the convergence of an oceanic plate with the North American plate in latest Devonian-Mississippian time resulted in the Antler Orogeny to the west in central Nevada (Beus, 2003). The colliding Antler arc shoved strata deposited in deep water over the continental shelf to form the Antler highlands (Blakely, 2014). The weight of the thickened crust caused eastern Nevada and western Utah to subside in a forearc basin and thick Upper Devonian and Lower Mississippian strata formed there. Farther east, thinner deposits accumulated on a broad carbonate shelf over much of the Western Interior (Figure 16a). The Grand Canyon sat on the periphery of orogenic activities, and was affected by them only indirectly (Beus, 2003). A broad, but gentle forearc bulge probably enhanced subaerial exposure in the Late Mississippian and helped induce channel incision on the karstified landscape (Figure 16b). Localized uplift also occurred as faults shifted the crust due to the contractions associated with the Antler Orogeny. For example, following the deposition of the Redwall Limestone and prior to the Surprise Canyon Formation, an anticline caused by east-west compression and reactivation of a buried fault was responsible for the truncation of the Horseshoe Mesa member of the Redwall Limestone in the Tanner Trail area of the eastern Grand Canyon.

In general, the Redwall forms soaring, mottled red cliffs spanning five- to eight-hundred feet in height, probably the most recognizable rock unit in the Grand Canyon (Figure 17). It is tallest in the west at roughly 800 feet at Iceberg Ridge, five miles west of the mouth of the Grand Canyon, and just over 400 feet in the easternmost Grand Canyon near the cliffs forming the Palisades of the Desert. The Redwall Limestone is deposited unconformably on Devonian sediments or, where the Temple Butte Formation is absent, on Cambrian sediments of the Muav Formation. Missing between these formations is latest Devonian and earliest Mississippian time. In the western Grand Canyon, the basal beds of the Redwall are of Early Mississippian and its abrupt contact with Devonian sediments is well marked by an erosion surface with up to 10 feet of relief. The more substantial part of the unconformity occurs in the east where the sediments are of the late Early Mississippian Period. The time difference between the basal sediments from west and the east is due to the eastward directed transgression. Locally, the unconformity is typically bordered by a conglomerate of angular dolomite or limestone clasts derived from the Temple Butte Formation, but becomes very difficult to locate when the strata on either side of the unconformity are dolomitic. These factors suggest a gentle uplift paired with mild erosion of the strata before deposition of the Mississippian Redwall began (Beus, 2003).

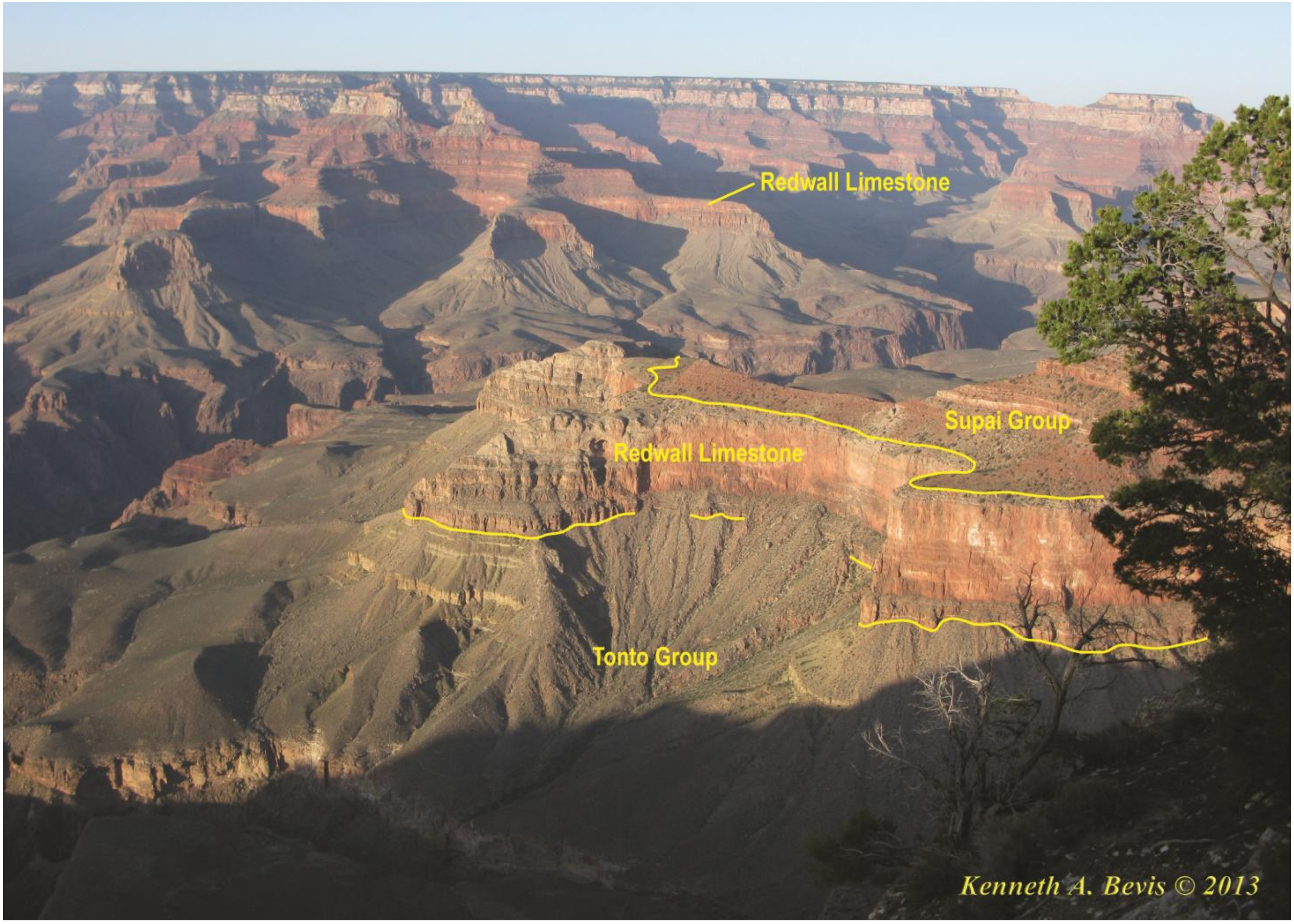

Figure 17. The Redwall Limestone literally forms a massive “red wall” spanning hundreds of feet in height broken only by the occasional stream drainage, one of the most easily recognized Paleozoic rock units in the Grand Canyon; in this view from Yavapai Point looking northeast, in the foreground, the Redwall forms the foundation of Cedar Ridge, and across the canyon in the background, it can readily be observed just above the gray slopes of the Tonto Platform.

Grand Canyon geologist Edwin McKee divided the Redwall into four members (Beus, 2003). He named the basal late Early Mississippian layer “Whitmore Wash” for its type section at that location. Composed almost entirely of pure carbonates, often “pelleted and locally skeletal or oolitic wackestones and packstones,” it is also accompanied by iron oxides and less than 2 percent of gypsum. It is deposited in thick beds, ranging 2 to 4 feet thick, and is dominated by fine-grained limestone in the west and dolomite in central and eastern Grand Canyon. The Whitmore Wash unit spans an area from about five miles beyond the western edge of the Grand Canyon at a thickness of approximately 200 feet to Iceberg Ridge in the east, thinning to around 100 feet. Due to extensive dolomization, fossils are rare in the Whitmore Wash member. Its upper contact is conformable and easily distinguished from the alternating dark chert and light carbonate beds of the Thunder Springs member. The initial deposition of the Whitmore Wash began as a major transgression resulted in oolitic shoals created by high energy currents in nearshore, shallow subtidal conditions, and were followed by continuous offshore deposition of skeletal grainstones and packstones in quieter water (Beus, 2003).

Following the transgression of the Whitmore Wash was the regression that deposited the Thunder Springs member of the Redwall Limestone in more shallow marine conditions. This second unit is composed of alternating thin and dark reddish brown or gray chert beds with light gray carbonate beds. In the west, the carbonate bed is a more pure limestone, while in the east it is more dolomitic. Beus (2003) reports that the chert layers are "silicified blue-green algal mats." Invertebrate marine fossils are very abundant in the chert beds and include corals, bryozoans, brachiopods, and crinoids. While fossils still appear in the carbonate beds, they are poorly preserved because of the dolomization. The member thickens from 100 feet in the east to about 150 feet in the west. Overlying the Thunder Springs is the Mooney Falls member, along an unconformable contact except in the extreme west. It is a low-angle disconformity suggesting only minor structural activity and erosion (Beus, 2003).

Like the previous two members of the Redwall Limestone, the third and thickest member, Mooney Falls, also thickens from east to west, ranging from around 200 to 400 feet. This member forms the sheer wall plastered in oxidized muds that inspired its name. It is characterized by pure limestone with thin beds or lenses of chert towards the top, but is locally dolomitized. The limestone is thickly bedded, micritic, and accompanied by carbonate grains including oolites, pellets, and skeletal fragments. Invertebrate marine fossils are also abundant in the Mooney Falls member and include solitary and colonial corals, spiriferid brachiopods, and crinoids. The top third of the member in the central and eastern Grand Canyon is also characterized by large-scale, tabular-planar cross-bedding. The second marine transgression was responsible for the deposition of the Mooney Falls in open marine waters (Beus, 2003). The upper contact of the Mooney Falls member with the overlying Horseshoe Mesa member is conformable and difficult to locate. Geologists identify the contact where the limestone transitions from thickly-bedded medium and coarse limestone to the aphanitic, thin-bedded, ledge-forming limestone of the Horseshoe Mesa (Beus, 2003).

The final and youngest member of the Redwall Limestone is the Horseshoe Mesa member of the early Late Mississippian Period. Also thinnest in the east, it ranges from 45 to 125 feet thick, and wedges out 30 to 40 miles south of the Grand Canyon due to erosion. In most of central Arizona, it is absent from the top of the Mooney Falls member. Composed of thin layers of light gray limestone, it has a texture ranging from mudstone to wackestone, and forms weak, receding ledges. Similar to the underlying Mooney Falls are chert lenses in the lower portion of the Horseshoe Mesa member. Invertebrate fossils are rare, but well preserved, with at least sixteen species of foraminifers identified. As the final regression occurred during the deposition of the Horseshoe Mesa member, the water became increasingly shallow resulting in restricted circulation. The upper contact of the Horseshoe Mesa member with the Watahomigi Formation of the Supai Group forms a major unconformity; locally the Horseshoe Mesa lies in contact with patchy lenses of the Surprise Canyon Formation along a lesser unconformable surface.

The Late Mississippian Surprise Canyon Formation occurs in isolated patches between the underlying Redwall Limestone and the overlying Watahomigi Formation of the Supai Group. It filled the caves and sinkhole depressions of a karstified limestone landscape and the erosional valleys of dendritic river systems with clastic and carbonate sediments (Figure 18a). Sediment-filled sinkholes and caves often had high porosity relative to the surrounding solid limestone and often served as conduits for mineralized fluids. Indurated breccia pipes containing uranium and copper ores were sometimes the result (Figure 18b). These patches occur in V- or gentle U-shaped cross-sections throughout the Grand Canyon and Marble Canyon to the east. Patches in the western Grand Canyon are typically 150 to 200 feet deep and up to half a mile wide, but in the central and eastern Grand Canyon and Marble Canyon, they are shallower and wider, usually reaching a thickness of only 50 feet. The largest pale river valley occurs in Quartermaster Canyon, a south rim tributary in western Grand Canyon, with a thickness of 400 feet. The dendritic drainage system flowed east to west, merging to form more continuous patches and fewer, larger channels to the west, where it met with the paleocoastline to form an estuarine environment. The depth of paleovalley incision suggests subaerial exposure associated with gentle uplift or a sea level drop by as much as 100 feet (Beus 2003). Blakey and Middleton (2012) indicate that much of the North American continent was gently uplifted in the Late Mississippian. Streams flowing into the marine trough formed by the still extant Antler foreland basin carved westward-deepening valleys, and these were subsequently flooded to form coastal estuaries and then filled by marine deposits to generate the Surprise Canyon Formation (Figure 18b).

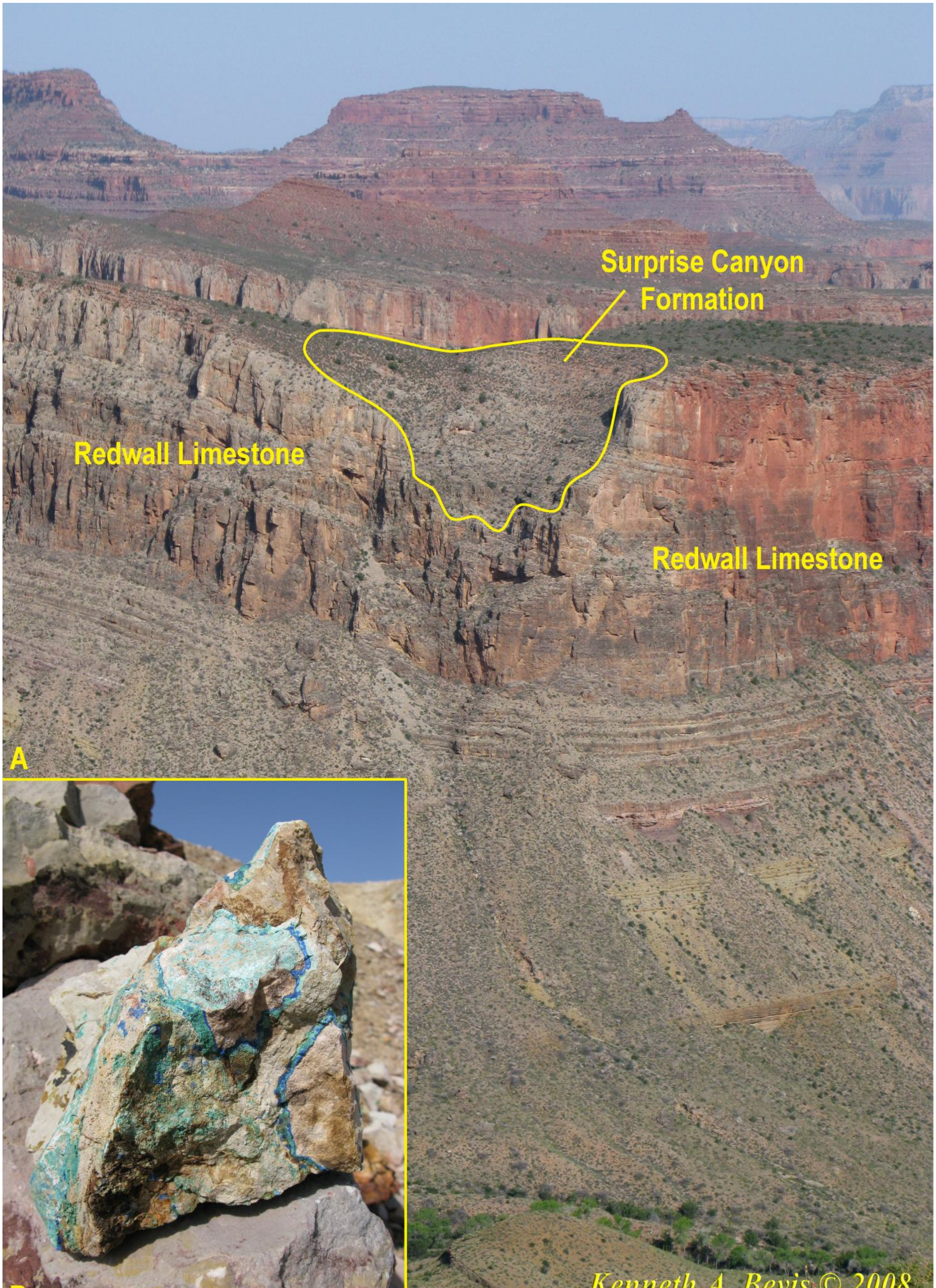
Surprise Canyon
Formation
Redwall Limestone
Redwall Limestone
A
B
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Figure 18. A sinkhole formed in the top of the Redwall Limestone on the south side of Cottonwood Canyon, filled with deposits of the Surprise Canyon Formation (A); because these sinkhole deposits filled with brecciated rock, they had high porosity relative to the surrounding solid limestone and often served as conduits of uranium and copper ore mineralization such as at Horseshoe Mesa (B).

According to Beus (2003), the Surprise Canyon Formation "has the most varied sedimentary lithology of any Paleozoic formation in Grand Canyon." It is laterally divided into eastern and western units; where in the east, the stratigraphy merges into a single red-brown, slope-forming conglomeratic

sandstone or siltstone, while in the western and central Grand Canyon, the Surprise Canyon is divided into three more distinct units and each unit has multiple lithofacies. The lower western unit is a fining-upward conglomeratic sandstone of fluvial origin, forming cliffs and slopes. Beus (2003) describes this basal unit as a "ferruginous pebble-to-pebble and local boulder conglomerate." Clasts are also present and dominated by chert, with minor amounts of limestone derived from the underlying Redwall, which is enclosed in a quartz matrix (with some hematite). The presence of imbricated cobbles provides evidence for a unidirectional, westward flowing, fluvial current. Conglomerates grade upward into quartz sand or siltstone or carbonaceous shale. The sandstone facies is "yellow to dark reddish brown or purple," flat-lying, characterized by trough cross-strata or ripple lamination, and contain Lepidodentron log impressions (Beus, 2003). The shale facies is dark and plant fossils have been found in the unit. Beus (2003) reports that this lower unit preserves "a record of continental and fluvial conditions changing to intertidal conditions as the sea transgressed eastward into the estuary and began to trap and rework clastic sediments within the tidal range."

The middle western unit of the Surprise Canyon Formation is a cliff-forming, skeletal limestone with a coarse grainstone texture and whole or fragmented shells of marine origin. The limestone occurs in thin beds with thicknesses up to 4 inches and alternates with thinner quartz sandstone beds. Due to erosion, this middle unit often truncates the lower unit, and after weathering has a yellowish brown, rusty or purple gray color. While the lower unit has evidence for a unidirectional current, the small-scale trough-cross strata of the middle unit suggest a bimodal current with the dominant direction upstream, or to the east (Grover, 1989). Paired with the abundance of marine invertebrate fossils, the bimodal current is suggestive of an estuary environment dominated by flood tides. This unit is absent east of the Grand Canyon's Fossil Bay area (Beus, 2003). The upper western unit is a marine siltstone and silty or sandy limestone, and forms weak slopes to receding ledges. It is dark red-brown to purple in color, and is planar bedded to ripple laminated. Nearly spherical algal stromatolites occur in some portions of the unit, but the relative absence of marine fauna suggest a restricted marine to upper tidal-flat environment as the karstified topography of the Redwall Limestone is finally filled in (Beus, 2003).

The fossil record of the Surprise Canyon is extremely rich and diverse. It contains more than 60 species of marine invertebrates with brachiopods being the most common form. In the middle unit, microfossil invertebrates are “moderately abundant” with 7 species of foraminifiers and 10 forms of conodonts (Beus, 2003). Bone fragments and sharks teeth are common in this unit. Plant fossils are most common in the western part of the lower unit, with multiple types of algal structures and “12 species of plant megafossils” (Beus, 2003).

The contact between the Surprise Canyon and the Watahomigi Formation is often obscured, but where it is exposed, a low-angle unconformity has been observed and the basal unit of the Watahomigi is a thin, widespread, locally discontinuous limestone pebble conglomerate with minor traces of chert clasts (Beus, 2003). If the conglomerate is absent, the contact can be identified by a purplish red calcareous siltstone and mudstone.

Geologists believe that the Redwall Limestone was deposited under shallow, open-marine conditions on a submerged continental shelf across northern Arizona during the Middle Mississippian. The four members of the Redwall were formed during two major transgressive-regressive cycles. Following the final phase of Redwall deposition, seas withdrew from the Colorado Plateau region for a time and the exposed landscape became karstified and incised by west-flowing rivers. A subsequent Late Mississippian to Early Pennsylvanian marine transgression produced the Surprise Canyon Formation. The rapid and significant lateral and vertical facies changes within the Surprise Canyon has lead geologists to infer deposition corresponding to a major east-west estuary system in the Grand Canyon area that emptied onto the paleocoastline near the Nevada-Arizona border. The eastern limit of marine fossils suggests a fluvially-dominated system in the eastern canyon progressing to a more marine-dominated environment in the central to western canyon at this time. The region remained largely unaffected by tectonic activity during the Mississippian, although the Antler Orogeny to the west likely caused sporadic upwarping of the crust, aiding the occasional exposure event and production of unconformable surfaces within the Redwall-Surprise Canyon sedimentary rock sequence.

The Supai Group(by Hannah Slover and Ken Bevis)

Stretching across the Grand Canyon from rim to rim is a thick band of red and tan cliffs and slopes that comprise the Supai Group (Figure 19). To even the untrained eye, the alternating stepped-ledge layers of sandstones, mudstones, and limestones of the Supai Group are easily distinguished from the soft, dark red slopes of the Hermit Shale above and the sheer cliff of the Redwall Limestone below. Helble (2011) suggests that these rocks reflect a diverse array of paleoenvironments that included continental, shoreline, and shallow-marine origins. The Supai Group is divided into four individual formations beginning with the oldest unit: the Watahomigi Formation characterized by red mudstone, siltstone, gray limestone, and dolomite; the Manakacha Formation dominated by quartz sandstone; the Wescogame Formation also of sandstone but intermixed with mudstone and limestone; and finally the Esplanade Sandstone with minor traces of mudstone and gypsum. Ages, determined primarily by brachiopods and fusulinids, indicate a correlation with the Pennsylvanian and Early Permian Periods. Common characteristics among the formations include mud cracks, ripple marks, and worm burrows (Hill, 2009). Vertebrate footprints occur in the Wescogame formation and Esplanade Sandstone

(Helble, 2011), probably representing reptile tracks. Deposited prior to the Supai Group during the Mississippian Period is the Redwall Limestone, with the Surprise Canyon Formation sandwiched in discontinuous patches between the two. Following the deposition of the Supai Group is the Hermit Formation, also of the Permian period, and often associated with the Supai Group because of similarities in sediment type and inferred origin even though an unconformity separates the two rock units (Blakey, 2003).

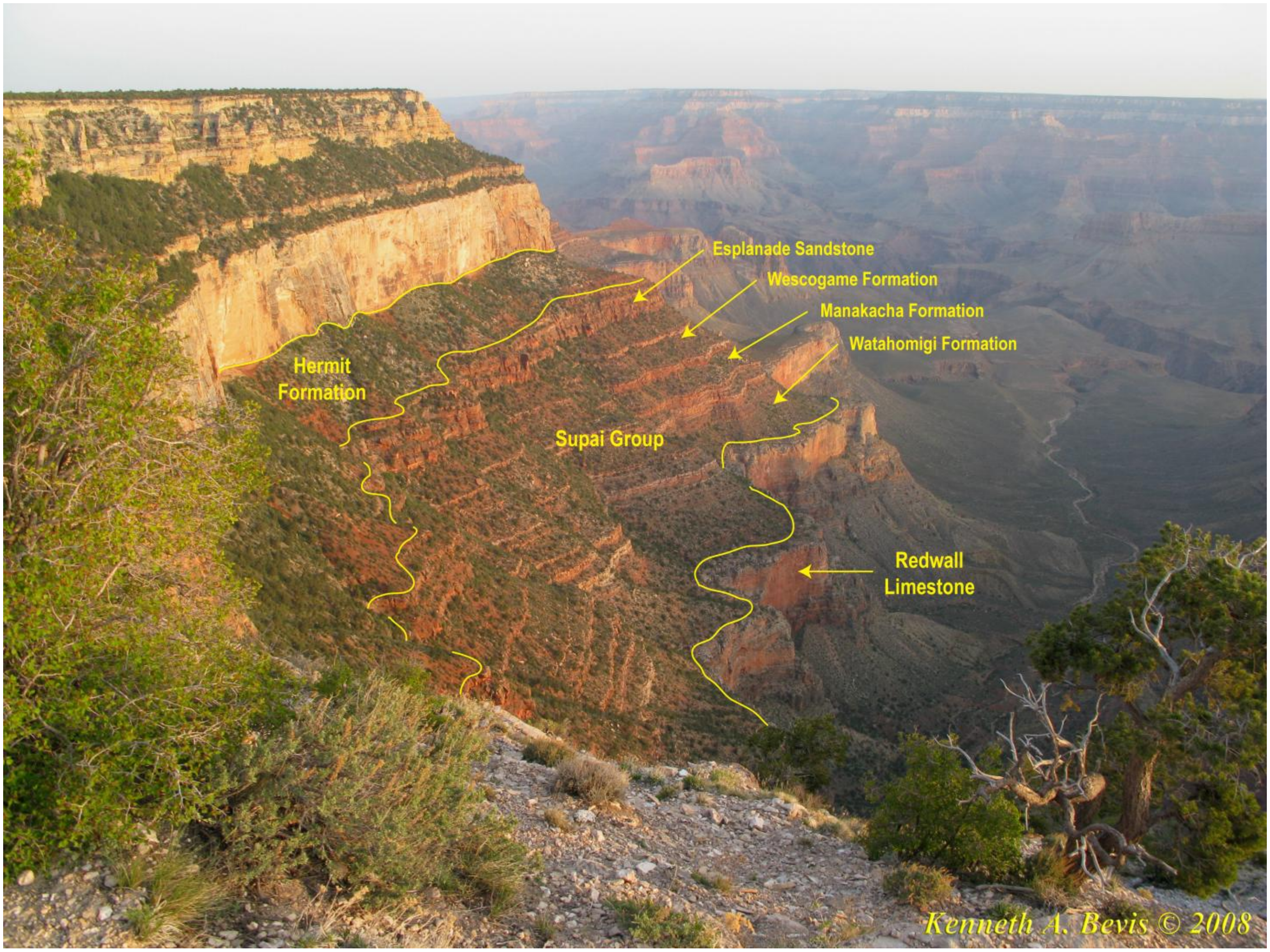


Figure 19. The Supai Group consists of alternating cliff- and slope-forming “redbeds”, layers of resistant sandstones and limestones interbedded with less resistant mudstones that tend to produce a distinctive stepped-ledge appearance in outcrop; the view here is from Shoshone Point.

The oldest rock unit of the Supai Group is the Watahomigi Formation which accumulated during the Early Pennsylvanian Period. It gradually thickens to about 300 feet in western Grand Canyon and the Grand Wash Cliffs compared to a thickness of 100 feet in the eastern Grand Canyon and in Sycamore Canyon to the South. Overall, it is composed of a mixture of red mudstone, siltstone, gray limestone, dolomite, and some sandstone. Informally, the Watahomigi Formation is divided into three horizontal sections: the lowest, a redbed slope, the middle, a carbonate ledge, and the upper section, another redbed slope. In addition to a horizontal division, the Watahomigi Formation can be divided laterally into an eastern and western facies, eastward being dominated by clastic redbeds and westward by the carbonates. The carbonate sediments in the east are mostly fine-grained limestone, while the limestone is more granular to the west with "abraded fossil fragments, accretal grains, and pellets" (Blakey, 2003). The majority of the mudstone is composed of "non-descript, slope-forming, poorly exposed units with thin, intercalated, bioturbated (disturbed by organisms), bright-orange, limey sandstone" (Blakey, 2003). When the redbeds are exposed, various bed-plane tracks, trails, and burrows along with a basal chert-pebble conglomerate are commonly observed (Blakey, 2003). This combination was deposited as a result of regional subsidence. Sediment from the low-energy, shallow-marine and coastal-plains blanketed the area to create the mudstones, carbonates, and minor sandstones (Blakey, 2012). Where the Watahomigi Formation lies directly on the Redwall Limestone, the contact is typically "sharp and unconformable," although the abruptness of the contact is less discernible between the Watahomigi and Surprise Canyon Formation. The upper contact with the Manakacha Formation appears to be conformable thought not necessarily of level stratigraphy since the contact joins the slope below with steep slope or cliff above (Blakey, 2003).

Following the deposition of the Watahomigi is the Manakacha Formation. Also formed in the lower Pennsylvanian Period, it marks a change in the depositional facies and their thicknesses that dominated earlier time periods during the Paleozoic era. Previously, carbonates and minor mudstones were the norm, but the Manakacha Formation is largely composed of quartz sand with minor traces of mudstones and limestones, and is thickest in the central Grand Canyon, rather than in the west as are other Supai Group formations. As Blakey (2003) explains "the deposition of the Manakacha reflects a significant [environmental] change across the western interior of the United States".   The formation is typically 300 feet thick in Grand Canyon while only 150 feet thick throughout the Verde and Chino valleys to the south. It consists of three types of sandstone containing "very fine- to medium-grained quartz grains to ooids, abraded fossils, and peloids" (Blakey, 2003). Units are typically held together with calcite, and the "more limey" sections are accompanied by Jasper (red chert). According to Blakey (2003), the first sandstone is cross-stratified and comprised of trough, planar, and compound bed sets, in thicknesses ranging from 1 to 30 feet. Within these strata are found climbing translatent strata. Supporting an interpretation of eolian deposition for the unit are thin laminae that show reverse grading while each lamina records the migration of a single wind ripple. The geometry of the strata leads geologists to suspect deposition at the forward base and between dunes, on eolian sheets of sand (supported by the horizontal to very low angles of sediments), and due to migration of the eolian dunes (supported by the climbing strata). The second sandstone is described as having "horizontally laminated to very low-angle, cross-stratified, fine-grained, calcareous, and silty units up to 20 feet thick" (Blakey, 2003). Of a deeper red color, this stratum is also characterized by rare ripple lamination and the abundance of local tracks and trails on the bedding planes. This evidence suggests a subaqueous genesis origin, but clues are limited. Finally, the third sandstone found in the Manakacha Formation is very fine-grained sand, ranging from structureless to greatly bioturbated layers. It is thought that both deposits have undergone bioturbation, but the structureless layers were so extensively mixed that all traces of burrows or tracks have been removed. This sandstone's color ranges from grayish orange to a bright reddish orange. The strata of the Manakacha Formation often display evidence of alteration by diagenesis (growth of calcite and dolomite crystals), and minor amounts of mudstone, fine-grained limestone, and dolomite are common throughout the formation. As the Manakacha formation moves toward the northwest, it increases in cross-bedded carbonates, implying deposition by fluid currents (Blakey, 2003). The mixture of river, shoreline, and shallow-marine sediment suggests complex intertonguing of low-energy, intertidal, estuarine, and eolian environments as the sea continued to transgress and regress (Blakey and Middleton, 2012). A regional unconformity occurs at its upper contact with the overlying Wescogame Formation, leaving absent the Middle Pennsylvanian period. The boundary is identified by scouring, channeling, and a coarse lag of conglomerates, although it is vague in the southern Colorado Plateau due to poorer outcrops, and in general is difficult to locate, and was likely associated with prolonged subaerial exposure (Blakey, 2003).

The third formation of the Supai Group is the Wescogame Formation. Deposited during the upper Pennsylvanian period, its lithologic components are exceedingly similar to that of the Manakacha. This sheet-like body of sand expands across the Grand Canyon and the western Mogollon Rim region while maintaining thicknesses from 100 to 200 feet. In the central Grand Canyon and western Chino Valley, sandstone is the dominant facies, while mudstone increases in the east and limestone to the west of Chino Valley. The Wescogame Formation is informally divided into an upper slope unit and a lower cliff unit. In addition, three north-east-trending facies belts have become associated with this formation. The easternmost belt is the redbed facies, located near the Hermit Basin and spanning to the western Verde Valley. The middle belt is the sandstone facies, extending west from the eastern belt through the Shivwits Plateau region. Finally, cross-stratified limestone and limey sandstone composes the limestone facies of the westernmost belt (Blakey, 2003). While the depositional environment is still the same low-energy environments of the Manakacha, a decrease in marine fossils leads geologists to suspect an increase in continental settings (Blakey, 2012). The lower contact is with the Manakacha Formation and, as discussed above, is often difficult to locate because of the variations of the conglomerates and debris that tend to obstruct the view. Most likely marking the disconformity between the Pennsylvanian-Permian boundary is the Wescogame's upper contact with the Esplanade Sandstone, an unconformable contact that is generally simple to locate by the presence of lag conglomerates and its high relief of up to 50 feet (Blakey, 2003).

The Esplanade Sandstone forms an easily distinguished cliff band in the Grand Canyon directly below the slope-forming Hermit Formation and is the final formation of the Supai Group (Figure 20). This rock unit is characterized by minor mudstones and bedded gypsum (Blakey, 2012) and has a higher percentage of sandstone than any other in the Supai group (Blakey, 2003). It accumulated during the Lower Permian, and it is correlative with the Pakoon Limestone, a carbonate mud stratum, intertonguing from the west (Blakey and Middleton, 2012). Blakey (2003) describes the formation as a wedge, thickening from around 200 to 250 feet in eastern Grand Canyon and western Mogollon Rim region to 800 feet in the northwest at the western end of the Grand Canyon along the Grand Wash Cliffs. The 800 foot thickness includes the Pakoon Limestone which thickens to about 300 feet along the Grand Wash Cliffs. The Esplanade Sandstone is compositionally very similar to the Manakacha Formation, but it does differ in its dominance of cross-stratification. It also exhibits evidence of eolian deposition due to the abundance of climbing translatent strata. Its color, originally a pale grayish orange to reddish orange, appears dark reddish brown due to staining from iron oxides by the overlying Hermit

Formation. The formation contains multiple stacked and interfingering sandstone units, ranging from 5 to 50 feet thick, and divided by “thin, red mudstone; by fine-grained carbonate units; or by prominent, irregular-bedding planes” (Blakey, 2003). It is informally divided into a basal slope unit and the main cliff unit. The Pakoon Limestone formed in the west grades laterally from this basal slope unit, a transition characterized by sandstone grading into dolomite and limestone (Blakey, 2003). The majority of eolian sand in the west is of marine origin while the east is of fluvial origin. The abundance of cross-stratification in the sandstone is supportive of eolian dune and sand sheet environments. The minor mudstones were likely deposited in tidal-estuarine and fluvial floodplains, the gypsum on coastal sabkhas, and the limestone in a shallow-marine environment (Blakey, 2012). Still debated amongst geologists, the Esplanade Sandstone’s upper contact with the Hermit Formation is generally believed to suggest a time of environmental transition and/or a regional unconformity. Different evidence includes 30 to 50 feet of local relief, channeling (originating from the Hermit Formation) into the Esplanade Sandstone, and cross-stratified sandstone 100 feet thick, beveled into the contact. In the west, the lower slope unit transitions into the Pakoon Limestone as indicated, while the upper cliff unit increases in carbonate content. In the Toroweap area, the top of the Esplanade Sandstone experiences bedded gypsum that is “intercalated with cross-stratified sandstone throughout a zone as much as 200 feet thick (Blakey, 2003).

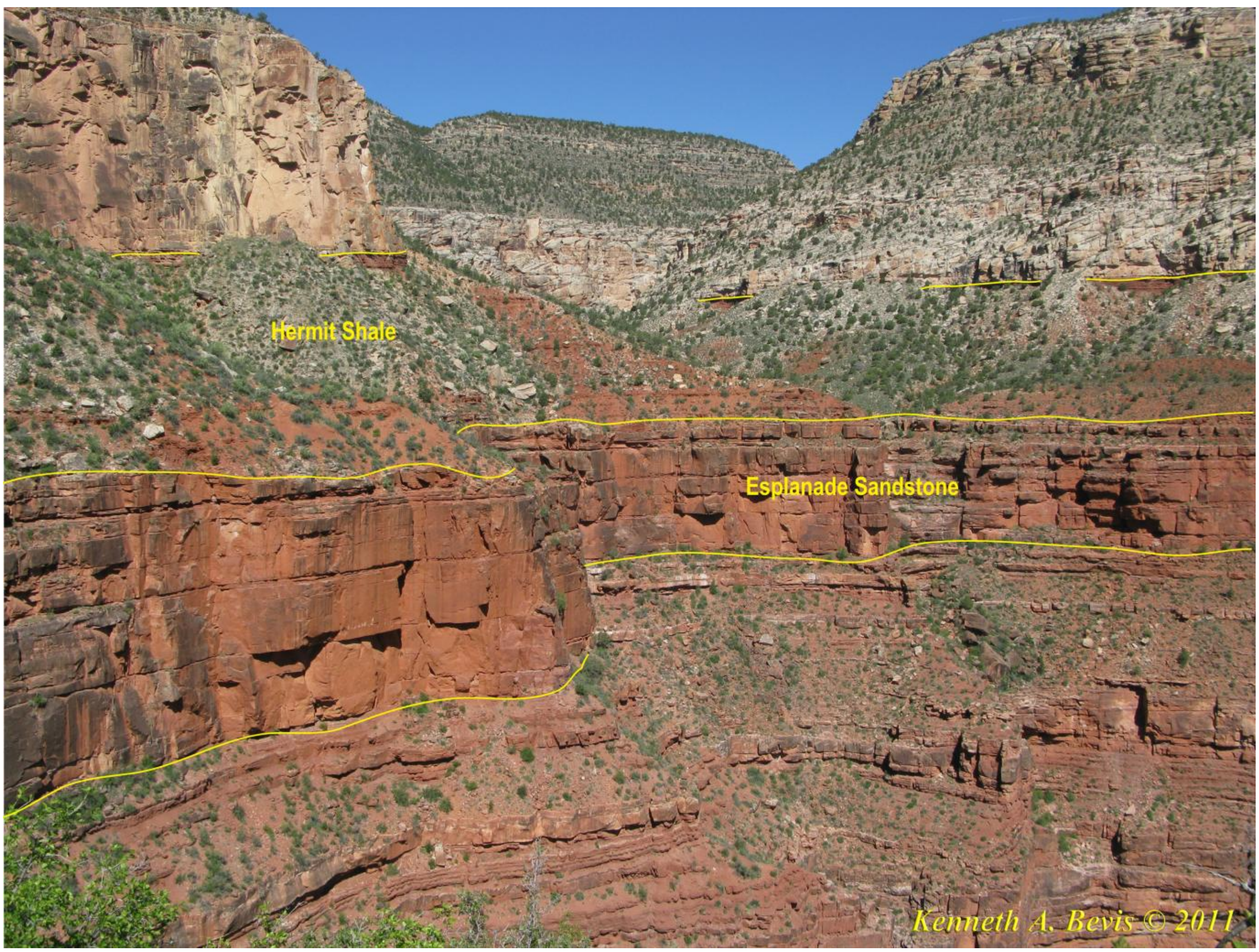


Figure 20. The Esplande Sandstone forms a prominent cliff-band below the Hermit Shale's equally distinct reddish mudstone slopes; the view here is from the Boucher Trail looking west across upper Hermit Canyon toward Dripping Springs.

The Supai Group is "sparsely and sporadically fossiliferous" (Blakey, 2003). The fossil types consist of trace fossils—including "trackways, burrows, impressions, resting marks, and feeding marks,"— invertebrates, and flora (Blakey, 2003). The abundance of swirls, crinkles, and irregular patterns make bioturbation the most common trace fossil in the Supai Group. Burrowed structures appear most commonly in silty sandstones and carbonates in a variety of types, but cylindrical, smooth burrows are dominant. Trackways occur specifically in the Wescogame Formation and the Esplanade Sandstone, usually created by terrestrial vertebrates. The fossil invertebrates are observed most commonly from the Watahomigi formation in brachiopod fauna, and marine fossils occur most often in deposits of the western Grand Canyon. However, any presence of abraded fossils in fine sandstone or limestone should not lead one to assign a marine deposition to the formation. The sedimentological evidence of eolian deposition suggests that a much more likely factor for the deposition of these particular fossils involves wind deflation of subaerially exposed marine sediments. Finally, flora is sparsely distributed throughout these formations and suggests coastal floodplains created during regressions (Blakey, 2003).

The origin of deposits that comprise the Supai Group is complicated and a subject still under consideration. The initial thought included a compilation of fluvial, deltaic, beach, shallow-marine, or estuarine origins, but after the evidence of wind ripples was identified, an eolian depositional setting seems much more plausible for much of the material. Controls during the deposition of the Supai Group were very complex, including the continuous eustatic rise and fall in sea level, regional subsidence, and an influx of sand from the north. As Blakey (2003) states, "The heterogeneity of much of the Supai is directly related to this complex intermixing of environments coupled with a broad range of available depositional material". In general, the Supai Group seems to have been deposited on a broad coastal plain associated with a passive continental margin, with individual settings ranging from shallow marine to continental depending on fluctuating sea level (Figure 21). However, it is thought that the eolian dune deposits occurred in a coastal-plain setting. Quartz sand was transported from the north by wind, but the continuous transgressions and regressions of the Pennsylvanian and Early Permian periods interrupted eolian deposition, preventing full development. While many carbonates were clearly deposited from a shallow-marine origin, minerals such as calcium carbonate can be precipitated within sediments as a result of extreme aridity which causes salty groundwater to rise upward by capillary action and evaporate (Blakey, 2003).

Though debate continues about the Supai Group's origins, its formations were likely deposited on a passive continental margin that experienced multiple, transgressive and regressive cycles of short duration. Environments transitioned from shallow marine to continental across a broad coastal plain, but became increasingly characterized by eolian deposition upward through the sequence. While blowing sand associated with an arid climatic setting dictated much of the deposition during the Pennsylvanian and Early Permian in the North American southwest, global tectonics also played a significant role. Assemblage of the supercontinent Pangaea was completed during this period, and compression generated by plate collisions along the southern and southwestern edge of the North American plate penetrated far inland to form the Ancestral

Rocky Mountains (Blakey and Ranney, 2008). Erosion of the Uncompaghre Uplift, a western extension of the Ancestral Rockies in western Colorado and eastern Utah, generated copious amounts of sediment that accumulated to vast thicknesses on the eastern Colorado Plateau well into the early Mesozoic. Deposits mainly accumulated in the Paradox Basin on the west side of the Uncompaghre highlands. To the west, in the Grand Canyon region, deposition from these eastern highlands was intermittent and limited to distal, fine clastics deposited on a continental shelf setting. Sandstones and mudstones of the Supai mark the first major influx of clastics into the region since the Middle Cambrian, provided for the most part by erosion of the Ancestral Rockies (Blakey and Middleton, 2012). Repeated glaciations in the Southern Hemisphere, where much of the Pangean landmass was located, probably induced the sea level changes experienced on western North America's passive margin as multiple, short-term marine transgressions and regressions. During higher sea levels, carbonate platform conditions overwhelmed clastic deposition to prevail in many areas, and during lower sea levels, clastic deposition in the form of eolian sands dominated. Over 150 million years, eolian deposits accumulated during regressions, were plained-off by transgressions or minor fluvial events, and repeated.

Deposition of the lower two-thirds of the Watahomigi Formation likely occurred in a shallow-marine to coastal plain setting along a fluctuating, low-energy shoreline (Figure 21a) (Blakey and Middleton, 2012). The mudstone was probably deposited on along a shoreline tidal flat. It was followed by carbonate deposition as the sea level rose or the Grand Canyon region subsided, with cycles of fine clastic influx followed by erosion marking short-lived regressions. Sediment accumulation was initially restricted to the Antler foreland basin until transgression allowed shallow marine conditions to invade much of the Colorado Plateau region (Blakey and Ranney, 2008). The upper third of the Watahomigi Formation and the entirety of the Manakacha Formation experienced periodic marine transgression into the Grand Canyon region paired with an increase in eolian material from the north during temporary regressions (Figure 21b). Periodic marine flooding of a broad, arid- and eolian-dominated coastal plain prevailed (Blakey and Ranney, 2008; Blakey and Middleton, 2012). After a relatively significant regional unconformity developed during prolonged subaerial exposure during the Middle Pennsylvanian, deposition of the Wescogame Formation in the Late Pennsylvanian marked a substantial change to more pervasive continental conditions paired with eolian deposition across the Colorado Plateau region (Blakey and Ranney, 2008) (Figure 21c). Minor transgressions paired with carbonate platform deposition continued to penetrate inland from the west giving the Wescogame a similar character to the Manakacha below (Blakey and Middleton, 2012). The erosion surface at the upper contact of the Wescogame, representative of the Pennsylvanian-Permian boundary, is thought to have been caused by streams crisscrossing newly exposed terrain resulting from a lowering of sea level. Finally, during the Early Permian, widespread eolian deposition formed the Esplanade Sandstone to the east, while the Pakoon Limestone member of the Esplanade Sandstone formed in shallow-marine, carbonate platform environments of clear water to the west (Blakey and Middleton, 2012) (Figure 21d). Interruptions were again caused by fluctuations in sea level coupled with fluvial activity, eventually transitioning to fluvial conditions as the eolian supply diminished (Blakey, 2003). Overall, the history of depositional settings of the Supai Group is very complex, and as studies continue more definitive, alterations will likely be made to the current interpretations described above.

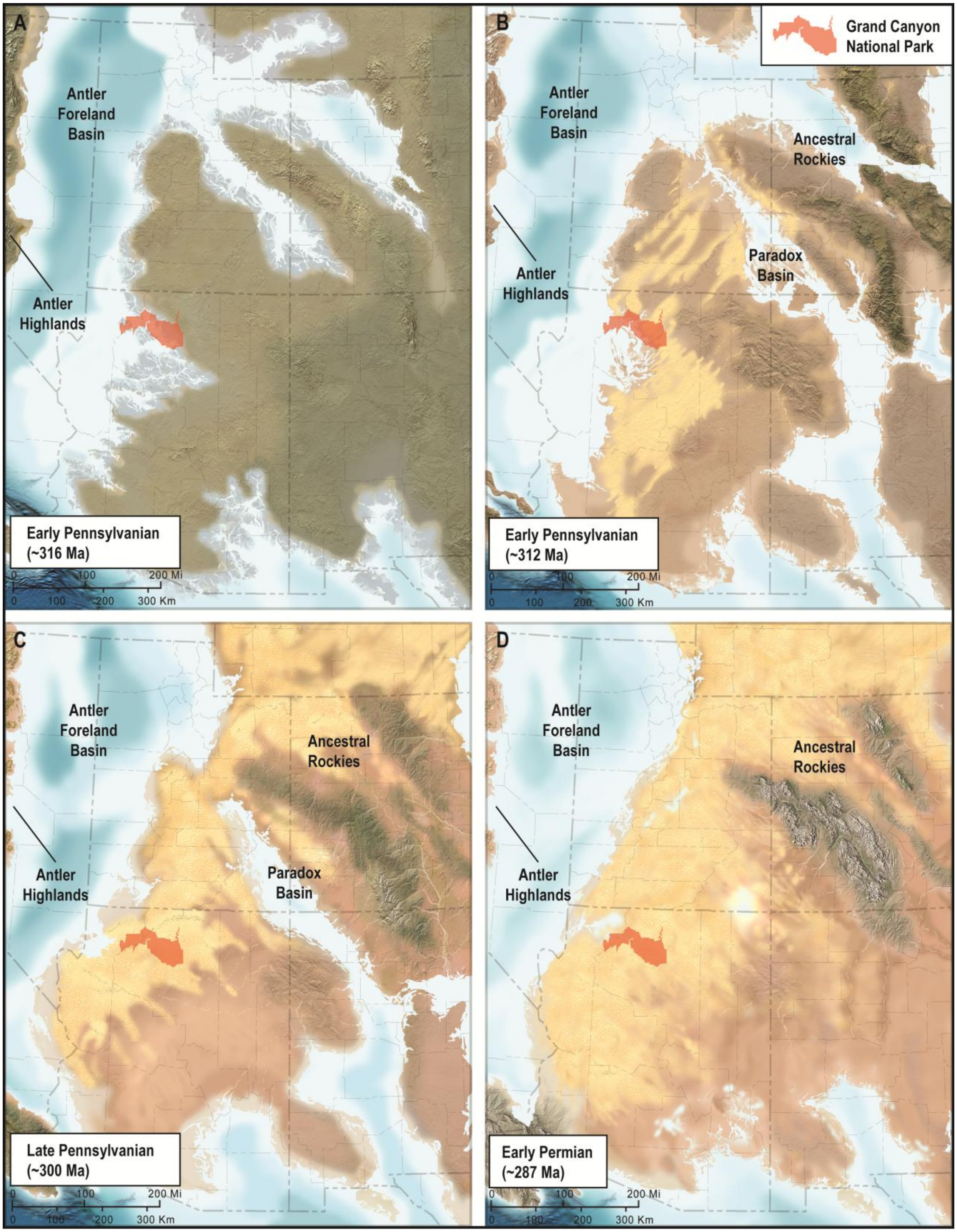


Figure 21. Early Pennsylvanian through Early Permian paleogeography of western North America during deposition of the Watahomigi Formation (A), Manakacha Formation (B),

Wescogame Formation (C), and Esplanade Sandstone (D), respectively; original maps are by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

The Supai Group was deposited on a passive continental margin. Recurring marine transgressions and regressions of the Pennsylvanian and Early Permian periods coupled with the influx of eolian sands caused a battle to ensue for depositional dominance. The sea would regress and eolian dunes would form on the broad coastal flat, only to be interrupted by a brief transgression, limiting eolian development. Carbonate units in the Supai Group are not solely derived from marine origin, but also of eolian (arenaceous limestone), and their interbedding is still not fully understood. To add complications, as the sea transgressed and regressed, debris was washed into the area by ephemeral streams, trapping mud and fine-grained sediment in lagoons, tidal flats, and river channels. Eolian deposition definitely played a significant part in the deposition of the Supai Group, but wind-generated sediments were continuously reworked, altered, or limited in extent by changes in sea level.

The Hermit Formation (by Meagan Redmond and Ken Bevis)

Along the upper walls of the Grand Canyon, the Hermit Formation occupies a distinctly reddish, slope-forming, bench above the Esplanade cliff-band and at the base of the contrasting white cliffs of the Coconino Sandstone. The formation overlies the equally reddish cliff bands of the Esplanade Sandstone, the last formation of the Supai Group, and the lowermost Permian rock unit in the canyon (Figure 19 and 20). All five of the sedimentary rock units from the Esplanade Sandstone, upward through the Kaibab Formation that creates the uppermost sedimentary rock layer within the Grand Canyon's walls accumulated during the Permian (Blakey and Middleton, 2012). Though many of the Permian age formations are well studied and documented, the Hermit Formation and the Coconino Sandstone are the least explored, with the Hermit Formation in particular being described as "one of the poorest known units in Grand Canyon" due to a seeming lack of interest by geologists in studying the formation (Blakey, 2003).

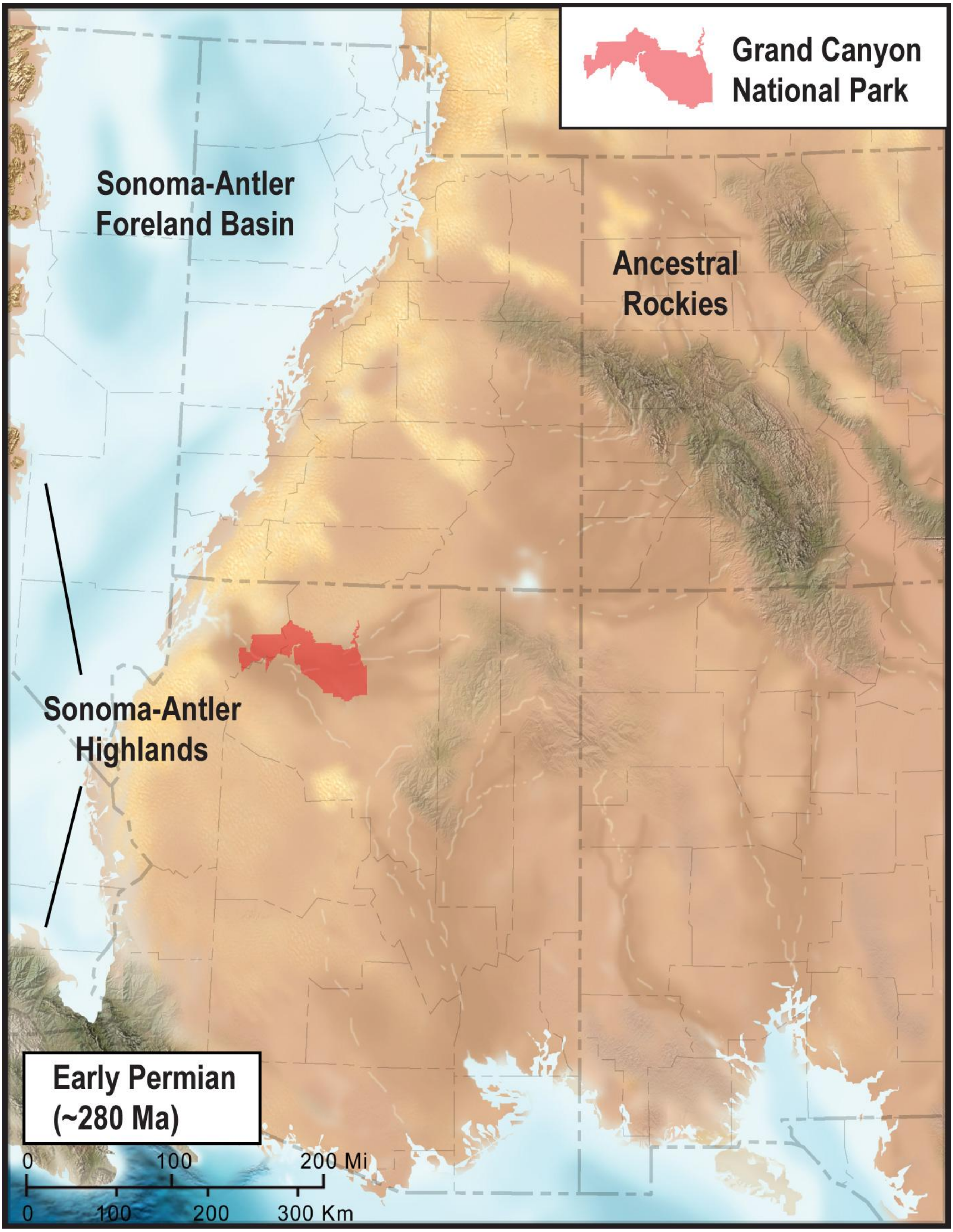
The Hermit Formation was once considered a member of the Supai Formation, mostly due to the fact that like the other formations of the group, the Hermit has a predominantly red strata and its depositional origin was thought to closely resemble that of the Supai Group's (Blakey, 2003). In 1922, the Hermit was seperated from the underlying Supai Formation as the Hermit Shale, and then in 1975 geologist Edwin McKee turned the Supai Formation into the Supai Group it is today, with its four formations (the Watahomigi, Manakacha, and Wesogame Formations and the Esplanade Sandstone) and renamed the Hermit Shale as the Hermit Formation. With that being said, almost all geologic sources that do discuss the Hermit Formation have a tendency to center their focus on only a few choice topics, such as the Hermit's composition and appearance in outcrop, and its relationship to the Supai Group.

Prior to Hermit Formation deposition, the western portion of the North American continent during Pennsylvanian time, consistently occupied a passive marginal setting on which sea levels rose and fell in a number of cycles of marine transgressions and regressions. The Grand Canyon region lay near the center of this transgression-regression activity and because of this, much of the geologic history of these sea level rises and falls are recorded within the Grand Canyon's formations. Within the Supai Group, an overall gradual marine regression punctuated by occasional minor transgressions is recorded. Beginning with deposition of the oldest formation of the group, the Watahomigi Formation, the Grand Canyon area "subsided and once again was covered by shallow-marine and coastal-plain sediments" (Blakey and Middleton, 2012). For a time, water transgressed into the area, and large influxes of sediment were being carried in from the north, likely coming from the rising Ancestral Rocky Mountains (Blakey and Middleton, 2012).

As the formations progress over time within the Supai Group, the area becomes more and more terrestrial, and this evidence is concluded with the uppermost formation of the group, the Esplanade Sandstone. The Esplanade Sandstone "contains the highest percentage of sandstone of any of the units in the Supia Group," and much of the sand is eolian in nature, suggesting wind-blown deposits and, thus, a dry, largely terrestrial environment (Blakey, 2003). In the end, this change from largely marine to largely terrestrial domination of the western coast of North America was the result eustatic sea-level changes, and not related to tectonics. The waning cycles of glaciation and interglaciation that lead to overall static sea levels, combined with the large influxes of sediment from the Uncompaghre highlands to the east, eventually lead to the area becoming more arid and terrestrial.

After deposition of the Early Permian eolian sandstones of the Esplande, all previous basins across the Colorado Plateau had been overwhelmed and filled with sediment (Blakely and Ranney, 2008). Lack of accomodation space may then have induced a phase of erosion. The Early Permian Hermit Formation is seperated from the slighly older Esplanade Sandstone by what McKee called a "regional unconformity" (Blakey, 2003). As Blakey (2003) describes, in some areas there is a definable transition from the underlying Esplanade to the overlying Hermit. Yet, in most locations at or near the Grand Canyon, the Esplanade and the Hermit are seperated by "an erosional surface with 30 to 50 feet of local relief." Other locations also indicate channeling into the Esplanade Sandstone; these channels are filled with Hermit sediments, and thus "originate from within the Hermit and not from the base [of the Esplanade]," and most of these channels are filled by sediments affiliated with the Hermit Formation (Blakey, 2003). This erosional relief, therefore, "may be related to channel cut-and-fill sequences," meaning that as stream channels cut across the Esplanade Sandstone, the Hermit Formation was also beginning to be deposited, locally filling in some of these channels with sediment.

The main body of the Hermit Formation is comprised of brick-red sandstone and mudstone, with minor pebble conglomerate and eolian loess and sand dune deposits associated with fluvial settings on a broad coastal plain (Blakely and Ranney, 2008; Blakely and Middleton, 2012). The rock unit formed as arid-land river systems spread sediment southwestward from the Uncompaghre uplift; the sandstones and conglomerates compose river channels deposits, while the mudstones were deposited on intervening floodplains (Figure 22). Early Permain Hermit deposition was a time of crustal stability in the Grand Canyon area, although areas to the south and west were beginning to subside in a forearc basin as the complex McCloud arc initially began to collide with the North American plate; this collision would eventually generate the Sonoma orogeny and reinvigorate the older Antler highlands (Blakely and Ranney, 2008).



Grand Canyon National Park
Sonoma-Antler Foreland Basin
Ancestral Rockies
Sonoma-Antler Highlands
Early Permian (~280 Ma)
0 100 200 Mi
0 100 200 300 Km

Figure 22. Early Permian paleogeography of western North America during deposition of the Hermit Formation; original map is by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

The Hermit Formation was deposited about 280 million years ago and is named after the Hermit Basin along Hermit Canyon (Wuerthner, 1998). According to the US Geological Survey, this formation is composed of "red, slope-forming, fine-grained, thin-bedded siltstone and sandstone." Blakey (2003) also describes the Hermit as consisting mostly of "interbedded, silty sandstone and sandy mudstone." The dark red-orange slope-forming characteristics of the Hermit help to distinguish the formation from the overlying tan-colored Coconino Sandstone (Timmons, 2003). However, the Hermit Formation's "fine-grained nature" causes the formation to be more suseptible to erosion, leading to "relatively poor exposures" and making the formation somewhat difficult to study; thus, the relative lack of interest by geologists (Blakey, 2003). The Hermit Formation is about "100 feet thick in the eastern portion of the Grand Canyon and near Seligman to over 900 feet in the Toroweap and Shivwits Plateau areas" and can average about 300 feet thick in the Mongollon Rim region east of Seligman (Blakey, 2003). The rock unit comprises a heterogeneous, chiefly a fine-grained, silisiclastic sequence, with neither vertial or lateral distribution of its components suggesting any distinct trends (Blakely, 2003). Silty sandstone and sandy mudstone are the most volumious and widespread sediments. The units are interbedded rhythmically, although mudstones generally increase upward within the formation. Sandstones range from structureless to ripple cross-laminated and trough cross-stratified; ledge-forming beds average 3 feet in thickness and are structureless. Ripple cross-laminated and trough cross-stratified layers are associated with point bar deposition in meandering streams. Sandy mudstone forms slopes in the Hermit Formation and is commonly featureless.

The Hermit Formation is sparsely fossiliferous, but a broad range of fossils have been identified (Blakely, 2003). Trace fossils, including burrowed structures, possible feeding and resting marks, invertebrate trackways, and bioturbation by plant roots are common. Plant fossils are widespread but few in number within the Hermit, but they do suggest the presence of broad floodplains developed on a semiarid to arid landscape. Timmons (2003) suggests that because of the continuous sea level fluctuations that had been occuring up to and during this time, "shallow marine, lagoon, and fluvial environments" were possible. She also goes on to say that the Hermit Formation, as well as the other formations of Permian age, were "the first layers deposited in an eolian (wind-dominated) environment in the Grand Canyon region. These layers mark the onset of the coastal sand dune environment that was dominant in the region through the late Jurassic Period" (Timmons, 2003). Blakey and Middleton (2012) indicate that the depositional environment for the Hermit Formation was likely a "broad coastal plain, chiefly in fluvial settings, but also wind-blown dust (loess) and scattered dune deposits of eolian origin" (Figure 22). The eolian deposits that contribute to the formation are also present within the overlying Coconino Sandstone, but in much larger quantities deposited in an "arid, inland dune setting" (Blakey and Middleton, 2012). This suggests that marine regression was approaching its maximum with the deposition of the Hermit Formation, and would later hit its peak with Coconino deposition. During this time, most of Colorado was dominated by the Ancestral Rocky Mountains, and westward draining "ephemeral river systems" were scattered over much of southwestern North America, with a shallow-marine shelf environment restricted to northwestern Utah (Figure 22). Throughout the Grand Canyon area, the contact between the Hermit Formation and overlying Middle Permian Coconino Sandstone is sharp and without gradation everywhere it has been observed (Blakely, 2003). The top of the Hermit is distinguished by descication cracks up the 20 feet deep that are frequently filled with the quartz sand of the Coconino, suggesting a rapid and intense return to aridity in Coconino time.

The Hermit Formation has undergone only limited investigation by geologists for the simple fact that its degraded slopes make observation difficult. The Hermit Formation was deposited in an arid climate, possibly on a broad coastal plain where fluvial settings dominated, but eolian sands continued to be intermittently carried across the Four Corners region. Tectonics did not play a critical role in the Hermit Formation's deposition; although the Sonoma orogeny caused areas to the southwest to subside as the McCloud arc collided with the North American plate. Eustatic sea-level changes associated with Southern Hemispheric glaciation continued to produce cyclical flucuations in sea level, periodically exposing coastal areas to wind erosion and sand transport. The overall marine regression that began at the start of Supai Group deposition is likely the underlying reason for the Hermit Formation's deposition. Factors such as the presence of mudstone interbedded with sandstone, an overall fining-upward sequence within the formation, and sparse terrestrial plant fossils, indicate that the Hermit was formed in an environment that was somewhat wetter than underlying and overlying eolian sandstones, but also one in which initially higher energy streams where gradually replaced by lower energy streams. It's dark red, gentle slope-forming appearance sets it apart from the overlying Coconino Sandstone, but an erosional disconformity seperates it from the otherwise similarly colored underlying Supai Group. The paleogeography of the Colorado Plateau region during the time of Hermit Formation deposition was likely a terrestrial-dominated area with a large, shallow body of water lying to the northwest. This trend of water transgressing and regressing from the northwest into and out of forearc basin created by the Antler-Sonoma orogenies would continue into the Triassic, although rock units of that age are almost entirely eroded from the Grand Canyon area.

The Coconino Sandstone (by Meagan Redmond and Ken Bevis)

Relatively little is known about the poorly studied Coconino Sandstone formation of the Grand Canyon (Middleton, et al., 2003). The formation is Early Permian in age, its deposition dating back to around 275 million years ago. Other formations within the Grand Canyon formed during the Early Permian include the Esplanade Sandstone of the Supai Group and the immediately underlying Hermit Formation. Much of the Early Permian was a time of only minor tectonic activity on North America's western passive continental margin; although by the latest Early Permian areas to the southwest were actively subsiding in a forearc basin as the Sonoma orogeny progressed associated with collision of the McCloud arc with the passive margin of the North American plate (Blakely and Ranney, 2008). Deposition of the Coconino and prior Permian age formations is considered to have been chiefly in response to "glacial-eustatic sea-level oscillations." Deposition of these rock units was affected by global-wide glacial activity that caused sea levels to rise and fall (Hopkins and Thompson, 2003). The Coconino Sandstone was deposited during a time when sea level had receded westward, and the Grand Canyon area was dominated by a terrestrial, desert-like environment. Unlike other Permian formations within the Grand Canyon, the Coconino is not subdivided into members.

The Permian west coast of North America underwent several cycles of marine transgressions and regressions while undergoing a major period of overall regression that ultimately led to the deposition of the terrestrial Hermit Formation and the Coconino Sandstone which had begun as far back as Wescogame Formation deposition, during the Late Pennsylvanian period (Blakey and Middleton, 2012). Coconino deposition occurred during a time of extreme aridity, with eolian sands covering two-thirds of Arizona and New Mexico, and yet "local high water tables furnished enough water to allow a diverse reptile fauna to exist" (Blakey and Middleton, 2012). No sedimentary rocks of the latest Early Permian are preserved in the walls of the Grand Canyon, presumably this stable shelf area simply acted as an eolian transport surface for sand dunes because it lacked the capacity to retain sediment. Equivalent latest Early Permain rocks of the Schnebly Hill Formation and De Chelly Sandstone were deposited in the Holbrook basin to the southeast were subsidence allowed accumulation of laterally adjacent and intertonging sand dune and tidal flat deposits.

The Coconino Sandstone forms a thick, light-colored cliff band directly above the Hermit Formation's contrasting reddish slopes (Figure 23). Taken in sequence, the rock unit is easily recognized from any North or South Rim overlook as the third layer from the top, where it forms the second riser in a series of cyclopethan stair steps descending into the canyon; the top step is the rim, the top riser is formed by the rim-capping Kaibab Limestone, the second step formed by the Toroweap Formation's well- vegetated slopes, the second riser the Coconino, the third step the Hermit Formation's brick-red slopes, and so on. Blakey and Middleton (2012) describe the Coconino Sandstone as pale-yellow in color, consisting "almost exclusively of large-scale cross-bedded sandstone." This sandstone is composed of "fine-grained, well-sorted" sand with large amounts of quartz and some feldspar, and the formation overall has a tendency to form massive, "high angle" cliffs in the Grand Canyon (Middleton, et al., 2003). McKee (1979) describes the lithology of the Coconino Sandstone as dominantly quartz arenite with a siliceous cement of a white to pink color with sparse amounts of feldspathic sediment present throughout. The grain sizes range from fine to medium and are well sorted, and the grains are well rounded to very well rounded with frosted surfaces that are indicative of wind transport (Blakely, 1990). These characteristics are indicative of mature grains that have undergone substantial transportation, opposed to immature sediments which would contain greater lithics and feldspars, would be less rounded, more poorly sorted, and consist of coarser particle sizes. McKee (1979) indicates that the predominant sedimentary structure found in the formation is steeply dipping, large-scale or mega-crossbedding, although well preserved ripple marks, and slump structures also occur.

Figure 23. The buff-colored cliffs of the Coconino Sandstone are easily distinguished from the overlying and underlying slopes of the Toroweap and Hermit Formations, forming the second riser (third layer) in sequence of a massive set of stair steps leading downward into the Grand Canyon; viewed at two locations, (A) Shoshone Point at sunrise, and (B) from the Boucher Trail looking up Hermit Canyon.

The Coconino Sandstone exhibits significant thickness variations which are generally inconsistent with those of its overlying Permian counterparts, the Toroweap and Kaibab Formations (Blakely, 1990). Bucking the normal trend of rock units in the Grand Canyon, it thickens southeastward and is 65 feet thick near the Grand Wash Cliffs at the western end of the canyon, 96 feet thick at Toroweap Point, and 300 feet thick in the Bright Angel Creek area (Babcock et al., 1974). At the central part of the canyon where the formation is thickest, the Coconino is 600 feet thick near Cottonwood Creek (Middleton, et al., 2003). Overall, the formation thins northward and westward. The directionality in which this “dry” formation thins indicates where marine conditions would most likely have prevailed; in studying sedimentary deposits, geologists know that where sand dunes and other associated terrestrial facies are thickest, the drier the area, suggesting that the depositional environment was more strongly influenced by terrestrial facies deposition. Where the terrestrial facies begin to thin, “wet” facies become thicker, indicating marine domination in the area. Because the Coconino Formation consists exclusively of homogeneous sands that consistently exhibit large scale cross-bedding, it is widely accepted that the depositional environment of this rock unit largely resembled wind-blown sand deposits spread over most of Arizona and New Mexico in an enormous desert erg, much like the African Sahara of today (McKee, 1979; Blakely, 1990; and Middleton, et al., 2003).

However, evidence within the Coconino suggests that there was some surface water present at times. As Middleton et al. (2003) explain, not only does the Coconino Sandstone have trace fossils of reptile footprints, but within the sedimentary structures of the Coconino there are “rain drop impressions.” Much like a modern desert, any rainfall that came during Coconino deposition would have been sparse and short-lived, and the reptiles that thrived in this environment would have been small and hardy, able to live on very little water. Middleton et al. (2003) also describes ripple marks present within the Coconino, and many geologists interpret that the ripple marks were likely the result of active crosswinds (Babcock et al., 1974).

Coconino deposition suggests that the paleogeography of the Colorado Plateau region was terrestrially dominated (Figure 24). The Ancestral Rocky Mountains and Uncompahgre Highlands, formed initially during latest Mississippian-Early Pennsylvanian period, dominated most of Colorado and parts of Utah and New Mexico to the east, while the regrowth of highlands to the west was underway related to the ongoing Sonoma Orogeny. A huge desert covered much of Utah, New Mexico, and Arizona in between remnant and growing highlands (Figure 24). Marine and shoreline environments would have been present only at the uppermost corner of Utah and at the southernmost ends of Arizona and New Mexico. These neighboring bodies of water would later transgress inland to inundate the region during Toroweap and Kaibab deposition, and marine depositional environments would transform the Grand Canyon area, ending Coconino deposition (Blakey and Middleton, 2012).

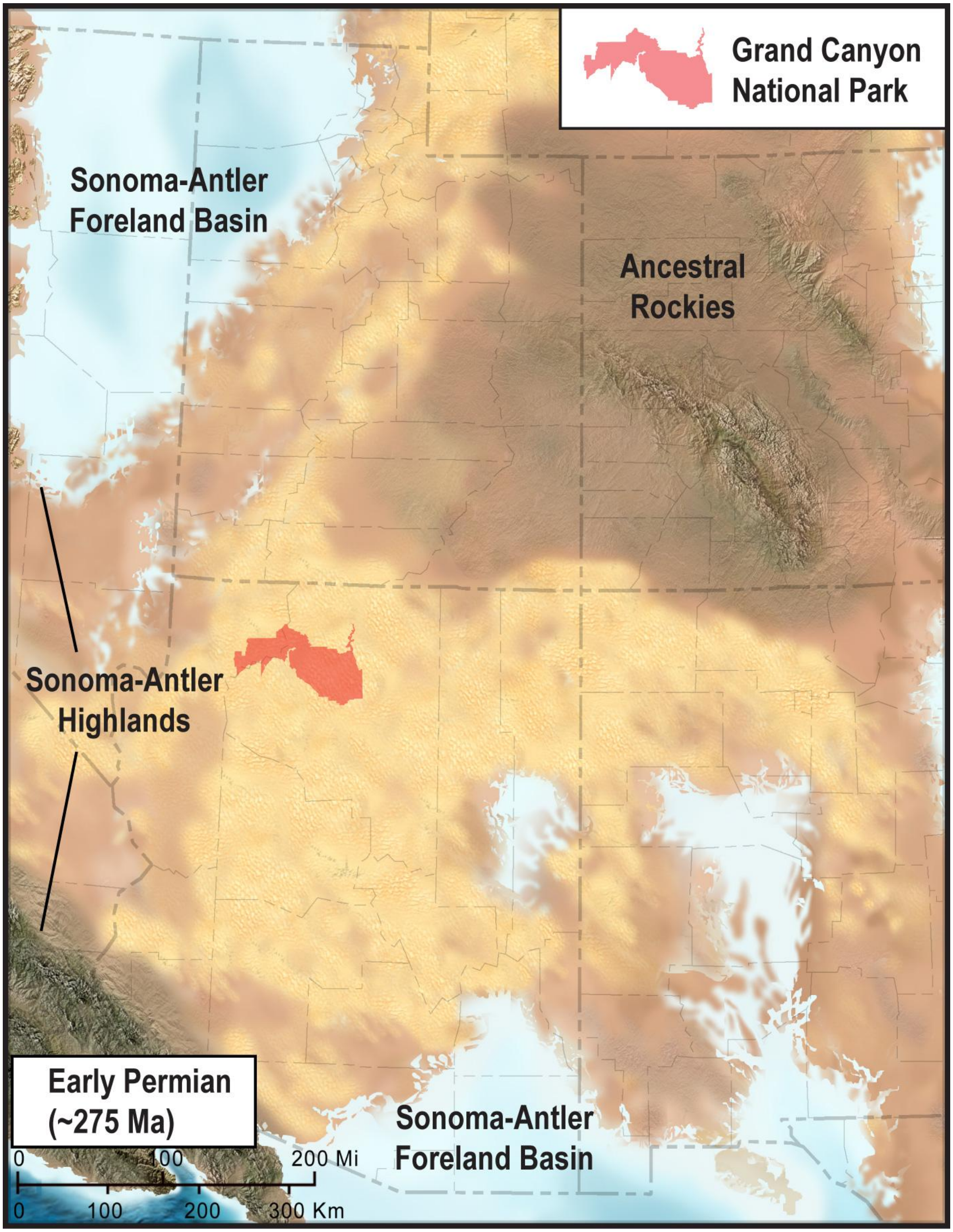


Figure 24. Early Permian paleogeography of western North America during deposition of the Coconino Sandstone; original map is by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

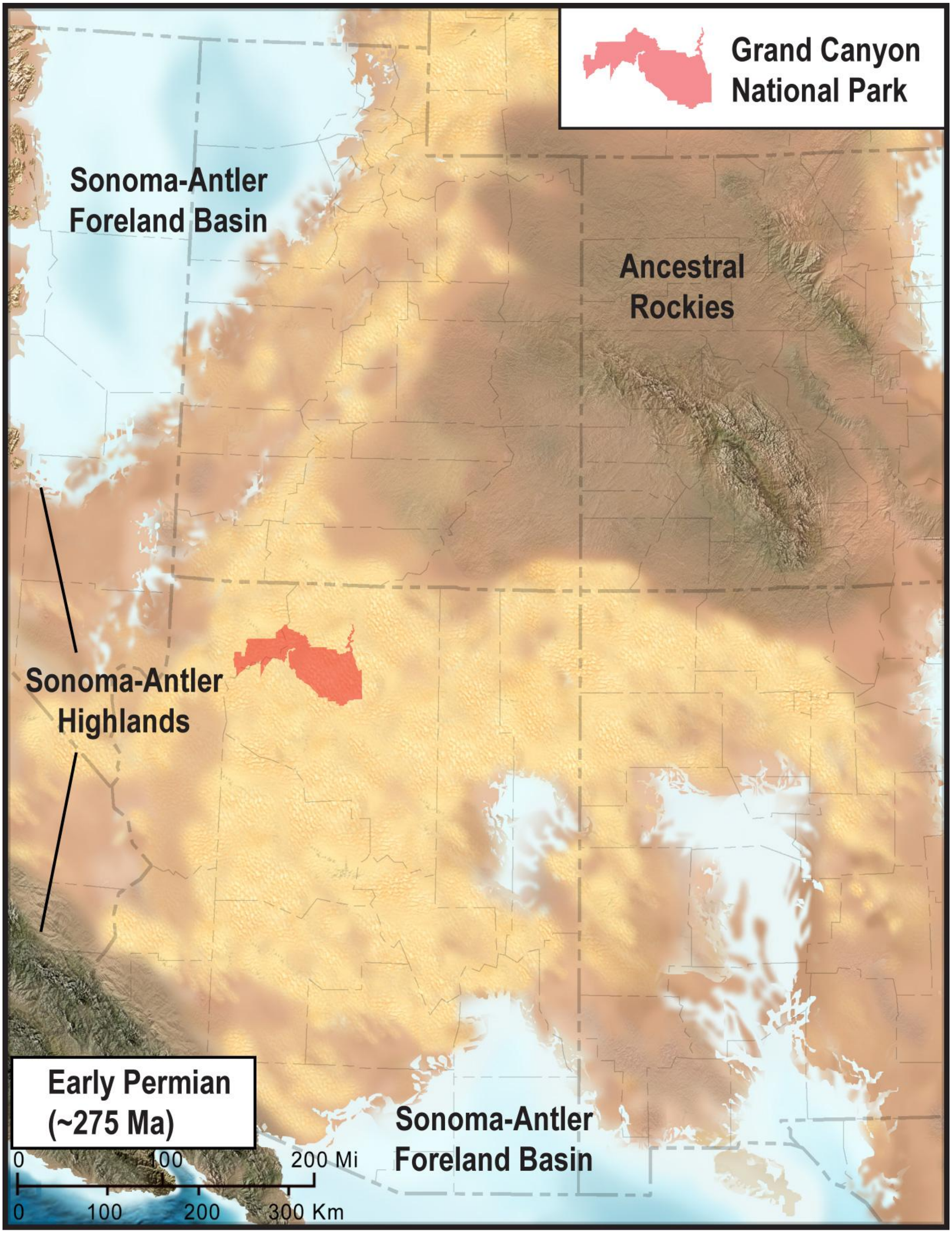
Even though relatively few geologists have studied the Coconino Sandstone, a deciphering of its sedimentology and paleontology indicates that it was deposited in a terrestrial setting dominated by an arid climate where there was relatively little water, except for an occasional brief rain. The Coconino Sandstone exhibits textbook examples of sand dune facies deposited by wind and its lithology, texture, sedimentary structures are all strong evidence of this conclusion. Similar to a modern desert, wind-blown eolian sand dunes dominated the Four Corners region and only hardy organisms such as reptiles and arachnids could live in this harsh environment. Coconino deposition took place at the very end of a significant period of marine regression, and its deposition was eventually terminated by a return of marine conditions to the Four Corners region as a transgression swept across the Grand Canyon area from the west to deposit the Toroweap and Kaibab Formations.

The Kaibab Limestone and Toroweap Formation (by Meagan Redmond and Ken Bevis)

The Kaibab and Toroweap Formations form the uppermost layers exposed in the walls of the Grand Canyon (Figure 23); recognized as a distinct pair of cliff and slope bands just below the rim. Both rock units were formed during the Early to Middle Permian period of the Paleozoic era (Turner, 2003), which began about 286 million years ago and lasted until 248 million years ago. Other Permian formations deposited within the Grand Canyon area include the slightly older Early Permian Coconino Sandstone and Hermit Formation; Late Permian rock units are not preserved (Blakey and Ranney, 2008). The overlying Kaibab and underlying Toroweap Formations are so compositionally similar and stratigraphically related that the two were originally paired by geologists into one formation, the Aubrey Limestone, which was later changed to the Kaibab Limestone. It would not be until 1938 that geologist Edwin McKee would ultimately split the Kaibab Limestone into the two formations that we see today, the upper Kaibab Formation and the lower Toroweap Formation (Turner, 2003).

Tectonic and sedimentation patterns developed earlier in the Pennsylvanian continued during much of Permian time (Fillmore, 2011). The Colorado Plateau occupied a broad, shallow-marine, continental shelf associated with the passive southwestern margin of the North America continental plate (Blakey and Ranney, 2008). Sediments continued to accumulate as a westward thickening wedge across the region on a gradually subsiding forearc basin comprised of thin crystalline basement and the accreted remains of the Mississippian Antler Orogeny and the ongoing Sonoma Orogeny (Figure 25). Although there is little to no direct evidence of tectonic activity in the Grand Canyon area occurring during Kaibab or Toroweap deposition, lying more or less in the center of a broad basin, the Permian period of the Colorado Plateau region as a whole was tectonically active. As Fillmore (2000 and 2011) describes, “The Permian saw the final assembly of the supercontinent known as Pangaea.” The Ancestral Rocky Mountains, the westernmost extension of the Marathon-Ouachita Belt that had formed during the Pennsylvanian in response to North America’s collision with the South American continent, continued to rise isostatically and influence regional drainage patterns (Fillmore, 2000 and 2011). More specifically, the Uncompahgre highlands and adjacent Paradox basin formed both the source and sink for vast quantities of sediment in the Four Corners area, and sediment was contributed distally to most if not all of the Grand Canyon’s Permian formational layers. Throughout Early

Permian time, subsidence of the Paradox basin would decrease gradually as it filled with westward-prograding sediments. By the end of the Permian period, the Uncompahgre uplift of the Ancestral Rockies was inactive, although sediment would continue to be generated and shed westward from this decaying highland well into the Triassic (Fillmore, 2011).


Grand Canyon National Park
Sonoma-Antler Foreland Basin
Ancestral Rockies
Sonoma-Antler Highlands
Early Permian (~275 Ma)
Sonoma-Antler Foreland Basin
0
100
200 Mi
0
100
200
300 Km

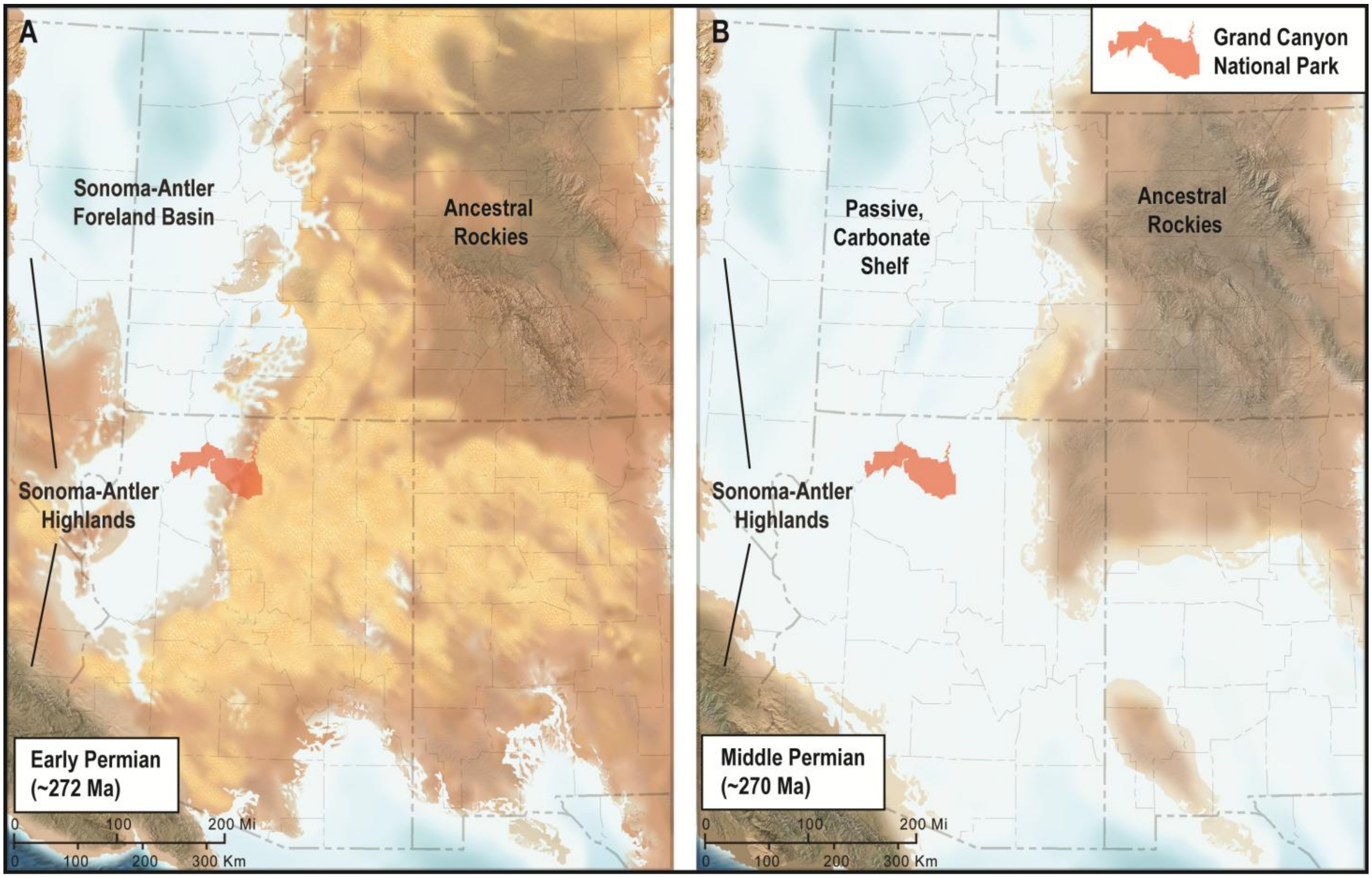


Figure 25. Late Early Permian and Middle Permian paleogeography of western North America during deposition of the Toroweap Formation and Kaibab Limestone, respectively; original maps are by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

Throughout the Permian and into the Triassic, various island arc terranes (generally known as the McCloud Arc) approached North America from the west, eventually colliding along east-dipping subduction zones to initially form a trough-like foreland basin that accommodated sediments from the east. Collision of arc terrains eventually formed the highlands of the Sonoma Orogeny (Figure 25), a probable erosional source on the west side of the basin that now existed between these growing highlands and the remnant Ancestral Rockies to the east (Blakey and Ranney, 2008). The region would experience a number of other mild uplifts during the Permian (Fillmore, 2011), but none would contribute much sediment to the Grand Canyon area's passive marginal setting. Instead, these uplifts mainly served to limit accommodation space and preservation of sediment in certain areas, especially during the Late Permian. Ensuing regional uplift prevented deposition or induced erosion of previously deposited sedimentary rocks over much of the Colorado Plateau for roughly the next 25 million years (until the Middle Triassic Period of the Mesozoic Era).

The Toroweap Formation underlies the Kaibab Limestone in the Grand Canyon (Figure 23), and the two formations are separated by a regional unconformity (Babcock, et al., 1974). The rock unit is distinguished as a slope-former because its higher mud content makes it relatively less resistant to erosion. The unconformity is an erosional surface, meaning that before the deposition of the Kaibab, the Toroweap Formation was subject to weathering and erosion through subaerial exposure, leading to the conclusion that marine conditions retreated from the Four Corners region for a time. The Toroweap as a whole was "deposited in open and restricted-marine environments, tidal flats, sabkhas, and eolian dunes fields" (Turner, 2003). Figure 25a depicts the paleogeography of these depositional environments. Turner further explains that "The Toroweap Formation, in its variety of lithofacies, reflects transgressions and regressions of an eastward-advancing and westward-retreating sea." This is further confirmed by Harris and Tuttle (2004), who describe that the Toroweap Formation's "nearly 200 feet of sandstone and limestone interbedded with gypsum records marine advances and retreats." Thus, through lithologic and fossil evidence within the Toroweap Formation, we can see that during its deposition sea level rose and fell multiple times, and the primary source of the water lay to the northwest, advancing southeast into the basin formed between the steadily eroding Uncompaghre and Sonoma-Antler highlands. This is also shown by the formation's distribution; it thins to the east and south from the Grand Canyon, presumably having less time to accumulate.

The Toroweap Formation has been divided into three members: alpha, beta, and gamma. These members from top to bottom include the Woods Ranch Member (or "alpha"), the Brady Canyon ("beta"), and the lowermost Seligman ("gamma") (Turner, 2003). The Kaibab Formation has been similarly subdivided and the top of the Woods Ranch Member ends where the "gamma" member of the Kaibab begins. We will discuss the members of the Toroweap in ascending order, so that changes in their composition can be associated with fluctuating depositional environments, sediment sources, and tectonic settings.

According to Babcock et al. (1974), the lowermost "gamma" member of both the Toroweap and Kaibab Formations can also be referred to as "the time of transgressing sea" and is a relatively thin unit. The Seligman Member marks the beginning of Toroweap deposition, and begins where the Coconino Sandstone Formation in the Grand Canyon ends. During Coconino deposition, the future Four Corners and Grand Canyon areas were mostly a terrestrial, desert-like environment with "enough water to allow diverse reptile fauna to exist" (Blakey and Middleton, 2012). However, as Babcock et al. (1974) suggest, marine conditions eventually returned to the region, and this transgression marked the beginning of Seligman Member deposition.

The Seligman Member largely consists of sandstone and evaporites, and is distinguishable in the Grand Canyon from the Coconino Sandstone because "it is the first non-cross-bedded" and evaporite-containing layer above the Coconino (Turner, 2003). In the Grand Canyon, this unit is less than 45 feet thick. The Seligman contains "thin-bedded dolomite, sandstone, and gypsum," with the sandstone and evaporites dominating southeastward and the dolomite becoming more apparent northwestward. This change in facies from northwest to southeast can be interpreted as "a transition from coastal sabkha environments on the northwest to a continental sabkha to the southeast" (Langenheim Jr. and Schulmeister, 1987). The evaporites within this member and the ripple marks in the sandstone indicate that the Seligman was likely deposited on a shallow continental shelf environment where "warm restricted-marine waters were evaporated, promoting the precipitation of evaporite minerals" (Turner, 2003). Essentially, the evidence within the Seligman suggests that the water level was deepest to the northwest, where ocean water was coming in and transgressing onto the North American continent. As the water became shallower to the southeast, it became cut off from its original water source and became shallower, allowing for evaporites to form. Evidence within all of the members of both the Toroweap and Kaibab, as well as in other formations within the Grand Canyon, suggest that the main water source was nearly always located to the northwest.

The Brady Canyon (or "beta") Member underlies the uppermost Woods Ranch (or "alpha") Member of the Toroweap, and is described by Babcock et al. (1974) as being "the time of maximum advance." It is during this time of Toroweap deposition that the marine transgression reached its peak. The Brady Canyon is the thickest member of the Toroweap within the Grand Canyon, and its 280 feet of thickness is almost entirely composed of dolomite (Turner, 2003). Dolomite would usually indicate that deposition occurred in deep water; however, the lithofacies within the Brady Canyon Member have a "mud-supported texture," which indicates deposition in quiet-water conditions consistent with the shallow water of a carbonate shelf (Turner, 2003). However, as the member moves southeast, the dolomite is "progressively replaced by wackestone, pelletal packstone, and quartzose dolomite," again showing the shallowing of seawater to the southeast (Langenheim Jr. and Schulmeister, 1987). The Brady Canyon is the only fossiliferous member of the Toroweap; its two major faunal facies are open-marine fauna to the west (brachiopods, crinoids, horn corals) and more restricted, molluscan fauna to the east (bivalves and gastropods). According to Turner (2003), these fossils within the Brady Canyon Member "suggest a late Leondardian age for the Toroweap Formation."

The contact between the lower Seligman Member and the Brady Canyon Member is described as being "on a regional basis, gradational" meaning that the Seligman gradually fades into the Brady Canyon (Langenheim Jr. and Schulmeister, 1987). However, the top of the Brady Canyon Member is more abrupt and easily distinguishable from the Woods Ranch Member, because at the top of the member there is dolomitic mudstone that has "abundant desiccation cracks" (Turner, 2003). These cracks indicate subaerial exposure, and this exposure indicates that at the end of the Brady Canyon Member's deposition, sea level dropped and water receded from the area temporarily.

The Woods Ranch ("alpha") Member is the uppermost member of the Toroweap and underlies the Kaibab Formation. Babcock describes the "alpha" member as being "the time of regressing sea." Much like the Seligman Member, the Woods Ranch primarily consists of the evaporite gypsum, sandstone, and limestone; yet, the Woods Ranch does contain some dolomite at its base, making it conformable to the underlying Brady Canyon. As Turner (2003) describes, "The Woods Ranch Member typically extends from the top of the aphanitic dolomite that forms the uppermost unit of the Brady Canyon Member to the base of the cliff-forming limestone of the overlying Kaibab Formation." The Woods Ranch is about 180 feet thick in the Grand Canyon, but unlike the other members of the Toroweap, the Woods Ranch "shows no consistent thickening or thinning trends" (Turner, 2003), meaning that it does not follow the typical pattern of thinning to the southeast or east like other rock units. Most likely, the Woods Ranch Member was deposited in a shallow shelf environment much like the environment that was present during Seligman Member deposition. The evaporites and carbonates within the Woods Ranch indicate periods when water was reintroduced back into the area, and the “eolian sandstones [within] suggest times of subaerial exposure” (Turner, 2003). The repetition of limestone, sandstone, and evaporate within the Woods Ranch Member’s deposition indicates that the member went through a type of “cyclic sedimentation”, possibly influenced not only by sea level but also by seasonal changes in depositional conditions. By the end of Woods Ranch Member, and ultimately Toroweap deposition, water completely receded from the area for a time and erosion took place, leaving behind the unconformity that separates the Toroweap Formation from the Kaibab (Turner, 2003).

The Middle Permian Kaibab Limestone is the uppermost rock unit within the Grand Canyon’s sedimentary sequence and is discernible from the lower Toroweap Formation by its grey, interlayered slopes and stepped-cliff appearance (Figure 23). It is likely that the Kaibab and not the overlying Triassic Moenkopi Formation is the cap rock of the Grand Canyon area because the Moenkopi had a “less resistant nature” than the Kaibab, and was ultimately eroded away along with the uppermost layers of the Kaibab from most of the Grand Canyon area (Hopkins and Thompson, 2003). The Kaibab Formation was named after the Kaibab Plateau on the North Rim of the Grand Canyon, where it was first discovered and is well exposed. As its name suggests, the Kaibab Limestone is composed primarily of limestone (Wuerthner, 1998), which can reach a thickness of anywhere from 300 feet and, in northwestern Arizona where it is thickest, an excess of 500 feet thick (Hopkins and Thompson, 2003). The Kaibab overall was likely deposited in a subtidal, shallow-marine environment where minor fluctuations in sea level could abruptly change the water depth and corresponding depositional environment (Figure 25b), much like the environment present at the time of Toroweap deposition. These changes were not due to tectonic activity, as Hopkins explains: “Considering the quiescent (inactive) tectonic setting of the Grand Canyon region during the Permian, it is most likely that these cycles were caused by glacial-eustatic sea-level oscillations.”

Because the Kaibab and the Toroweap were originally grouped together as one rock unit (McKee, 1938), the Kaibab was also divided into three members: the lowermost gamma, the middle beta, and the uppermost alpha. However, because the rocks within the lower gamma member were confined to the Mogollon Plateau south of the Grand Canyon, “Geologists [now] interpret the gamma member as a facies within the Fossil Mountain (or “beta”) Member” (Hopkins and Thompson, 2003). Even though the Kaibab technically has three members, only two distinct members are now recognized and named: the uppermost Harrisburg Member and the middle/lower Fossil Mountain Member. The Harrisburg Member is typically thinner than the underlying Fossil Mountain Member, but its actual thickness at the time of deposition is impossible to determine accurately because subsequent erosion has stripped away the member’s uppermost portion.

Although the lowermost “gamma” member is no longer formally recognized as its own distinct layer within the Kaibab, the unit is still significant for understanding the depositional environments of the other members of the Kaibab Formation. As with the basal unit of the Toroweap Formation, the gamma layer represents a time of sea transgression. An unconformity separates the Woods Ranch Member of the Toroweap Formation from the gamma member of the Kaibab, indicating that, for a time, water receded from the area and allowed for subaerial exposure and erosion to occur. Then water returned to the area, just as it had after the dry spell during Coconino deposition, when the arrival of water was signified by deposition of the Toroweap's Seligman Member. We see evidence of this in the Kaibab's gamma member because it consists of "sandstone and dolomite with a rich [marine] fossil assemblage" (Fillmore, 2000). As Fillmore explains, "The gamma member was laid down as the sea advanced eastward across Utah and Arizona….", and just as with the Toroweap, as marine waters advanced eastward, deeper water conditions would help to deposit the overlying beta member.

The Fossil Mountain Member is described by Fillmore (2000) as being a "massive, cliff-forming cherty limestone" that was deposited when sea level was at its highest. This member typically contains abundant and diverse marine-fauna such as brachiopods, bryozoans, crinoids, sponges, and solitary corals. However, as your location within the member moves eastward, the fauna changes to suggest more restricted conditions and the member becomes more siliciclastic-dominated, indicating yet again that the water source lay to the west (Hopkins and Thompson, 2003). As the member thickens westward it can become up to 300 feet thick; and though the Kaibab Formation as a whole is largely known for its limestone composition, the Fossil Mountain Member is "75 percent sandstone or sandy dolostone" with scattered chert, dolomite, and limestone throughout (Hopkins and Thompson, 2003). The Fossil Mountain was likely deposited on a shallow "Kaibab Sea" that regularly experienced sea level rises and falls, extending eastward into the Grand Canyon area during transgression and receding westward toward the continental margin during regression (Fillmore, 2000). The difference between the beta member of the Kaibab Formation and the beta member of the Toroweap Formation is that, during the time of Fossil Mountain deposition, water covered nearly all of Utah, Arizona, and most of southern New Mexico, making this transgression much more extensive than the one that occurred during the Toroweap's Brady Canyon Member (Blakey and Middleton, 2012).

The uppermost Harrisburg Member of the Kaibab Formation was formed during a time when the Kaibab Sea that had generated Fossil Mountain deposition was receding westward. As Hopkins and Thompson (2003) describe, "units that comprise the Harrisburg sequence reflect deposition within predominantly restricted-marine environments during cyclic westward retreat of the Kaibab Sea." These repeated "shifts" in depositional environments due to sea level change can be observed throughout the Grand Canyon region with the regular alterations between carbonate, siliciclastic, and evaporite deposits. The Harrisburg Member in particular consists of an array of gypsum, dolostone, sandstone, redbeds, chert, and minor limestone (Hopkins and Thompson, 2003). The large amount of evaporites present within the Harrisburg Member indicates that the climate was arid, the influx of water supply from the ocean was routinely restricted, and water was generally receding from the area (Fillmore, 2000). This restriction of water and associated unstable environment is also reflected in the member's fauna content. Within the Harrisburg Member, there are fossils of pelecypods, gastropods, some ostracods, and generally fauna that were "hardy individuals tolerant of a greater range in environmental conditions" (Hopkins and Thompson, 2003).

The Toroweap and Kaibab Formations compose the uppermost layers of the Grand Canyon's walls, forming a lower slope-forming band and upper cliff-forming band at the canyon's rim. Both rock units were deposited during a time of quiescent tectonic activity. During the time of their deposition, sea level rose and fell in a series of cycles influenced by the waxing and waning of glaciations occurring to the south. As sea level rose during the lowermost "gamma" member of the Toroweap, the Seligman Member, sand dunes that were left behind during the time of Coconino deposition were covered and the landscape began to change from terrestrial- to shallow marine-dominated. By the time of the “beta” Brady Canyon Member, sea level across the Grand Canyon area and greater Four Corners region had reached its peak, covering half of Utah and parts of northwestern Arizona. During deposition of the Woods Ranch Member, sea level began a step-wise fall and marine waters gradually receded to the west. Following Woods Ranch Member deposition, subaerial exposure of the region allowed for erosion to take place, indicated by an unconformity that separates the Toroweap Formation from the Kaibab Formation. The Kaibab’s gamma member marks a new phase of marine inundation of western North America and initiates another cycle of rising and falling sea level. The only significant difference between this new cycle of sediment deposition and that of the cyclic deposition of the Toroweap is that, during deposition of the Kaibab’s beta member, sea level rose enough to cover an even greater area, allowing for most of Utah, Arizona, and the lower half of New Mexico to be almost completely under water. However, depositional environments associated with both formations were unstable and ever fluctuating, indicated by the relative lack of fossil preservation.

<u>Mesozoic Sedimentary Rocks of the Grand Staircase</u>

The Chinle and Moenkopi Formations (by Meagan Redmond and Ken Bevis)

The Moenkopi and Chinle Formations comprise two stratigraphically related sedimentary rock units that formed during the Triassic Period of the Mesozoic Era, the time interval from 245 to 208 million years ago (Stokes, 1986). However, “no Mesozoic rocks are exposed within the Grand Canyon” (Colbert, 1974) and because they are not Paleozoic in age, the Moenkopi and the Chinle Formations do not usually appear within the Grand Canyon’s sedimentary sequence. Cedar Mountain, actually a relatively nondescript butte located near Desert View in the southeast corner of Grand Canyon National Park, offers the only outcrops of the Moenkopi and Chinle formations within the park’s boundaries (Figure 26). Yet Mesozoic rock formations do represent an important part of the Colorado Plateau’s geologic history, and they <u>*do*</u> crop out in areas near the Grand Canyon, making the formations important to include when discussing the geology of this region. For this reason, we include these early Mesozoic rock units in Figure 1. Alternating, stair-stepped, cliff-slope exposures of these Mesozoic rocks primarily occur to the north in an area aptly called the Grand Staircase, preserved for our viewing pleasure in such parks as the Vermillion Cliffs and Escalante-Grand Staircase National Monuments. The Moenkopi Formation is believed to have been formed during the Early (and possibly early Middle) Triassic Period, while it is inferred that Chinle deposition occurred during the Late Triassic (Reppening and Cooley, 1969), the two rock units being separated by a significant region-wide unconformity. To understand how these rock units were formed, it is first important to explore what was happening tectonically during the time of their deposition and to understand what the Colorado Plateau region’s landscape, in which the Grand Canyon is now situated, looked like (what were the rock unit’s paleogeographic and paleoenvironmental settings during the time of their formation?).

Figure 26. The only outcrops of the Triassic Moenkopi and Chinle Formations within the boundaries of Grand Canyon National Park occur at Cedar Mountain, in the southeast corner of the park; here viewed from the observation deck at Desert View.

The Triassic period is the first of three time intervals that comprise the Mesozoic era (the second being the Jurassic, which is followed by the Cretaceous). During the Mesozoic Era, which lasted about 183 million years, the world was changing as the supercontinent Pangaea fragmented and the Earth's distribution of landmasses and ocean basins began to resemble what we see today. The initial shifting of tectonic plates during the Triassic affected "western North America only indirectly because the region was far removed from the rifting margins of the breakup" (Fillmore, 2011). Yet the shifts still had some impact on the Colorado Plateau region as continental masses split apart to form constricted seaways that ultimately grew to form ocean basins, and marine transgressions and regressions came and went across the landscape, leaving their mark as the sedimentary rocks of the Mesozoic (Fillmore, 2011).

Throughout the Triassic period, the Ancestral Rocky Mountains (and the main Ouachita-Marathon mountain belt further to the southeast) "slowly disappeared under the assault of erosion and regional subsidence", but "endured as low hills….., sporadically submitting small amounts of fine-grained sediment to the west" (Fillmore, 2011). This sediment contributed to the material deposited during the Triassic and therefore to our two formations. Triassic sedimentationas a whole is "characterized by a variety of marine and non-marine deposits" (Stokes, 1986), suggesting the importance of marine transgression and regression and the lateral shifting of transitional marinedepositional environments during this period. A more significant and long lasting shift from a primarily marine to a more non-marine environment likely happened between the Early Triassic Period, when the Moenkopi Formation was deposited, and the Late Triassic, when the Chinle Formation was deposited.

"In a continuation of Late Permian geography, the west margin of Pangaea (being western North America) was initially an ocean, although a volcanic island [arc] lay an unknown distance offshore to the west" (Fillmore, 2011), the McCloud arc of Blakey (2014). The volcanic island arc and the North American plate finally fully collided in the Triassic to form the Sonoma highlands (or the Sonoma orogenic belt) to the west in central Nevada and a parallel foreland basin in the Four Corners area (Fillmore, 2011). The accreted arc terrane and foreland basin cut off a portion of the west-lying ocean, creating an interior basin occupied by an arm of the ocean opening to the northwest known as the "Moenkopi Sea" (Figure 27) and allowing for the Moenkopi Formation to accumulate (Fillmore, 2011). The climate at the time of Moenkopi deposition was likely still "arid" and prevailed "across the Four Corners region during this time" (Fillmore, 2011).

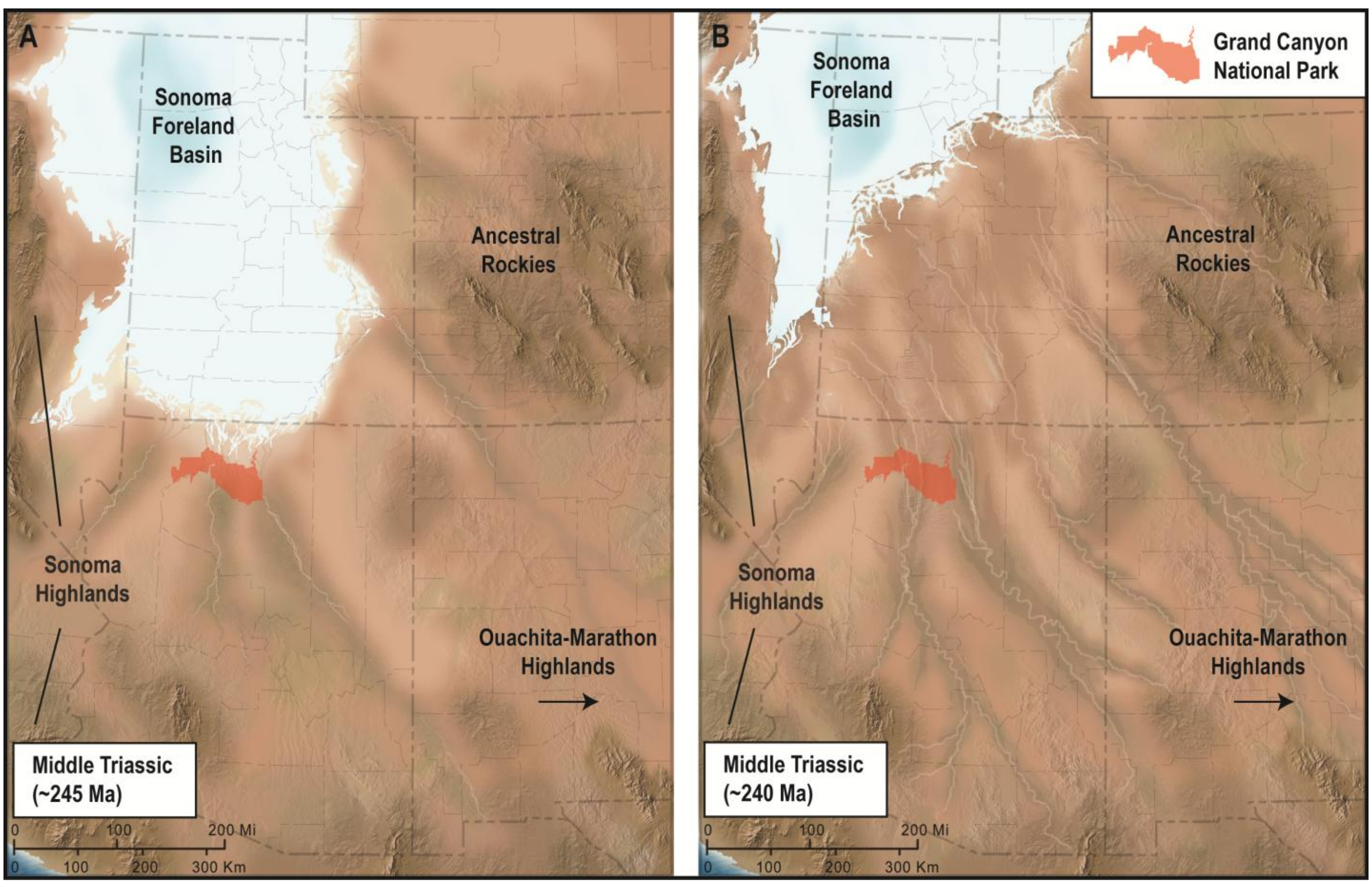


Figure 27. Middle Triassic paleogeography of western North America during early and later depositional members of the Moenkopi Formation; original maps are by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

The Moenkopi Sea was likely the remnant of a more extensive marine transgression which had flooded the entire passive margin of western North America and covered much of Utah and Arizona during the earlier Permian period, depositing the Kaibab Formation (Blakey and Middleton, 2012). In the Triassic, remnants of the Ancestral Rocky Mountains still lay the east of the interior basin, which since Late Paleozoic time had “endured and continued to shed sediment westward and influence drainage patterns” (Fillmore, 2000). Early Triassic Moenkopi deposition in areas nearer the Grand Canyon was overall the result of the interplay between marine and terrestrial paleoenvironments (Figure 27); the shallow Moenkopi Sea that dominated western Utah and northwestern Arizona and a coastal plain setting dominated by aridland braided rivers occupied eastern Utah and Arizona. Stokes (1986) describes the general paleoenvironment of the Moenkopi Formation in Utah as “mainly marginal marine mud flats”. The Moenkopi Sea likely did not extend into the area that is now the Grand Canyon, but rivers flowing into this sea (from the remnant Uncompaghre highlands to the east) probably did, depositing correlative sediment with a more terrestrial signature. In all likelihood, this depositional scenario played out over the course of Moenkopi accumulation, with earlier Grand Canyon area environments dominated more by marine conditions (Figure 27a), and later accumulation dominated by terrestrial fluvial conditions as the seas regressed from the region (Figure 27b).

The Sonoma mountain belt and volcanic arc that formed during Moenkopi time ultimately formed the western margin of the Late Triassic Chinle foreland basin in what would become Nevada and Arizona (Fillmore, 2000 and 2011). Trapped between the Sonoma highlands to the west and the ever diminishing Uncompaghre Uplift of the Ancestral Rockies to the east (as well as the Ouachita-Marathon mountains further to the southeast), the interior basin gradually filled with sediment. As millions of years passed, the interior basin was transformed into a terrestrial environment dominated by a sprawling, lowland river drainage system flowing generally to the northwest in the direction of the retreating Moenkopi Sea (Figure 28). This shift in paleo-environmental conditions resulted in the Chinle Formation’s deposition, which now overlies the Moenkopi (Blakey and Middleton, 2012). Fillmore (2000 and 2011) describes much of the deposition during the Late Triassic Chinle Formation as being “floodplain” and “river channel deposits”. Fossil plants and pollen preserved in the Chinle’s fluvial sediments record a dramatic shift in paleoclimate toward “wet subtropical” and “warm, strongly seasonal monsoonal” conditions influenced mainly by an equatorial paleolatitude of 5-15° N (Fillmore, 2011), while the volcanic arc surrounding the westernmost side of the Chinle basin would “episodically” erupt to leave evidence of its activity in the form of copious volcanic ash within some of the rock unit’s members.

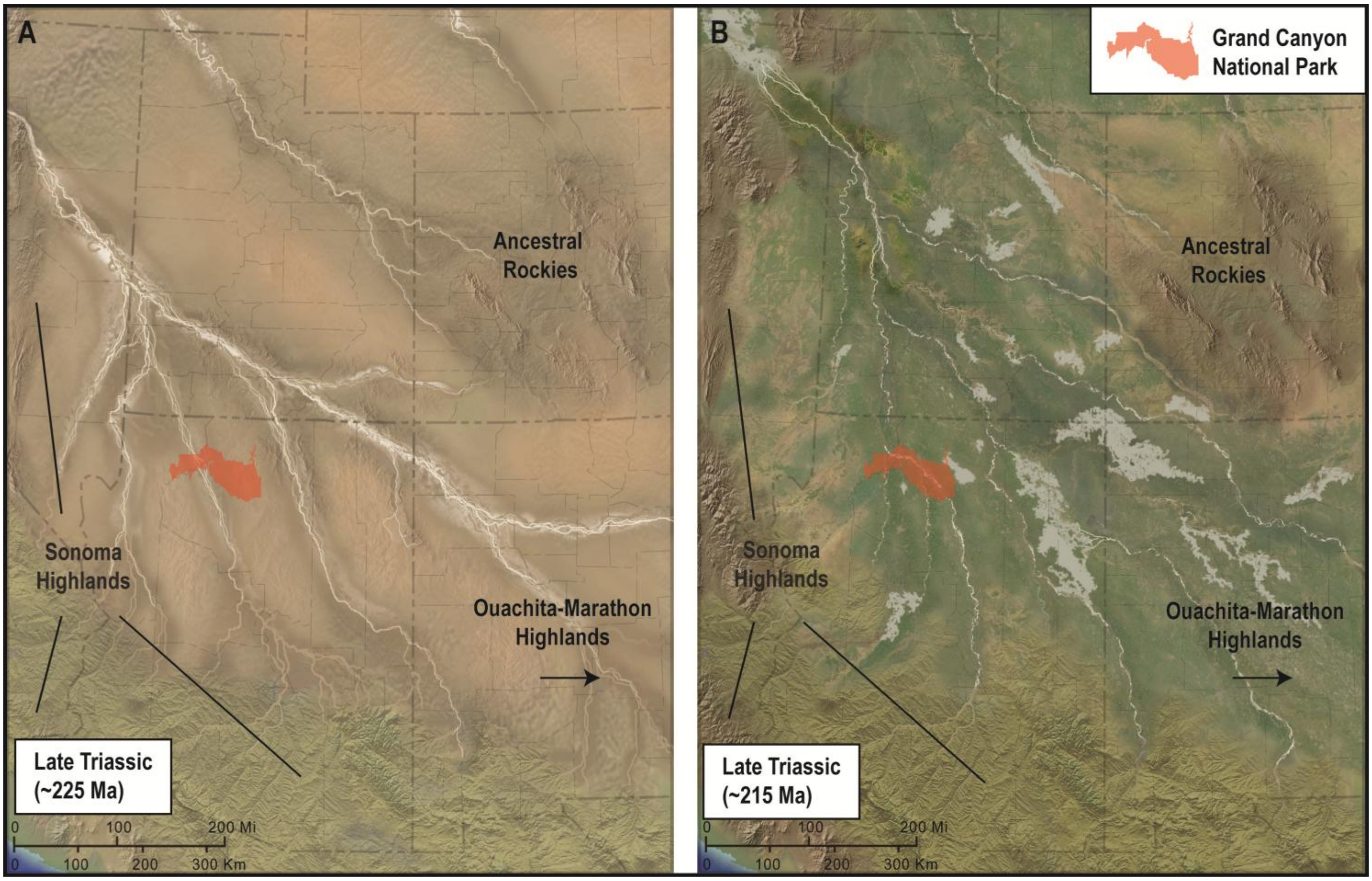


Figure 28. Late Triassic paleogeography of western North America during early and later depositional members of the Chinle Formation; original maps are by Ron Blakey and presented in Blakey and Ranney (2008) and Blakey and Middleton (2012).

Fillmore (2011) indicates that Chinle Formation deposition gradually shifted from a high energy, braided stream dominated system (Figure 28a) to low energy, "sluggish river and swampy lake deposits" (Figure 28b) as the western Sonoma volcanic arc, southeastern Ouachita-Marathon belt, and eastern Ancestral Rockies continued to supply volcanic material and sediment throughout the Late Triassic. By the end of the Triassic, the Uncompaghre and Ouachita-Marathon highlands had all but disappeared, although Sonoma highland sedimentation continued into the Early Jurassic. Evidence of prevailing paleoenvironmental conditions changes slightly depending on where one examines the Moenkopi and Chinle Formations, but the basic conclusion remains the same. During the Moenkopi Formation's deposition, the region would have been arid and largely covered by a west to east transitional setting from shallow sea to coastal plain (Figure 27), while during the Chinle Formation's deposition that sea regressed and the area became terrestrial, covered with riverine and swampy lake deposits dominated by a wet, subtropical climate (Figure 28).

If the Moenkopi and the Chinle formed in the same area as the modern Grand Canyon, then why are they not present in the canyon's walls? According to Edwin Colbert of the Museum of

Northern Arizona; “It seems obvious that rocks representing the three Mesozoic periods once were part of the Grand Canyon sequence…Their present absence from the canyon is merely the result of extensive erosion since Cretaceous time” (Colbert, 1974). As stated previously, although the Moenkopi and the Chinle Formations do not crop out in the Grand Canyon itself, they do appear extensively in areas just north and east of the canyon. The discussion which follows examines these two associated formations as they are described within the Escalante-Grand Staircase area of Utah, where they are widely distributed and well studied (Sprinkel, et al., 2003). The removal of the Moenkopi and Chinle Formations from a broad swath of the Grand Canyon region suggests area-wide uplift in part controlled by the Late Cretaceous-Early Tertiary Laramide Orogeny that built the modern Rocky Mountains to the east and by enhanced downcutting of drainage systems across the region caused by Late Tertiary and Quaternary Basin and Range extension and downdropping to the west.

The general descriptions of the Moenkopi Formation do not vary much between Colorado, Utah, or in other states where the rock unit occurs. Since the Moenkopi does not appear in the Grand Canyon, but its general descriptions remain largely the same wherever it is present, we can assume that the descriptions of the Moenkopi for the Colorado Plateau are largely what would be present if the formation appeared at the Grand Canyon. The Moenkopi on the Colorado Plateau is characterized as being deep red in color, fine grained, exhibiting many thin, horizontal lines, and has a tendency to weather into multiple steep slopes and benches (Fillmore, 2011). The Moenkopi is also “dominated by thin alterations of sandstone, mudstone, and shale, which contribute to its horizontally striped appearance” (Fillmore, 2011). These typical characteristics stand out for most of this formation’s members. Figure 29 provides an illustrative example of these features from the Vermillion Cliffs near Lee’s Ferry, AZ.

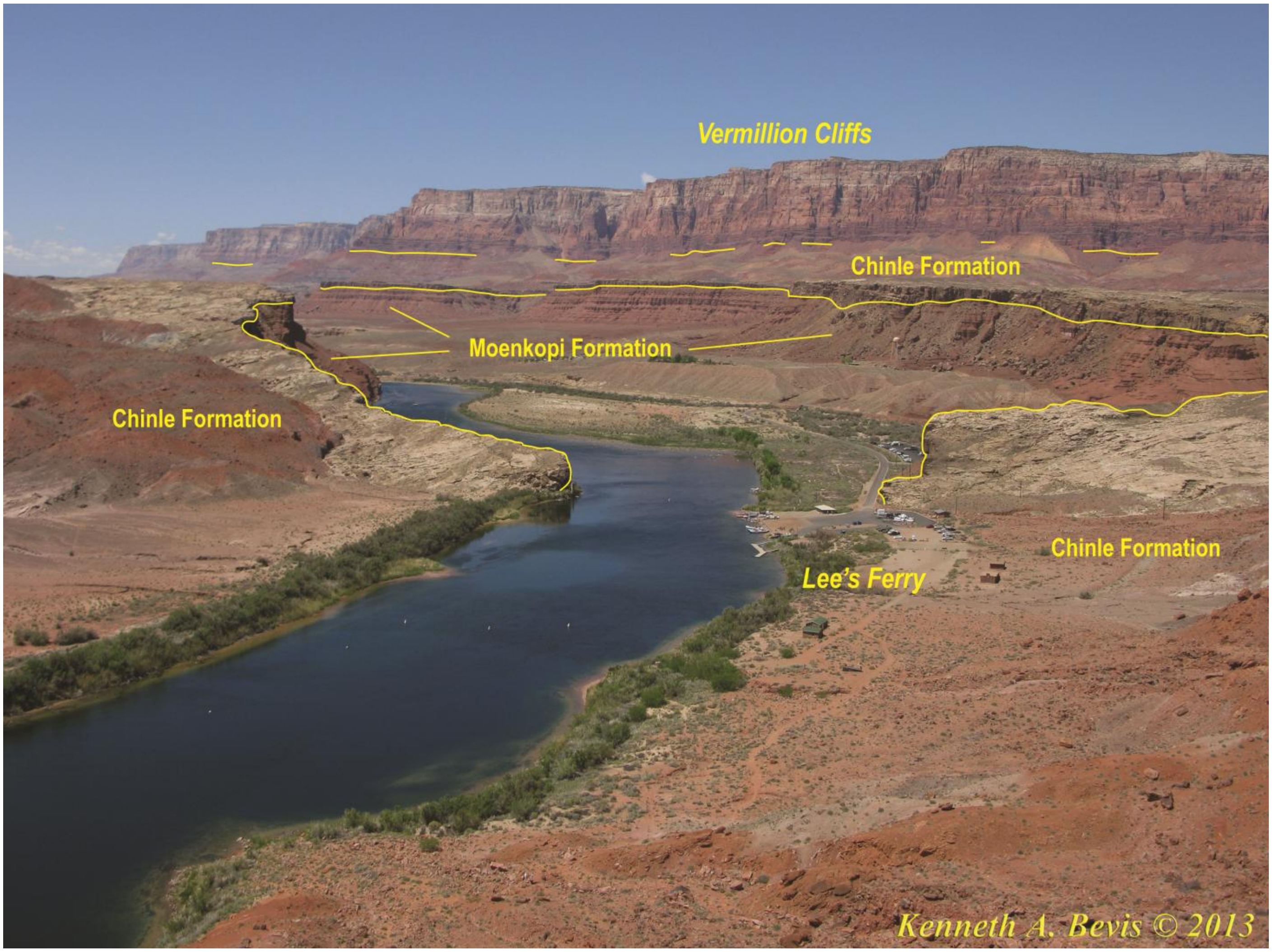


Figure 29. The Moenkopi Formation exposed in a monoclonal fold at the base of the Vermillion Cliffs near Lee's Ferry, AZ as viewed from the Spencer Trail.

The Moenkopi Formation can be divisible into 4-6 formal members, depending on where the formation occurs and what park or monument it is being described in (Sprinkel, et al., 2003). We focus discussion on the Moenkopi Formation's members as described within the Grand Staircase-Escalante National Monument just north of the Grand Canyon in Utah, covering the more well-known members of the formation, as we will do later with the Chinle Formation. The Moenkopi and the Chinle have paired stratigraphic relationships and it is best to describe their members in ascending order, so that one can better see the sequential facies changes within both formations (from Moenkopi through Chinle) through time, with the Moenkopi being deposited first and the overlying Chinle formation being deposited later.

The Moenkopi Formation is 910-1150 feet thick in the Grand Staircase section of the Grand Staircase-Escalante National Monument in Utah, but only 440-730 feet thick in the Escalante Canyons section to the northwest, indicating that the formation thickens westward (Sprinkel, et al., 2003). Fossils within the Moenkopi formation represent “a mix of terrestrial and marine taxa,” such as plants, brachiopods, fish, reptiles, amphibians, and reptile footprints (Sprinkel, et al., 2003). This mix of terrestrial and marine fossils within the members of the rock unit indicates that the Moenkopi Sea underwent multiple short-lived transgressions and regressions. The reason for these frequent changes in the level of the Moenkopi Sea likely had to do with the fact that the open marine water source of the Moenkopi was located to the northwest and deposition occurred on a “west-sloping” coastal plain. As geologist Robert Fillmore explains: “The very low relief of the coastal plain allowed the smallest sea level changes, or even tidal fluctuations, to shift the shoreline [of the Moenkopi Sea] tens of kilometers…” (Fillmore, 2011). As you will see, evidence of these fluctuations is present within the Moenkopi Formation’s members.

In the Grand Staircase section at the monument’s southwestern end, the Moenkopi is divided into six members. In ascending order, the members are the Timpoweap, the Lower Red, the Virgin Limestone, the Middle Red, the Shnabkaib, and the Upper Red. In the Escalante Canyons section to the northeast, however, the formation is divided into only four members which in ascending order include the Black Dragon, the Sinbad, the Torrey, and the Moody Canyon. From northeast to southwest across the monument, most of these members correlate to one another, although they have different names and sometimes different characteristics. In ascending order, the correlations are: the Black Dragon – no Grand Staircase Member equivalent; the Sinbad – correlates with the Timpoweap in the Grand Staircase; the Torrey – correlates with the Lower Red, Virgin Limestone, and Middle Red; and the Moody Canyon – correlates with the Shnabkaib and the Upper Red members in the Grand Staircase (Sprinkel, et al., 2003). Although the members of the Moenkopi formation may have different names in different places and even some differences in characteristics, the formation and its members remain largely the same no matter where the formation occurs.

The lowermost and oldest member of the Moenkopi Formation in Utah is the Black Dragon Member. The Black Dragon is about 40 feet thick, and consists typically of laminated to very thin-bedded siltstone and silty sandstone. This unit is ripple-marked and intercalated with thin-bedded, fine-grained, micaceous sandstone (Sprinkel, et al., 2003). The ripple marks and abundance of sand indicate that this member must have been formed in a transitional marine environment, likely nearshore environments of the Moenkopi Sea. The member shows no evidence that any major tectonic activity was occurring during the time of its deposition (Fillmore, 2011). The base of this member, which directly overlies Permian formations, consists of accumulations of chert or quartz pebble conglomerates, and some gypsum. Gypsum is an evaporate mineral, which is “precipitated from mineral-rich waters [within restricted basins] when evaporation rates exceed the influx of water” (Fillmore, 2000). The presence of gypsum suggests that, as previously stated, the Moenkopi Sea was largely cut off from its main source of water (the ocean, which lay to the west) and that the sea still largely dominated the Utah area. According to Sprinkel, et al. (2003), “it is thought that the [Black Dragon Member] was deposited on a mudflat or tidal flat and in associated lagoons,” probably on the eastern edge of the Moenkopi Sea.

Overlying the Black Dragon Member is the Sinbad (or Timpoweap) Member. As Fillmore (2011) describes, “The contact between the Black Dragon and the Sinbad is a gradual one, from red deltaic deposits upward into fossiliferous limestone. This points to a “progressive rise in sea level" associated with marine transgression. This transgression was one of the four main changes in sea level that the area experienced; however, the deposits within the Sinbad "mark the highest level that the sea reached during the Triassic period" (Fillmore, 2000). Figure 27a is representative of this member's paleogeography. This maximum transgression is the reason for the abundant limestone that is present within the Sinbad Member. Although Fillmore (2000) is describing the Sinbad member within the eastern Colorado Plateau region, his description does not differ from that of the member in the Grand Staircase-Escalante National Monument area. In the monument, the member is 55 feet thick in the Circle Cliffs area, and is also described as being "fossiliferous" with "limestone, dolomite, and calcareous siltstone" (Sprinkel, et al., 2003). It intertongues with the Black Dragon member below and the Torrey Member above, indicating a close stratigraphic relationship and proximity of adjacent transitional marine environments. The limestone beds of this member weather out as bench-forming ledges and are typically thin (Sprinkel, et al., 2003). The Timpoweap Member, which correlates with the Sinbad and is located at Buckskin Mountain within the monument, resembles its partner in color (light brown to yellow gray), but consists of several rock types: carbonate rocks, sandstone, chert breccias, and siltstone. What can be inferred about the two members' relationship is that both were deposited in a marine environment which "represented a transgression," but the difference between the two is that the thin, even beds of the Sinbad in the Circle Cliffs are an indication of "quiet water deposition," while the chert breccias in the Timpoweap "might indicate deposition in more turbulent waters" (Sprinkel, et al., 2003). Basically, the western side of the monument sat in deeper water, while the Timpoweap sat in shallower, more wave and/or tidally influenced waters on the eastern side.

The Lower Red Member sits on top of the Timpoweap Member in the Grand Staircase area of the monument. It is 140-220 feet thick and thickens westward along with the rest of the Moenkopi, and it consists of "red to chocolate-brown interbedded and thin-bedded siltstone and fine-grained sandstone" (Sprinkel, et al., 2003). Like most of the members within the Moenkopi, this member mainly forms slopes, but with alternating slight ledges. These ledges weather "platy" and display abundant ripple marks, indicating that we are still in a transitional marine environment. This is further verified by the interpretation of Sprinkel, et al. (2003) that: "The Lower Red Member was deposited on a tidal flat traversed by meandering streams." Yet, the presence of meandering streams suggests some sort of regression or prograding deltaic environment, being that in deeper waters (such as were present in earlier members) there were mostly lagoons and tidal flats. It appears that sea level in this area might have gotten shallower at this time. The member also correlates with the lower part of the Torrey Member in the Circle Cliffs area (Sprinkel, et al., 2003).

The Virgin Limestone Member overlies the Lower Red Member in the Grand Staircase area. It is 10-30 feet thick and thickens to the west. The member is described as being "conspicuous because it is ledge forming" (Sprinkel, et al., 2003). It consists of yellow-brown sandstone, siltstone, and limestone, which is consistent with all of the characteristics of the Torrey Member of the Circle Cliffs area to which it is correlative, though the descriptions may differ slightly. The member is not fossiliferous (Sprinkel, et al., 2003).

The Torrey Member overlies the Sinbad Member in the Circle Cliffs area of the monument. It is 240-310 feet thick and like other members in the area, it thickens northwesterly. "The upper contact with the Moody Canyon Member is intertonguing and sometimes difficult to place" (Sprinkel, et al., 2003). It has very fine to fine-grained sandstone and silty sandstone. These sandstones form thin to medium-bedded ledges, cliffs, and slopes, which are common throughout the Moenkopi formation. The "slope-forming constituents" are generally very micaceous, and the "ledge formers" display "ripple marks, load casts, drag structures and animal trails" (Sprinkel, et al., 2003). Features such as ripple marks and drag structures suggest flowing water, which could indicate stream transport. Sprinkel, et al., (2003) have inferred that: "The Torrey Member was deposited in a deltaic and shoreline environment," which would have produced these ripple marks and drag structures. This transition from marine waters of moderate depth in the Sinbad Member to delta and shoreline environments in the younger Torrey Member indicates regression and/or progradation of a terrestrial environment. As Fillmore states, "This change heralds the retreat of the shallow sea and the resurrection of the delta" (Fillmore, 2011). From a paleogeographic standpoint, the Moenkopi Sea at the time of the Torrey Member's deposition would have looked slightly shrunken, with some of the eastern Utah area that was once covered by water being exposed (Figure 27b).

The Middle Red Member of the Moenkopi that overlies the Virgin Limestone Member in the Grand Staircase "is the thickest of the Moenkopi members in this section and the least resistant to erosion" (Sprinkel, et al., 2003). It too has a slope-forming nature, and it is 280-400 feet thick. The member consists of "medium-brown to chocolate-brown mudstone and siltstone" along with fine-grained silty sandstone. Gypsum "veinlets" within this member indicate a correlation with the upper section of the Torrey Member which also contains gypsum. However, the gypsum content in this member "increases upward," which could indicate an increased evaporation rate and/or a decrease in the influx of water as the Moenkopi Sea regressed. Like most members of the Moenkopi, this member also displays ripple marks in the sandstone. The upper section of this member is believed to correlate with the lower section of the Moody Canyon Member in the Escalante Canyons area of the monument (the Moody Canyon correlates with the Shnabkaib Member in the Circle Cliffs, which sits just above the Middle Red Member). In an indirect way, this makes sense, because the Moody Canyon Member is resting on the Middle Red Member, and one is just fading into the other (Sprinkel, et al., 2003).

The Shnabkaib Member overlies the Middle Red Member in the Circle Cliffs area and is 150-250 feet thick. "It is a ledge- and slope-forming unit" that has silty gypsum and slopes that consist of sandstone and "red and green-gray siltstone" (Sprinkel, et al., 2003). In order for gypsum to form, we would need "calcium and sulfate ions" in order to "precipitate the gypsum." Therefore, the Shnabkaib Member was "probably deposited in restricted embayments of a sea surrounded by low tidal-flat and mud-flat areas" where a "fresh supply" of these needed ions could be deposited regularly (Sprinkel, et al., 2003). Essentially, the Shnabkaib Member was deposited in the receding Moenkopi Sea.

Finally, we arrive at the uppermost members of the Moenkopi within Grand Staircase-Escalante National Monument; the Upper Red Member and the Moody Canyon Member. Although these two members are thought to correlate, they are not exactly alike. The Upper Red Member in the Grand Staircase section is 90-180 feet thick, while the Moody Canyon Member in the Escalante Canyons is 200-300 feet thick. Both members are generally red-brown in color, and both can have interbedded siltstone and sandstone or mudstone. However, within the Moody Canyon

Member, “siltstone units occasionally display ripple marks, and mudstone is mostly laminated” (Fillmore, 2011). Both of these members are slope-forming, but the Upper Red Member is basically divided into two halves, with the upper half weathering into ledges and the lower half being the slope former. The depositional environment for both of these members is very similar: “The Upper Red Member is mainly a tidal flat deposit”, while the Moody Canyon Member is “believed to have been deposited in shallow quiet water, such as in ponds or lagoons on a tidal flat” (Sprinkel, et al., 2003).

From the beginning of the Moenkopi Formation onward, we have seen what appears to be a gradual marine regression to the northwest from the Utaharea punctuated by one or more minor transgressions, but no significant tectonic activity. However, it is sediments within the Chinle Formation, deposited after the Moenkopi Formation, that show evidence of a substantial transition from a marine environment to a terrestrially-dominated, as well as the considerable influence of tectonic activity.

The Chinle Formation is described in Colorado Plateau literature and in the Grand Staircase-Escalante National Monument as being lithologically “heterogeneous” and extremely colorful (Fillmore, 2011). The Chinle Formation overlies the Moenkopi Formation, however, “the contact between the Lower Triassic Moenkopi and the Upper Triassic Chinle Formation is an unconformity marked by the absence of Middle Triassic strata”, an erosional surface which “represents an omission of up to 10 million years” (Fillmore, 2011). The top of the Moenkopi contains distinctive northwesterly oriented furrows, presumably cut by stream erosion and into which the initial deposition of Chinle material occurred. This missing information in the rock record is the reason why it is difficult for scientists to determine if Moenkopi deposition continued into Middle Triassic time, because the geologic history needed to determine this simply isn’t there.

Fossils within the Chinle Formation include petrified wood (which is the most abundant), as well as fossils of fish, bivalves, amphibians, dinosaurs, and dinosaur tracks” (Sprinkel, et al., 2003); all indicative of a terrestrial setting. The rock unit’s sediments are inferred to be composed of “varying amounts of fluvial and lacustrine interbedded sandstone, mudstone, claystone, siltstone, limestone, gritstone, and conglomerate” (Sprinkel, et al., 2003). In the Grand Staircase area of Grand Staircase-Escalante National Monument, the rock unit is 500-930 feet thick and 425-750 feet thick in the Circle Cliffs area. It’s members in ascending order include; the Temple Mountain, the Shinarump, the Monitor Butte, the Petrified Forest, the Owl Rock, and the Church Rock; although some of these members are not present in every location within the monument (Sprinkel, et al., 2003). Figure 30 nicely displays exposures of the Chinle Formation at the mouth of the Paria River canyon near Lee’s Ferry, AZ.

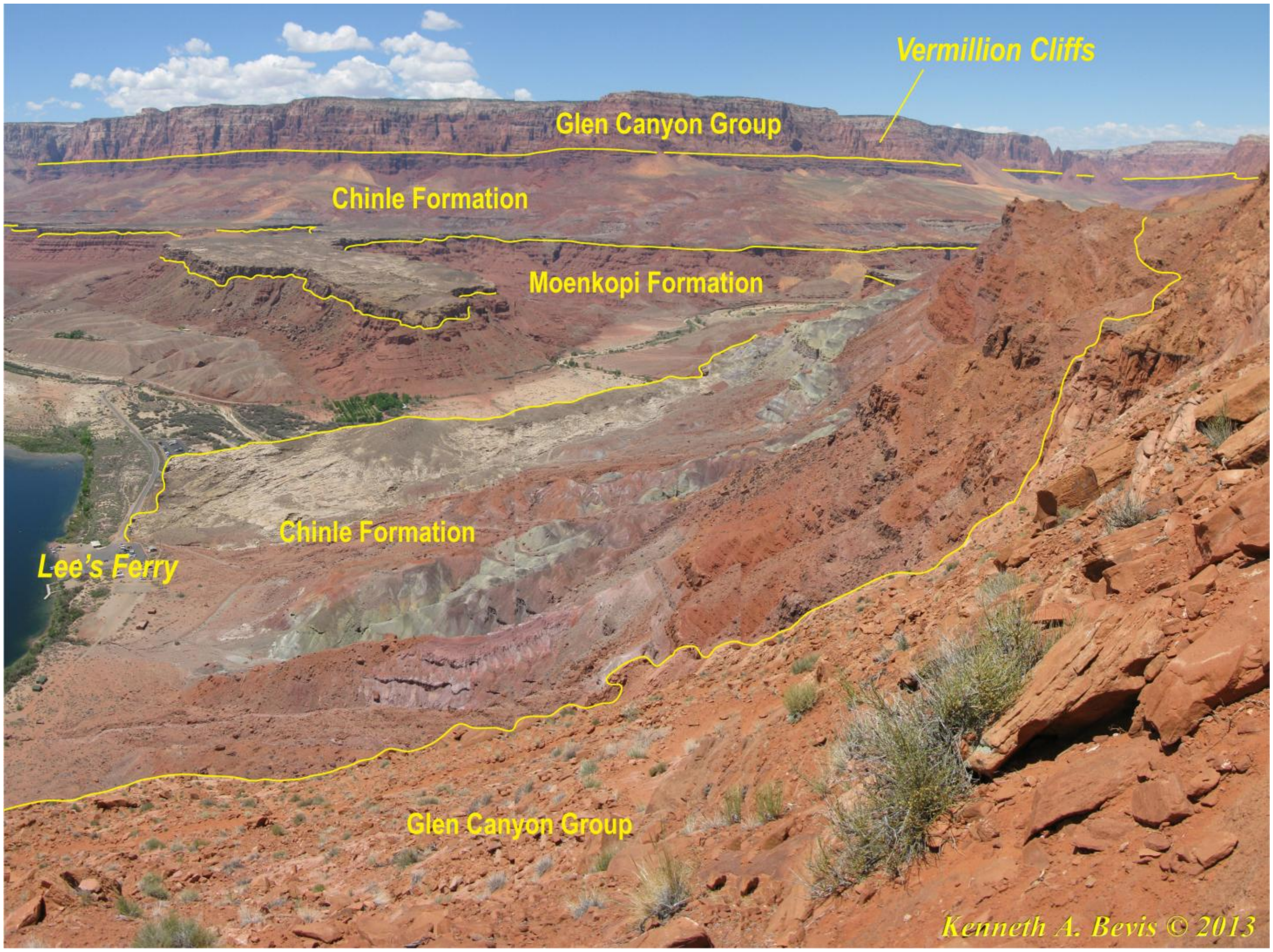


Figure 30. Exposures of the Chinle Formation at the mouth of the Paria River canyon near Lee's Ferry, AZ as viewed from the Spencer Trail.

The Temple Mountain Member only occurs within the Circle Cliffs area of the Grand Staircase-Escalante Monument, and on the Colorado Plateau, "the Temple Mountain Member…is mostly confined [well to the north] in the San Rafael Swell" (Fillmore, 2011). It is 50 feet thick, and "it forms slopes and ledges between the Moody Canyon Member of the Moenkopi Formation and the resistant cliff-forming sandstones of the Shinarump Member" (Sprinkel, et al., 2003). In the Utah-Colorado border region, the member's most distinctive characteristic is its "mottled" appearance, with "colors of purple, red, brown, and gray white" (Fillmore, 2011). The member consists of well-cemented siltstone containing immature quartz, along with some evidence of root casts and some considerable amount of dark minerals, such as tourmaline, which indicates that the source of the siltstone was not erosion of the Moenkopi. It is believed to be "mostly a fluvial deposit that was highly weathered before the Shinarump and later sediments were deposited" (Sprinkel, et al., 2003). These features, as well as the erosional disconformity between this and the underlying Moenkopi Formation, leads to an interpretation that at this point in time, this area had been transformed into a terrestrial environment. However, the actual deposition of the member itself was controlled by what Fillmore (2011) calls a "reinvigoration" of streams flowing through the area for a brief period of time during the Middle Triassic in response to "a drop in sea level and probable uplift southeast of the Colorado Plateau"; these rivers carved out "an early incarnation of the Chinle drainage network", the channels that would eventually receive Temple Mountain and Shinarump deposition.

The Shinarump Member primarily consists of relatively high energy stream-channel deposits: "Streams meandered northwestward [away from North American continental sourcelands] across an old Moenkopi surface where local thick soils [had] developed. The streams locally cut deep channels into this surface" (Sprinkel, et al., 2003). Unlike the Temple Mountain Member which is not widespread, "The Shinarump Member holds the basal position of the Chinle over much of the Colorado Plateau, particularly in northern Arizona and southern Utah" (Fillmore, 2011). Deposition at this time was dominantly fluvial, filling an internally drained basin (Figure 28a): "During Chinle time, southwest Utah may have been an area of interior drainage, with the Shinarump streams depositing their loads of suspended materials over a broad flat that had developed on the old Moenkopi surface" (Sprinkel, et al., 2003). Fillmore (2011) also notes that "Shinarump deposits mark the paths of energetic braided rivers." Throughout the Utah area, the Shinarump can largely be characterized as having trough cross-bedded sandstone, siltstone, and mudstone, as well as having fragments of petrified wood (as most of the members in the Chinle do) (Sprinkel, et al., 2003). However, the Shinarump also contains "volcanic-derived pebbles," indicating that the volcanic arc associated with the ongoing Sonoma Orogeny that developed during Moenkopi deposition continued through Shinarump time. As Fillmore (2011) describes, "…numerous tributaries from the remnant Ancestral Rockies and the chain of volcanic mountains to the west-southwest contributed significant amounts of sediment" to the braided rivers of the Shinarump.

The Monitor Butte Member overlies the Shinarump and underlies the Petrified Forest Member, forming gradual intertonguing contacts with its neighbors both above and below. It forms stepped-ledge slopes and because of its dominant gray coloration it is commonly included in the Shinarump. It contains gray-green mudstones with interbedded sandstones and siltstones, and it contains petrified wood fragments in some places and shales and coal bed deposits in others, probably related to the prevailing tropical climate and thickly forested conditions of the great paleovalley system developed throughout the time. "Generally, sandstone beds are more common in areas overlying channels of the Shinarump Member, indicating that stream courses established by Shinarump streams were maintained during Monitor Butte time" (Sprinkel, et al., 2003). It would appear that with this evidence there was not much change between the depositional environment of the Shinarump Member and that of the Monitor Butte Member. However, Fillmore (2011) describes evidence that proves otherwise: "The fine sandstone, mudstone, and shale of the Monitor Butte are a striking contrast with coarse underlying strata, reflecting an equally dramatic shift in depositional setting." The volcanoes that surrounded the Chinle basin to the west and the southwest were becoming more active, releasing "huge quantities of fine ash" into the air that were then carried by wind into the Chinle basin. As the ash accumulated, rivers became clogged and an overall lower energy system of "lakes, rivers, and wetlands evolved" (Fillmore, 2011). The abundant organic matter that accumulated in these long-lived lakes would later form bogs and eventually the coal that we find within the member

(Sprinkel, et al., 2003). The Moenkopi Sea at this point in time had completely disappeared from the Chinle's interior basin far to the northwest.

The Petrified Forest Member overlies the Monitor Butte. However, "whereas the lower three members of the Chinle generally form cliffs or stepped-ledge slopes, the upper three members generally form slopes, steep slopes, and higher in the section, ledgy slopes" (Sprinkel, et al., 2003). The most "conspicuous" characteristic of the Petrified Forest Member is its bright coloration, which is usually differentiated in bands. It is because of these magnificent bands of color that the Petrified Forest Member is the most eye-catching unit within the Chinle, and explains how areas with extensive exposure of the Petrified Forest may have gotten the name "The Land of the Sleeping Rainbow" (Sprinkel, et al., 2003). An interesting sedimentological feature of this member, along with all of the petrified wood that it contains, is the fact that it also contains bentonite. "Bentonite, or montmorillonite, is a type of clay produced from the decomposition (devitrification) of volcanic ash. It swells dramatically when wet" (Sprinkel, et al., 2003). Although this member is aesthetically pleasing, because it has "abundant swelling clays," it becomes extremely slick when moistened by rain or snow melt, and it is prone to slope failure, making it particularly dangerous as landslides occurring within this area of Utah have their roots in this member. Furthermore, roads traversing this member have a nasty reputation for turning into slick mud during the occasional summer storm. "The Petrified Forest Member was deposited on a fluvial plain largely as overbank deposits," indicating that streams formed in Shinarump time persisted through Petrified Forest time. "The plain was dotted with lakes and ponds" (just as it had been during deposition of the Monitor Butte Member), and at times, as the sediments accumulated, volcanic ash settled on the plain, which was altered to bentonite clay and locally into siliceous brown nodules" (Sprinkel, et al., 2003). Fillmore (2011) indicates that the continued presence of volcanic ash in the Petrified Forest Member was likely caused by the "continued flood of airborne volcanic ash into the basin from the erupting volcanoes to the southwest." This volcanic activity was sourced in the active volcanic arc to the southwest, likely the result of continuing subduction along the western margin of the North American continental plate. This member also contains abundant fossil wood, and huge logs (some 6 feet in diameter and 90 ft long) have been found (Sprinkel, et al., 2003). These huge logs (as well as coal deposits) suggest that during the time of Petrified Forest deposition, climate in the Chinle basin area still resembled monsoon-like, tropical conditions, allowing for abundant vegetation to grow (Fillmore, 2000).

The Owl Rock Member overlies the Petrified Forest Member, and "it crops out in the steep slope interval formed by the three upper units of the Chinle Formation" (Sprinkel, et al., 2003). It is 150-250 ft thick, and because bentonite is present in this member, though in a lesser amount than in the Petrified Forest Member, some investigators believe that the upper part of the Petrified Forest and the lower part of the Owl Rock form a gradational contact. The lesser bentonite suggests "that volcanic activity in the southwest was waning" (Fillmore, 2011), although there were still signs of some tectonic activity: "Regionally, the Owl Rock Member reflects continued subsidence of the Chinle basin, centered around the Four Corners area, but with a much reduced sediment supply from surrounding highlands." The paleogeography during Owl Rock deposition is depicted by Figure 28b, and can basically be compared to that of the present-day Everglades, "in which freshwater trickled northwestward to an eventual rendezvous with the sea in central Nevada" (Fillmore, 2011). This member is "composed of thin lenticular beds of green limestone interbedded with red, brown, and green-gray sandstone and mudstone." In the Grand Staircase-Escalante National Monument, it was believed to be deposited "on a fluvial environment on a plain dotted with lakes" (Sprinkel, et al., 2003), while more broadly, on the greater Colorado Plateau, it was considered to be "deposited in a widespread lake and marsh system linked by the occasional wandering river" (Fillmore, 2011).

The uppermost member of the Chinle is the Church Rock member. In both Colorado and Utah, the member is described as having "orange red sandstone and siltstone" (Fillmore, 2011). He further notes that, "Based on similar lithologies and stratigraphic position, the Church Rock Member has been correlated with the widely scattered Rock Point Member in northeast Arizona and western New Mexico." This shows that the member itself is widespread, and suggests a broad geographic distribution of its depositional environment. In the Colorado Plateau area, "the Church Rock and its equivalents were deposited in a variety of settings, all signaling an increasingly arid climate" (Fillmore, 2011). However, in the Grand Staircase-Escalante National Monument area, "The sediments were deposited in lakes and by streams on an alluvial plain that [still] sloped away from the Uncompahgre uplift in eastern Utah during Late Triassic time" (Sprinkel, et al., 2003). This transition from drier conditions in Colorado to wetter conditions in Utah suggests that fluvial drainage associated with deposition of this member was chiefly derived from remnant highlands along the west coast. There are no volcanics within this member, or evidence of tectonic activity, indicating that the area was becoming stable once more.

At the beginning of the Triassic period, Pangaea was undergoing break up, and this tectonic activity had an indirect impact on the depositional environment of the Moenkopi Formation. To the west of the Pangaean supercontinent, a volcanic island arc made its final landfall with North America. This collision and subduction created the Sonoma Orogeny in central Nevada and its accompanying Sonoma highlands, followed by active arc volcanism along the North American southwest coast. Consequently, a foreland basin developed between the Sonoma highlands and the much subdued Ancestral Rocky Mountains, leaving an interior basin occupied by the Moenkopi Sea over most of Utah. Tectonic activity generally remained mild throughout the time of the Moenkopi's deposition. By the time of the Chinle Formation's deposition, the Moenkopi Sea had almost entirely receded back into the ocean on the west coast of the North American continent, leaving behind a northwestward draining paleoriver valley system in the basin between the Sonoma highlands and Ancestral Rockies. Tectonic activity ramped up during Chinle Formation deposition, with evidence of further subduction and growth of a volcanic arc to the southwest represented by copious volcanic ash within some of the Chinle's later members.

Based on lithological and paleontological evidence, the Moenkopi Formation is inferred to have been deposited in moderately deep to shallow marine and transitional marine environments, and as sea level steadily regressed, they were gradually replaced by terrestrial environments upward within the formation. Between the Moenkopi and Chinle Formations there is an erosional unconformity, indicating that marine conditions completely disappeared for a time, allowing substantial subaerial exposure, weathering, and erosion of the top of the Moenkopi. Dominated by fluvial deposits that contain abundant plant and animal fossils, the Chinle Formation is believed to have formed in a wetter climate than those that preceded or followed its deposition. Starting with the deposition of the Shinarump Member of the Chinle Formation, accumulation of sediment was renewed as streams and rivers began to flow across the previously deposited Moenkopi Formation. In the Monitor Butte Formation, lakes and ponds began to dominate the interior basin as volcanic ash from the west clogged earlier formed rivers, creating abundant vegetation and adding to the water supply in the area. From the Petrified Forest Member through the Owl Rock Member, previously established drainage in this area prevailed, although volcanic activity was becoming significant. Finally, in the Church Rock Member, volcanic activity waned, and the environment began to transition into drier terrestrial conditions, with streams still present in western areas of the basin.