National Park Service
U.S. Department of the Interior

Natural Resource Stewardship and Science

Paleontological Resource Inventory and Monitoring

Northern Colorado Plateau Network

Natural Resource Technical Report NPS/NCPN/NRTR—2012/585



ON THE COVER
Images clockwise from top left: a trionychid turtle from Fossil Lake, the ancient lacustrine system that includes Fossil Butte National Monument (FOBU); foliage of the distinctive Chinle Formation plant *Sanmiguelia* from Arches National Park (ARCH); Mississippian-age coral from Timpanogos Cave National Monument (TICA); an artiodactyl track from Pleistocene sediments associated with Coalpits Lake in Zion National Park (ZION).

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NOTE:
This report provides baseline paleontological resource data to National Park Service administration and resource management staff. The report contains information regarding the location of non-renewable paleontological resources within NPS units.

It is *not* intended for distribution to the general public.

May 2012

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Abstract

Paleontological resources (fossils) are any remains of past life preserved in a geologic context. Paleontological resources are non-renewable resources found in at least 234 National Park System units. Despite the abundance and diversity of these resources throughout the National Park Service (NPS), until recently few parks have had adequate baseline paleontological resource data.

The NPS National Inventory and Monitoring Program and the NPS Geologic Resources Division have developed various strategies to provide parks with the baseline paleontological resource data necessary for greater understanding and appropriate management of these resources. This report presents a paleontological resource summary for the parks of the Northern Colorado Plateau Network (NCPN). The summary was compiled through extensive literature reviews and interviews with park staff and professional geologists and paleontologists. Preliminary paleontological resource management recommendations are also included for each park. Summaries have now been prepared for all of the 32 NPS Inventory and Monitoring Networks.

The parks of the NCPN contain some of the most significant fossil-bearing strata in the National Park System. Research in these areas began in the vicinity of Fossil Butte National Monument (FOBU) during the late 19th century and spread into Utah (Dinosaur National Monument [DINO]) and Colorado (just outside of Colorado National Monument [COLM]) in the first decade of the 20th century. Most fossil occurrences in the NCPN are from the Mesozoic (251.0 to 65.5 Ma [million years ago]) (see the appendix for a geological timescale). Mesozoic NCPN fossils range from plants to shelled invertebrates to vertebrate bones to tetrapod tracks. Many NCPN units also have Quaternary fossils (younger than 2.6 Ma), a handful have Paleozoic marine fossils (542 to 251 Ma), and one, FOBU, is part of one of the world's best Eocene fossil deposits (55.8 to 33.9 Ma). Fossils from NCPN parks are found in museums throughout the world, with fossil fish from FOBU and its vicinity as particularly common items. The parks and monuments of the NCPN also have connections between their paleontological resources and cultural resources.

Two NCPN units, DINO and FOBU, are among the most significant fossil sites in North America. DINO's Dinosaur Quarry and microvertebrate sites are some of the most productive sites in the world-famous Morrison Formation, and FOBU includes some of the most productive fossil sites of the Green River Formation, one of the best-known fossil lake deposits in the world. Until relatively recently, these two monuments completely overshadowed the fossil resources of the other NCPN units, but recent surveys have established the rest of the NCPN units as untapped sources of significant paleontological wealth. Tetrapod tracks from the Triassic and Jurassic are particularly abundant, and are known from Arches National Park (ARCH), Canyonlands National Park (CANY), Capitol Reef National Park (CARE), COLM, DINO, Pipe Spring National Monument (PISP), Timpanogos Cave National Monument (TICA), and Zion National Park (ZION). Other common fossil types include marine invertebrates from the Paleozoic, found at ARCH, CANY, CARE, DINO, Golden Spike National Historic Site (GOSP), TICA, and ZION; Cretaceous marine invertebrates, found at ARCH, Black Canyon of the Gunnison National Park (BLCA), Bryce Canyon National Park (BRCA), CARE, Curecanti National Recreation Area (CURE), DINO, and possibly Hovenweep National Monument (HOVE) and ZION; and Quaternary-age packrat middens, found at ARCH, CANY, CARE, COLM, DINO, GOSP, NABR, and possibly CURE. The fossil record of the NCP parks and monuments is strongest from the middle of the Pennsylvanian (about 310 Ma) to the middle of the Late Cretaceous (up to about 80 Ma). DINO and TICA have fossils from older periods of the Paleozoic. BRCA, Cedar Breaks National Monument (CEBR), and FOBU have fossils from the Eocene. Finally, the majority of NCPN units have some kind of fossils from the Quaternary, usually packrat middens or other paleoecological materials. Together, they provide a detailed picture of the time between the assembly of the supercontinent Pangaea to the early stages of the formation of the Rocky Mountains, with shorter glimpses of earlier deposition, an early Cenozoic episode of lake formation and desiccation, and the past few tens of thousands to hundreds of thousands of years of climate change and ice ages.

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Dedication



This report is dedicated to our colleagues from the Utah Geological Survey [from left to right] Jim Kirkland, Grant Willis, Scott Madsen and Don DeBlieux.

Introduction

Paleontological resources, or fossils, are any remains of past life preserved in geologic context. There are two main types of fossils: body fossils and trace fossils. Body fossils are the physical remains of an actual organism (shells, bones, teeth, plant leaves, etc.) while trace fossils (burrows, coprolites, footprints, trackways, etc.) preserve evidence of an organism's activity or behavior. Fossils are non-renewable natural resources that possess scientific, educational, and interpretive value. In the National Park Service (NPS) there are three fossil context categories: in situ (fossils found in rock layers), in museum collections, or in cultural resource contexts (building stones, archeological resources, historic documents, and ethnographic legends or stories). This report addresses fossil discoveries in all of these contexts.

The establishment of baseline paleontological resource data is essential for the appropriate management of fossils found within National Park System units. Although at least 234 units have been identified with paleontological resources, few parks have adequate baseline paleontological resource data.

In conjunction with the NPS Geologic Resources Division and the NPS Inventory and Monitoring (I&M) Program, paleontological resource inventories have been initiated in dozens of parks Servicewide and are further described in the Methodology and Inventory Strategies section of this report. This report presents paleontological resource inventory and monitoring data compiled for the parks of the Northern Colorado Plateau I&M Network (NCPN).

The Northern Colorado Plateau Network includes sixteen NPS areas in Arizona, Colorado, Utah, and Wyoming as shown in Figure 1: Pipe Spring National Monument (PISP) is within Arizona; Black Canyon of the Gunnison National Park (BLCA), Curecanti National Recreation Area (CURE), and Colorado National Monument (COLM) are entirely within Colorado; Dinosaur National Monument (DINO) and Hovenweep National Monument (HOVE) include land in both Colorado and Utah; Arches National Park (ARCH), Bryce Canyon National Park (BRCA), Canyonlands National Park (CANY), Capitol Reef National Park (CARE), Cedar Breaks National Monument (CEBR), Golden Spike National Historic Site (GOSP), Natural Bridges National Monument (NABR), Timpanogos Cave National Monument (TICA), and Zion National Park (ZION) are within Utah, and Fossil Butte National Monument (FOBU) is within Wyoming.

The NCPN parks detailed in this report range in size from 16 ha (40 acres) (PISP) to 136,620.99 ha (337,597.83 acres (CANY). The parks and monuments of the NCPN are renowned for their natural scenic beauty, with ARCH, BLCA, BRCA, CANY, CARE, CEBR, NABR, and ZION all originally established to preserve areas with outstanding features. DINO and FOBU were established to protect areas of notable fossil resources. GOSP, HOVE, and PISP were established to protect cultural sites. The NCPN also includes a cave monument (TICA) and a recreation area (CURE). All of the units have some form of paleontological resources, and most of them are extensively fossiliferous.

NCPN Geological Background

The geologic history of the NCPN (discussed in greater detail in the individual park summaries) is best understood from rocks that formed between about 310 Ma to about 50 Ma (million years ago). Older rocks are present in some areas but are generally deeply buried by thick Upper Paleozoic–Mesozoic sequences, while younger rocks are uncommon because of the uplift of the Colorado Plateau, which has led to widespread erosion (see the appendix for a geological timescale). The most complete geological record is present at DINO, while the oldest rocks are present at BLCA and CURE. Most of the other NCPN units contain extensive exposures of rocks from some part of the above-mentioned 310–50 Ma interval, with the exception of GOSP and TICA, which have older Paleozoic bedrock. Deposition in the NCPN units usually occurred just above or just below sea level, as long cycles of shallow continental seas transgressed over the land or regressed (retreated) from it. This situation only changed relatively recently in geological time, when the Colorado Plateau began to undergo uplift near the close of the Cretaceous, starting about 70 Ma.

The oldest rocks exposed in an NCPN unit are metamorphosed sedimentary rocks found at BLCA and CURE, which predate 1,700 Ma. These rocks represent a time of continental assembly. The paleogeography featured a series of island arcs offshore of the growing proto-North American craton, and has been compared to the modern Banda Sea of Indonesia. Around 1,700 Ma, some of these island arcs collided with the continent. These fragments were then altered by multiple subsequent episodes of accretions and intruding igneous rocks. The BLCA and CURE rocks are separated from the next-oldest NCPN rocks, at DINO, by nearly 1 billion years. Neoproterozoic deposition at DINO dates to approximately 770 Ma, represented by the Uinta Mountain Group. At this time, the NCPN area had been part of North America for hundreds of millions of years, and what would become a familiar pattern of shallow marine transgression and regression had begun. North America was located near the equator as part of the ancient supercontinent Rodinia, with the future Antarctic and Australian landmasses near what is now western North America. The supercontinent

broke up a few tens of millions of years later, between 720 and 670 Ma. The Uinta Mountain Group is thought to be related to the similar Chuar Group of the Grand Canyon and Pahrump Group of Death Valley (the ChUMP hypothesis). Another isolated window of Precambrian deposition can be seen at TICA, and dates to about 600 Ma.

Between about 540 and 310 Ma, rock exposures in the parks and monuments of the NCPN are largely limited to DINO and TICA. Cambrian deposition at these two units is represented by shallow marine formations populated by brachiopods (lamp shells), trilobites, and unknown organisms known only from their burrows and trails. The following Ordovician, Silurian, and Devonian periods are poorly represented in the parks and monuments of the NCPN. Some of this poor record is due to deep burial of the relevant rocks. At DINO and TICA, where older and younger rocks are present, the absent interval is probably the result of an episode of uplift during the Devonian that led to the erosion of the formations deposited during the missing periods. This episode of uplift is known as the Antler Orogeny (mountain-building event), the result of an island arc colliding with western North America. The orogeny caused highlands to rise from Nevada to southern Canada, and their subsequent erosion to the east contributed to the NCPN formations of the later Paleozoic. Only TICA has any rocks that are definitely from the Ordovician to the Devonian, and then only from the upper Devonian. The record becomes considerably stronger in the Mississippian, with both DINO and TICA having extensive outcrops of Mississippian-age rocks. From the sedimentology, it is apparent that multiple large-scale transgression-regression cycles of a shallow warm sea occurred during the Late Devonian and Mississippian. These seas were home to an abundance of lifeforms, including algae, foraminifera (single-celled animals that secrete a hard covering), sponges, colonial and solitary corals, bryozoans (moss animals), brachiopods, bivalves, cephalopods (nautilus-, squid-, and octopus-like animals), gastropods (snails), trilobites, ostracodes (tiny shell-dwelling crustaceans), crinoids (sea lilies), blastoids (extinct crinoid-like echinoderms), echinoids (sea urchins), holothurians (sea cucumbers), conodonts (extinct eel-like chordates), and fish. Early land plants such as *Lepidodendron*, lycopsids, and horsetails lived on the shores.

Rocks from the early part of the Pennsylvanian can be seen at DINO and GOSP. Like the Mississippian rocks, they are shallow marine in origin and are extensively populated by invertebrates. To the east, in what is now western Colorado, tectonic activity was beginning. This episode of tectonism created several major structural features that had important implications for the history of the NCPN to the present. Large areas of Colorado were uplifted as part of the Ancestral Rocky Mountains, which included several sub-ranges that trended northwest-southeast. The Uncompahgre Uplift is the subsection with the most direct influence on the NCPN, occupying a sizeable part of the eastern section of the network. During the late Paleozoic and early Mesozoic, it would have formed a natural divide between BLCA, COLM, CURE, DINO, and FOBU from ARCH, CANY, CARE, and NABR. A structural depression called the Paradox Basin formed off of the southwest flank of the Uncompahgre Uplift; ARCH, CANY, and NABR are among the areas underlain by this basin. Another basin formed north of the uplift. The erosion of the uplift contributed to the thick late Paleozoic–early Mesozoic sedimentary sequences of the basins. In the greater context of North America, the Pennsylvanian and following Permian include the final stages of the assembly of the supercontinent Pangaea. Unlike Rodinia, there were no continental landmasses west of the NCPN, and it is unclear why the Ancestral Rocky Mountains formed where they did.

North of the Uncompahgre Uplift, in DINO, the Pennsylvanian closed with a marine regression followed by a dune field. South of the uplift, the Paradox Basin was beginning the cycles of marine deposition for which it is now famous. Cyclic alternation of different sedimentary rocks in the basin is probably tied to global sea level changes from polar glaciation. The best-studied are the salt cycles, produced by periodic drops in sea level that left the water in the relatively confined basin to evaporate in the near-equatorial setting. Thick deposits of salt are present at depth in much of the Paradox Basin, and have continued to affect the geology up to the present; for example, Salt Valley of ARCH is the result of partial dissolution of an underlying salt wall, and the Needles District of CANY owes its topography to its underlying salt. During those times when the Paradox Basin wasn't undergoing evaporation and mineral precipitation, it hosted diverse marine invertebrates. ARCH and CANY include outcrops of the Paradox Basin's marine history.

Permian deposition can be found in many of the NCPN units. North of the Uncompahgre Uplift, at DINO, it occurred under fluctuating coastal conditions. South of the uplift, at ARCH, CANY, CARE, and NABR, terrestrial deposition predominated, forming the Cutler Group. Here, eolian (wind-deposited) sand dunes emerged east of a shallow seaway, which occasionally transgressed over them. At CARE, and in the southwest part of the NCPN at ZION, outcrops from the seaway can be observed. Most of the terrestrial formations represent hostile environments and are poorly fossiliferous. The most abundant fossils are tetrapod tracks, with some plants and tetrapod body fossils also known. The end of the Permian is poorly known in the NCPN.

The Early Triassic is widely represented in the NCPN by the Moenkopi Formation and its equivalents. The basic geography of the NCPN during this time featured a broad coastal plain on the south and east, and a shallow continental sea on the north and west. Several transgressions and regressions led to interbedded sequences of marine, continental, and evaporitic rocks. The Moenkopi Formation includes all three types of deposition, although there is some interest in limiting it to the coastal plain rocks. It can be viewed in ARCH, CANY, CARE, DINO, NABR, PISP, and ZION, and the equivalent marine Thaynes Limestone is present at FOBU. Historically, fossils are

not well-known from the Moenkopi Formation, except in the marine beds. Recent work, including at ARCH and CARE, has uncovered many tetrapod tracks, expanding our knowledge of the terrestrial forms.

Following the Moenkopi Formation is the Chinle Formation, one of the best-known Triassic formations in the world. Excellent exposures of the Chinle Formation can be found in several NCPN parks and monuments, such as ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. The Chinle Formation is fully terrestrial, representing a setting of rivers, floodplains, swamps, lakes, and conifer forests. Mountains and volcanoes rose in the distance to the south and west. The swamps and lowlands were dominated by large salamander-like amphibians and crocodile-like phytosaurs, feeding on diverse fish communities. In the uplands were armored herbivorous aetosaurs, cow-sized "mammal-like reptiles," terrestrial bipedal reptiles related to the ancestors of crocodilians, and early predatory dinosaurs. The climate is thought to have been seasonal, and substantial quantities of volcanic ash were deposited on the floodplains and later altered into clay. The lands of the NCPN were still near the equator during the Late Triassic. After having moderated from the drier conditions preceding the Triassic, the climate again gradually became drier over time as Pangaea drifted north and the NCPN area moved from equatorial to arid-zone latitudes. Near the end of the Triassic, the dissolution of Pangaea was under way.

As the Jurassic began, approximately 200 Ma, the drying trend continued and a vast erg (a "sea" of sand dunes) formed, with sand trapped in an interior basin. The erg was briefly interrupted by a more humid period featuring a broad floodplain. These uppermost Triassic and Lower Jurassic rocks, dotted by dinosaur tracks and other trace fossils, are known collectively as the Glen Canyon Group. They are well-exposed through much of the NCPN, including at ARCH, CANY, CARE, COLM, DINO, PISP, and ZION. A marine transgression near the end of the Early Jurassic signaled the onset of another series of transgression-regression cycles, the result of a fluctuating sea on the north and west of the NCPN. Collectively, the marine and continental rocks are known as the San Rafael Group, and can be found at ARCH, BLCA, CANY, CARE, COLM, CURE, DINO, and ZION. These rocks record a complex series of changing environments; although many of the same formations are found in the parks and monuments, they often represent distinct environments. The terrestrial formations include ergs, tidal flats, and sabkhas (salt flats). They are typically sparsely fossiliferous, although dinosaur tracks are abundant in some places, such as the ARCH area. The marine formations have much more abundant fossils. Unsurprisingly, CARE, DINO, and ZION, which have the greatest marine influence for their Middle Jurassic rocks, also have the most extensive fossil records. Many of the forms that dominated the Paleozoic seas went extinct at the end of the Permian, so the Middle Jurassic seas had significantly different faunas, dominated by mollusks such as bivalves, gastropods, and various types of cephalopods. Crinoids were also still abundant, and fully marine tetrapods such as dolphin-like ichthyosaurs, long-necked plesiosaurs, and large-skulled pliosaurs made up the highest trophic level. During this time, North America continued moving north, nearing the end of the arid-zone latitudes; southern Utah reached 22° N during this time.

The seas retreated during the early part of the Late Jurassic, leaving behind extensive floodplains. The more humid conditions supported a great diversity of life and greater biomass in the last sedimentary formation of the Jurassic in the NCPN, the well-known Morrison Formation. The Morrison Formation is interpreted as a semi-arid seasonal savanna-like setting. During deposition of the upper portion of the formation, a vast basin is thought to have extended from near Grand Junction, Colorado, to Albuquerque, New Mexico. This basin was periodically occupied by a shallow saline lake which has been named Lake T'oo'dichi'; the setting is described as analogous to the modern Lake Eyre basin in Australia. The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such well-known forms as *Allosaurus*, *Apatosaurus* (=*Brontosaurus*), *Brachiosaurus*, *Camarasaurus*, *Camptosaurus*, *Ceratosaurus*, *Diplodocus*, and *Stegosaurus* populated the floodplains of the Morrison Formation. It also has a substantial plant fossil record, abundant freshwater bivalves and gastropods, and diverse turtles, small reptiles, crocodile relatives, and early mammals. In the NCPN, the Morrison Formation is exposed at ARCH, BLCA, CARE, COLM, CURE, DINO, and HOVE, and is fossiliferous in most of these parks and monuments. DINO includes some of the most productive fossil sites in the entire formation, COLM is directly adjacent to historically-important fossil localities, and ARCH, CARE, and CURE also include significant discoveries. In the western NCPN and farther west, a mountain-building event known as the Sevier Orogeny was in progress by the end of the Jurassic, forming an area of highlands. It continued through the succeeding Cretaceous Period and into the Eocene. The new mountains depressed the crust to the east, forming a basin in much of the NCPN and providing much of the sediment that filled it.

The Early Cretaceous of the NCPN was similar to the Late Jurassic in terms of paleontology, paleoecology, and sedimentology: a tetrapod assemblage dominated by abundant and diverse large dinosaurs, although representing some different groups, roamed a seasonal semi-arid setting of rivers, floodplains, and lakes. ARCH, CARE, COLM, DINO, and HOVE include rocks from this interval. The paleontology of these rocks has become an important subject of investigation since the mid-1990s; major finds have been made at DINO, in and around ARCH, and near CARE. Near the end of the Early Cretaceous, a shallow continental sea known as the Western Interior Seaway or Cretaceous Interior Seaway began to encroach on the NCPN from the east. Its approach is often signaled by the deposition of the Dakota Formation, a widespread rock unit that exhibits vertical transition from

terrestrial fluvial deposition at the base to coastal marine deposition at the top. Because the Western Interior Seaway arrived at different times in different places, the time of the transition varies within a few million years of 100 Ma, becoming younger going east to west until reaching western Utah; CEBR and ZION are near the maximum western limit. The Western Interior Seaway is one of the best-known former continental seaways in the world. Its rocks are famous for their fossils: abundant microscopic organisms, bivalves, ammonites, sharks, bony fish, giant sea turtles, diverse large marine reptiles, pterosaurs (flying reptiles), and early toothed birds. On the surrounding coasts, early flowering plants and conifers supported a vertebrate fauna of turtles, small reptiles, crocodile relatives, diverse small theropods, tyrannosaurs, armored dinosaurs, horned dinosaurs, hadrosaurs (duckbilled dinosaurs), and early mammals. The Western Interior Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America. It underwent several large-scale and numerous small-scale transgressions and regressions. Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the NCPN. At the same time the Seaway was present, volcanoes to the west introduced large amount of ash, which over time have altered to bentonitic clays. ARCH, BLCA, BRCA, CARE, CEBR, COLM, CURE, and DINO have extensive exposures of Western Interior Seaway rocks and their continental equivalents. The marine rocks are historically the most fossiliferous, but screenwashing of terrestrial rocks has recently revealed rich terrestrial assemblages in the Dakota Formation and younger coastal rocks in and around BRCA, CEBR, and ZION.

As the Western Interior Seaway withdrew, the region began to take on its present topographic landscape with the onset of the Laramide Orogeny, sometime before 70 Ma. This event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains. The Laramide Orogeny differed from the Sevier Orogeny by producing deeper, high-angled faults instead of the relatively shallow, low-angle faults of the older mountain-building event. One characteristic type of structural feature formed by the Laramide Orogeny is monoclines, large step-like folds. Many modern features of the NCPN were formed during the Laramide Orogeny, such as the Uinta Mountains of northeastern Utah and northwestern Colorado, and Waterpocket Fold of CARE. The geologic record of most the parks and monuments of the NCPN ends near the end of the Cretaceous, but the early Cenozoic is represented at BRCA, CEBR, FOBU, and just outside of DINO. In southern Utah, such as at BRCA and CEBR, there is evidence for a series of shallow lakes. Recent fieldwork in ARCH supports an isolated remnant of Eocene rocks and fossils. To the north, in FOBU and near DINO, there are rocks from much more extensive lakes, deposited as part of the Eocene-age Green River Formation. At various times during the early and middle Eocene, three large lakes existed in northeastern Utah, northwestern Colorado, and southwestern Wyoming: Fossil Lake, in the relatively narrow Fossil Basin (including FOBU) along the western border of Wyoming; Lake Gosiute, in several structural basins north of the Uinta Mountains; and Lake Uinta, in the Uinta Basin south of the mountains. These three lakes preserve diverse fossils of angiosperm plants, mollusks, crustaceans, insects, fish, amphibians, turtles, small reptiles, crocodilians, birds, and mammals. Fossil Lake, which appears to have been the deepest, most stable, and least saline of the three lakes, has by far the most diverse fish assemblage, as is well-known from FOBU and the vicinity.

Rock formations deposited between the Eocene lakes and the Quaternary are uncommon in the parks and monuments of the NCPN, being represented by a few poorly-known and sparsely fossiliferous conglomeratic units in BRCA, CEBR, DINO, and ZION. The rarity of rock formations does not mean that the NCPN units experienced no geological activity, though. Tectonic extension during the Oligocene caused the collapse of the eastern Uinta Mountains near DINO and the reorganization of drainage in the area. Extension spread through the western NCPN beginning about 15 Ma and continuing to the present. This east-west extension, known as Basin and Range extension for its characteristic geography of small north-south mountain ranges and narrow basins, has rearranged the topography of the Great Basin and adjacent areas of the NCPN. It is also responsible for some of the large fault systems of southern Utah, such as the Sevier and Paunsaugunt fault systems. Erosion and, in some areas, salt dissolution at depth have also greatly modified the landscape.

In addition to periods of uplift and extension, igneous activity affected parts of the NCPN, particularly areas on the periphery of the network. Oligocene-age volcanic tuffs are found at BLCA and CURE, from eruptions north and south of the Black Canyon. To the west, the Marysvale Volcanic Field of south-central Utah directly affected the southwestern NCPN, including BRCA, CARE, CEBR, and ZION. Marysvale volcanics extend from Salina and the Fish Lake Plateau in the northeast to the Markagunt Plateau in the southwest. Volcanism began in the Marysvale field by 34 Ma and tailed off by around 14 Ma, but small-scale eruptions occurred as recently as 500,000 years ago. The southern part of the volcanic field collapsed under its own weight around 20 Ma, causing faulting. Related to the Marysvale eruptions is a series of deeper intrusions, mostly formed between 22 and 20 Ma, known as the Iron Axis. Deep intrusions also formed several mountain ranges in southern Utah, such as the Abajo Mountains, the Henry Mountains, and the La Sal Mountains. More recently, during the Quaternary, small volcanic eruptions occurred in the Navajo Lake Volcanic Field, between Cedar City and Panguitch Lake. Small Quaternary volcanic features can be observed at ZION.

The past 2.6 million years are known as a time of glacial advances (“ice ages”) and retreats. The large continental ice sheets did not reach the NCPN, but alpine glaciers formed on some of the high peaks and plateaus, including on Brian Head

near CEBR. A much more direct influence on the NCPN was a series of lakes in western Utah, made possible by the different climate conditions. A wide area of western and northern Utah is part of a large basin, known as the Bonneville Basin. Lakes sporadically occupied parts of the basin as long ago as 5 Ma, becoming entrenched several hundred thousand years ago. The largest iteration, Lake Bonneville, occupied 51,700 km^2 (20,000 $miles^2$) at its greatest extent, around 19,000 years ago. Lake Bonneville underwent catastrophic flooding around 17,400 years ago, and then dropped to a level near that of the present Great Salt Lake some 3,000 years later as the climate changed. Bonneville Basin lake sediments can be found at GOSP and near TICA. Large Quaternary lakes also occupied areas near FOBU (Bear Lake in Utah and Idaho) and parts of ZION.

All of the parks and monuments of the NCPN have some form of Quaternary sediment, and many of them have Quaternary fossils. Quaternary fossils in the NCPN are typically either megafaunal remains or paleoecologically-important fossils like dung, packrat middens (former rodent nests cemented by urine and now well-preserved by the arid climate in crevices and caves), or pollen; caves are important locations for finding well-preserved bones, dung, and middens. Regarding the megafauna, a variety of large mammals including sloths, bears, canids, big cats, proboscideans, horses, pronghorns, deer, bovids, and camels inhabited the region, although most of them either left or went extinct as the last glacial period ended and humans entered the region 13,000 to 11,000 years ago. Megafaunal fossils are not common in the NCPN units, but are present at ARCH, COLM, FOBU, NABR, TICA, and ZION. Paleoecological fossils are much more abundant; the last 40,000 or so years of NCPN history are well-represented by fossil packrat middens and other such fossils from ARCH, CANY, CARE, CEBR, COLM, DINO, GOSP, NABR, TICA, ZION, and possibly CURE. These fossils show changes in climate through vegetation, illustrating how the modern northern Colorado Plateau vegetation developed as conditions changed.

During the past 12,000 years, there has been a shift from the Pleistocene ice-age climate. The Pleistocene-Holocene boundary was a time of climate change, shifting to a drier and warmer climatic regime. Humans have been present in the NCPN for at least 11,500 years. They have interacted with the area's fossil resources in a variety of ways. For example, sites left by the Fremont people at DINO often include marine fossils, sometimes drilled to serve as ornaments. Fossil resources are part of the modern culture of the communities near DINO and FOBU, where paleo-tourism and (in the FOBU area) the commercial collection of fossils are important components of the economy. Commercial collection has been established in Fossil Basin since the end of the 19^{th} century, and commercial collectors have made many important discoveries in the basin.

Paleontological resources have been reported from NCPN lands from a scientific perspective since the 1870s, when Othniel Charles Marsh found sparse dinosaur fossils near what is now DINO, and the first collections were made from Fossil Butte. Elmer S. Riggs began extracting fossils from the Morrison Formation just outside of COLM in 1900, and in 1909 Earl Douglass opened the Dinosaur Quarry at what would become DINO. These early discoveries can be observed in many institutions; the fish of Fossil Basin can be found in the collections of many museums, and the dinosaurs and other vertebrates of DINO and the COLM area form important parts of the collections of several leading museums. Outside of these three areas, fossil discovery was sporadic in the parks and monuments of the NCPN for many decades. Interest in paleontological resources in the NCPN units began increasing in the 1980s, and several paleontological surveys have been made in individual parks and monuments, including at ARCH, BLCA, BRCA, CEBR, COLM, CURE, DINO, GOSP, and ZION. These surveys have resulted in abundant fossil discoveries.

Collectively, the paleontological resources of the NCPN contribute to a greater understanding of the history of life on earth. Continued paleontological resource inventories will serve to expand this ever-widening base of paleontological knowledge represented throughout the National Park Service. Although at least 234 parks have already been identified as containing paleontological resources, much of what is to be known about the history of life on earth remains to be discovered.



Figure 1. Map of Northern Colorado Plateau Network parks. Courtesy NCPN Staff.

Report Organization and Methodology

This report summarizes the paleontological resources of the Northern Colorado Plateau Inventory & Monitoring Network (NCPN). The report consolidates baseline paleontological resource data for each network park to support management operations, planning and science-based decision-making as required by NPS Management Policies and the Paleontological Resources Preservation Act (2009). This section of the report details the organization of the report and summarizes how the report was written.

Each network report includes an abstract, introduction with geologic background, methodology chapter (this chapter), description of resource management strategies, and individual park narratives, which are described below. Some network reports include information for non-network parks, those that are not officially part of the Inventory and Monitoring Network, but are within the geographic boundaries of the network. Such parks were included in instances where paleontological resources were documented. For more information regarding different paleontological resource inventory strategies, refer to the Paleontological Resource Management chapter.

This 2012 Northern Colorado Plateau Network Paleontological Resource Inventory report represents a substantial update of the previous NCPN paleontology summary completed by Koch and Santucci (2002). The Methodology section, below, describes the changes between the original report and this document.

Organization of the Park Narratives

Each park narrative contains the following sections.

Geologic Background

The Geologic Background section contains basic geographic information for the park and lists pertinent geology and paleontology references. A brief geologic history is included to provide context for the fossils discovered within or near the park.

Sources of Paleontological Resources

The Sources of Paleontological Resources section reviews each fossil-bearing rock unit within the park. The history of paleontological resource field work in the park is summarized. A brief geologic description is included as well as the scope and significance of paleontological resources known to occur within the park. The section is organized stratigraphically from oldest rock unit to youngest.

Potential Sources of Paleontological Resources

The rock units listed in the "Potential Sources" section are units mapped within the park which potentially contain fossils. Although fossils have not been previously documented from those units within the park, fossils have been discovered within those units at localities outside of the park. In some cases the lack of documented fossils within a particular park or a specific geologic unit in a park, may simply be due to the fact that field inventories for fossils have never been undertaken in the park. As with the "Sources" section, the geologic units are listed stratigraphically from oldest to youngest.

Park Collections

The Park Collections section summarizes museum collections of fossils associated with the park. Such collections may be housed within the park or in external museum or university repositories. Fossil collections maintained by the park may include specimens discovered from both within and outside the park. It is important to recognize that the park's museum Scope of Collections may include fossils from localities outside of the park boundaries.. Regional museums and other outside repositories may curate fossils collected within the park prior to or subsequent to the park's establishment. Both contexts are included in the report. The Smithsonian Institution National Museum of Natural History and U.S. Geological Survey maintain significant collections of National Park Service fossils.

Cultural Resource Contexts

Resource stewardship strategies and management plans typically address paleontological resources which are known to occur within the rocks of the park. However, fossils may also occur in a variety of cultural resource contexts including: in association with archeological resources, ethnographic stories and legends, prehistoric and historic structures, or other documented historical occurrences. Fossils may be found as tools, jewelry, spiritual or ceremonial objects, or other cultural objects in archeological sites. A few ethnographic stories and legends told by American Indians and "mountain men" of the American West incorporate fossils. Building stones often contain fossils. In addition, archives, journals, memoirs, and photographs include other historical accounts of fossils in NPS areas. Kenworthy and Santucci (2006) summarizes such occurrences and provides examples.

Paleontological Resource Management, Preliminary Recommendations

In addition to geology and paleontology content, the authors suggest "next steps" for paleontological resource management and stewardship at each park. The recommendations are based on discussion with park staff and other geologists or paleontologists, or as part of broader NPS paleontological resource management strategies. These strategies are summarized in the next chapter. Parks may choose to address the recommendations based upon available staff time, training, and budget. Contact the NPS Geologic Resources Division for technical assistance or further guidance.

Literature Cited and Additional References

Each park narrative has a bibliography which is divided into two sections. References mentioned in the park narrative are included in the Literature Cited section. References discovered during the data-mining process that maybe of secondary value or interest are listed in the Additional References section. These publications were not included in the narrative but are relevant to the park, commonly containing additional detail beyond the scope of the narrative.

Data Sets

The Data Sets section includes a list of research projects, museum collections, NPS Geologic Resources Inventory-produced geologic map data, archives, or other files and information relevant to the fossils of a particular park.

Figures, Photographs and Graphics

The first network-based paleontological resource summaries (completed between 2002 and 2011) did not include figures, photographs or other graphics aside from the report cover and a network map. This second generation of Inventory & Monitoring Network based paleontological resource inventory reports includes basic geologic maps, stratigraphic columns, paleontological resource summary tables, and photographs for each park when available. Many of the geologic maps are adapted from the NPS Geologic Resources Inventory (GRI)map data for each park. These data are available in GIS format and support accompanying GRI reports, all of which are available from the GRI publications page: http://www.nature.nps.gov/geology/inventory/gre_publications.cfm. The stratigraphic columns illustrates a compiled section of all the rock units mapped in the park along with rock unit name, age, map symbol, thickness, and lithology. Also new to the second generation network-based reports is a summary table listing all the fossiliferous geologic units within the park, fossils collected within the park and the inferred depositional environment. Photographs of park fossils are inserted where appropriate.

Report Methodology

Inventory & Monitoring Network-based paleontological resource inventories compile baseline paleontological resource data for each of the parks within a particular network, and as such require considerable research and data mining. To minimize costs, field work is not normally conducted as part of the report preparation process. This section details the research methodology and report preparation, review, publication and distribution for network-based paleontological inventories.

Between 2002 and 2011, network-based inventories were completed for all of the 32 I&M networks. In 2001, the Northern Colorado Plateau Network technical committee supported a modest funding proposal to develop a prototype paleontological resource inventory for the network. Koch and Santucci (2002) represented the first attempt at compiling baseline paleontological resource data at a network level. That report laid the foundation upon which similar work would be completed for each of the I&M Networks in the National Park System. Between 2002 and 2006, 13 network reports were completed using small amounts of funds provided from each network. Funding from the NPS I&M Program was designated to systematically complete the remaining 19 reports between 2006 and 2011.

This second generation report represents a substantial revision of the original Northern Colorado Plateau Network summary, originally completed in 2002 (Koch and Santucci 2002). This updated Network-based paleontological resource inventory report represents a new template and format for future revisions to other network summaries.

Although field work is still not typically performed in association with Network-based reports, in a few instances field work complimented the original park inventories In some cases, field work has been conducted in response to recommendations provided in the initial Network-based paleontological resource inventory reports. Any new field data are incorporated into the updated reports.

Research

One of the most important components of any paleontological resource inventory is an intensive literature search and data mining. The American Geological Institute's GeoRef is the leading database for geology references, indexing more than 3.3 million references to North American geology dating back to the 1660s. To compile the appropriate literature base for this report, the authors conducted intensive GeoRef searches for general and park-specific publications.

Additional databases or information catalogs include the NPS Integrated Resource Management Applications (IRMA) portal (https://irma.nps.gov/App/Portal/Home), the U.S. Geological Survey (USGS) library system (http://library.usgs.gov/) and publications warehouse (http://pubs.er.usgs.gov/), as well as museum, university, and other institutional libraries and archives.

The literature search also includes unpublished material, or "gray literature," from park files, museum archives, local newspapers, field notes and other sources. In addition to literature searches, interviews with park staff, university faculty, geologists from the USGS and state geological surveys, and even local amateur geologists or paleontologists can often yield unpublished information regarding park paleontological resources. These interviews frequently result in capturing data that might otherwise be undocumented and potentially lost or unrecognized.

Report authors develop a comprehensive list of geologic units within each park by consulting published geologic maps and other sources. Geologic maps show the types of rocks and the associated geological formations within a park area. These maps

alone can often suggest the potential for paleontological resources to occur within a park. The NPS Geologic Resources Inventory (GRI) program is producing standardized digital geologic map data (GIS) for all Inventory and Monitoring Program parks. When available, the updated network-based paleontological resource inventory reports adapt GRI map data for each of the parks.

The nomenclature for geologic formations evolves over time and can differ based upon different geological interpretations. To maximize consistency with other NPS geology products, this report generally followed the geologic nomenclature used in GRI maps and report when they were available. The USGS maintains the National Geologic Map Database (NGMDB), which includes a searchable lexicon of valid geologic formation names (http://ngmdb.usgs.gov/Geolex/geolex_home.html). This lexicon provides a basic summary for each formation, summarizes current and past usage of the various formation names found in the literature, and provides an annotated bibliography for each formation.

Narrative

Information from these various sources is compiled into a narrative report for each park. The narrative is organized according to the scheme detailed earlier in this section.

Review

The park narratives and compiled report are extensively peer reviewed by professional geologists or paleontologists, as well as NPS staff from each park and the Geologic Resources Division. Upon completion of a compiled, draft report, network staff have the opportunity to provide review. Network staff also determine if the previous technical review was adequate for publication.

Publication and Distribution

The peer-reviewed, network-approved reports are published in the NPS Natural Resource Technical Report series. To minimize production costs, hard copies are not generally produced. The reports are distributed in PDF format to park and network staff and available to other NPS staff via the Integrated Resource Management Applications portal. Report reviewers and paleontologists working in the parks may also receive copies. The reports are not intended for, or distributed to, the general public due to the sensitive nature of paleontological resource locality information.

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Inventory and Monitoring Strategies and Resource Management Guidance

Paleontological Resource Inventory Strategies

In order to better document fossil occurrences and to provide baseline paleontological resource data to National Park Service areas, the NPS Geologic Resources Division (GRD) and the NPS Inventory and Monitoring Division (I&M) have established three paleontological resource inventory strategies. These strategies include: comprehensive park-specific inventories, Servicewide thematic inventories, and Inventory and Monitoring Network-based summaries.

Comprehensive Park-Specific Surveys

Comprehensive park-specific surveys identify all known paleontological resources within a single NPS unit. A team of specialists from within the NPS and from educational institutions and cooperators work with the targeted park to identify and address all aspects of its paleontological resources. The surveys commonly include sections addressing: park geology, paleontological resources inventory (species list and other compilations), localities, museum collections and curation, interpretation, management and protection, and recommendations. Paleontological resource training for park staff is an important component of many comprehensive park paleontological resource inventories.

Park-specific comprehensive paleontological resource inventories have been completed for Amistad National Recreation Area (Visaggi et al. 2009), Arches National Park (Swanson et al. 2005), Bighorn Canyon National Recreation Area (Santucci et al. 1999), Chesapeake and Ohio Canal National Historical Park (Clites and Santucci 2010), Death Valley National Park (Nyborg and Santucci 1999), Santa Monica Mountains National Recreation Area (Koch et al. 2004), Walnut Canyon National Monument (Santucci and Santucci 2001), Yellowstone National Park (Santucci 1998, the first park

to have such a survey completed), and Zion National Park (DeBlieux et al. 2005). As of April 2012, similar paleontological resource inventories are currently underway at Big Bend National Park, Glacier Bay National Park and Preserve, and Golden Gate National Recreation Area.

Servicewide Thematic Inventories

Servicewide thematic paleontological resource inventories compile data regarding specific types of paleontological resources that occur in parks throughout the NPS. These reports offer brief summaries of each occurrence of a particular type or category of fossils documented within a particular park. These reports are based primarily on literature review and communication with park staff and professional geologists and paleontologists.

The first thematic paleontological resource inventory identified 19 NPS units that preserve fossil vertebrate tracks (Santucci et al. 1998). Subsequent discoveries have increased the number of parks identified with fossil vertebrate tracks to 25 and warranted an updated publication (Santucci et al. 2006). Another thematic inventory identified paleontological resources associated with NPS caves in 35 parks (Santucci et al. 2001). Servicewide thematic inventories have also been initiated for fossil fish (Hunt et al. 2006), fossil coprolites (Hunt et al. 2012), fossil packrat middens (Tweet et al. in prep.), and fossils found in cultural resource contexts (Kenworthy and Santucci 2006).

Inventory and Monitoring Network-Based Summaries

Network-based paleontological resource inventories compile baseline paleontological resource data for each of the parks within a particular I&M Network. Between 2002 and 2011, Network-based inventories were completed for all 32 I&M networks. This report represents an overhaul and substantial revision of the Northern Colorado Plateau I&M Network report, the first I&M Network report completed. For more information on the content of network-based summaries and how they are completed, refer to the "Report Organization and Methodology" chapter. Citations for the completed reports are listed in the Literature Cited section.

A comprehensive summary report is currently being developed to present an overview and highlights of the 10 year effort to document fossils throughout the NPS based on the network-based strategy. Four summary reports are planned, including reports which present data based on geologic time, taxonomy, museum collections and fossil type specimens. These reports are anticipated to be finalized in 2013.

Paleontological Resource Monitoring Strategies

Following the compilation of baseline paleontological resource inventories, monitoring of fossil localities represents the next phase in paleontological resource management. Santucci and Koch (2003) presented a preliminary strategy for monitoring paleontological resources and identified natural and human-related impacts to fossils. Santucci et al. (2009) presented a more comprehensive strategy for paleontological resource monitoring which identified natural and human-related stressors upon paleontological resources, described five vital signs, and provided examples of study design and case studies. The five vital signs include: (1) erosion (geologic factors); (2) erosion (climatic factors); (3) catastrophic geohazards; (4) hydrology/bathymetry; and (5) human access/public use. The vital signs discussions provide detailed recommendations for resource managers, incorporating expertise, personnel, and equipment needed, approximate cost, and labor intensity. Santucci et al. (2009) is one chapter in the Geological Monitoring manual that provides guidance for monitoring various geologic resources and processes. The manual (Young and Norby 2009) was published by the NPS Geologic Resources Division and Geological Society of America.

Paleontological Resource Management Legislation and Guidance

I&M Network reports such as this are meant to provide network parks with sound baseline paleontological resource data that can stimulate future research, interpretation, education, or resource management projects. Science-based management of resources identified through these inventories is mandated by many NPS policies including the newly enacted Paleontological Resource Preservation Act (PRPA).

2009 Paleontological Resource Preservation Act

The Paleontological Resource Preservation Act (P.L. 111-11; "PRPA") was signed into law on March 30, 2009 as part of the Omnibus Public Lands Act of 2009. PRPA will serve as explicit authority for the management, protection and interpretation of paleontological resources in parks. PRPA regulations are currently (March 2012) being finalized by the National Park Service, Bureau of Land Management and the U.S. Forest Service. PRPA specifically provides the NPS with the following mandates to enhance paleontological resources stewardship:

- §6302 calls for the management and protection of paleontological resources using scientific principles and expertise. The NPS should develop plans for inventory, monitoring, and the scientific and educational use of paleontological resources. Planning should emphasize interagency coordination and collaboration and where possible include non-federal partners, the scientific community and the public.
- §6303 calls for the establishment of education programs to increase public awareness about the significance of paleontological resources.
- §6304 calls for the development of a specific permit for the collection of NPS paleontological resources. The new

legislation and other existing authorities clarify and reaffirm issues of property ownership, accountability, access and confidentiality of locality information associated with the management of NPS paleontological resources.

- §6305 calls for the curation of NPS paleontological resources, along with any associated data or records, in approved repositories.
- §6306 provides clarity regarding prohibited acts involving paleontological resources and specifies criminal penalties associated with these prohibited acts.
- §6307, along with other existing authorities, enables the NPS to seek civil penalties and restitution for the violation of any prohibited activities involving paleontological resources.
- §6308 provides the NPS a confidentiality provision with an exemption from the disclosure of any information associated with the nature and specific location of NPS paleontological resources.
- §6310 directs the Secretaries of the Interior and Agriculture to issue regulations appropriate to carry out the Act

NPS Management Policies

In addition to the PRPA, several other authorities influence the management of NPS paleontological resources. The 2006 National Park Service Management Policies (§1.4.6) stipulates that paleontological resources are considered park resources and values that are subject to the "no impairment" standard set forth by the NPS Organic Act in 1916. Basic guidelines for management of paleontological resources are found in sections 4.8.2 and 4.8.2.1 of the 2006 NPS Management Policies (http://www.nps.gov/policy/MP2006.pdf).

Other Resources

Paleontological resource management issues were the subject of a Report of the Secretary of the Interior in May 2000, viewable online at: http://www.blm.gov/heritage/docum/fossilrpt.pdf. That report was prepared in response to a Congressional request for an assessment of the need for a unified federal policy on the collection, storage and preservation of fossils and for standards that would maximize the availability of fossils for scientific study.

The Paleontological Resources Management section of Natural Resource Management Reference Manual 77 provides guidance and additional information regarding the implementation and continuation of paleontological resource management programs (http://www.nature.nps.gov/rm77/paleo.cfm).

As described above, Santucci et al. (2009) provided guidance for paleontological resource monitoring.

The NPS manages more than 445,000 catalogued fossil specimens in collections across the country. Many thousands more specimens await cataloging. For information regarding fossils in museum collections, refer to the NPS Museum management handbook, particularly Part I, Appendix U (http://www.nps.gov/museum/publications/handbook.html).

Education and Outreach Opportunities

Fossils occupy an enduring place in the hearts and mind of the public, presenting excellent opportunities for education and outreach. These opportunities support provisions in the Paleontological Resources Preservation Act (above) as well as the NPS natural resource stewardship mission. The NPS Geologic Resources Division coordinates two fossil-related education and outreach opportunities with a national scope: National Fossil Day and the Junior Paleontologist program. National Fossil Day is a celebration organized to promote public awareness and stewardship of fossils, as well as to foster a greater appreciation of their scientific and educational value. Learn more at the NFD website: http://nature.nps.gov/geology/nationalfossilday/index.cfm. The Junior Paleontologist program (part of the Junior Ranger program) provides opportunities for children to explore the ways that paleontologists work, learn about earth's history, and protect fossils and the rocks in which they are found. For more information, visit the Junior Paleontologist program webpage: http://nature.nps.gov/geology/nationalfossilday/jrpaleo.cfm. Contact the Geologic Resources Division for information on how to participate in either program.

In conclusion, our knowledge of the fossil record is only as good as our previous field season. The potential for new paleontological discoveries is proportionally related to our understanding as managers and stewards of this non-renewable evidence of life from the past. We believe that the baseline information provided in these reports and the resulting increased understanding of paleontological resources will inevitably result in paving the way for future fossil discoveries in NPS areas.

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General NCPN Information

Northern Colorado Plateau I&M Network: http://science.nature.nps.gov/im/units/NCPN/index.cfm

Geological Survey Websites

Arizona Geological Survey: http://www.azgs.az.gov/

Colorado Geological Survey: http://geosurvey.state.co.us/Pages/CGSHome.aspx

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American Geological Institute: http://www.agiweb.org/

Library catalogs

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Museums

American Museum of Natural History, New York City: http://www.amnh.org/

Brigham Young University Museum of Paleontology: http://cpms.byu.edu/ESM/index.html

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Denver Museum of Nature and Science: http://www.dmns.org/

Field Museum: http://fieldmuseum.org/

Museum of Northern Arizona: http://www.musnaz.org/

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Utah Field House Museum of Natural History State Park Museum: http://stateparks.utah.gov/parks/field-house

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Utah State University Eastern Prehistoric Museum: http://www.ceu.edu/museum/

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Dixie Geological Society: http://sites.infowest.com/dgs/

Four Corners Geological Society: http://www.fourcornersgeologicalsociety.org/

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U.S. Geological Survey/National Oceanographic and Atmospheric Administration North American Packrat Midden Database: http://esp.cr.usgs.gov/data/midden/

Utah Geological Association: http://www.utahgeology.org/

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Resource Management/Legislation Documents

NPS 2006 Management Policies (§1.4; Park Management): http://www.nps.gov/policy/mp/chapter1.htm

NPS 2006 Management Policies (§4.8.2.1; Paleontological Resources): http://www.nps.gov/policy/mp/chapter4.htm

NPS 1998 Omnibus Management Act (paleontological resource summary): http://www.nature.nps.gov/geology/paleontology/paleo_5_1/index.htm

NPS Natural Resource Management Reference Manual #77 (paleontology section): http://www.nature.nps.gov/rm77/paleo.cfm

Assessment of Fossil Management on Federal & Indian Lands: http://www.fs.fed.us/geology/fossil.pdf

National Park Service Paleontology Program Sites

NPS Geologic Resources Division: http://www.nature.nps.gov/geology/

NPS GRD Geologic Resource Inventory products (scoping summaries, GRI reports, digital geologic maps): http://www.nature.nps.gov/geology/inventory/gre_publications.cfm

NPS Paleontology Program: http://www.nature.nps.gov/geology/paleontology/

NPS Park Paleontology Newsletter (published 1999-2004): http://www.nature.nps.gov/geology/paleontology/newsletter.cfm

I&M Network paleontological resource summary project website and PDFs of all reports (InsideNPS): http://inside.nps.gov/waso/custommenu.cfm?lv=4&prg=753&id=4518

NPS Technical Information Center (Denver, repository for technical (TIC) documents): http://etic.nps.gov/

Other Geology/Paleontology Tools

Bates, R. L. and J. A. Jackson, editors. American Geological Institute dictionary of geological terms (3rd Edition). Bantam Doubleday Dell Publishing Group, New York.

NeoMap (Neogene Mammal Mapping Portal): http://www.ucmp.berkeley.edu/neomap/

Neotoma (Quaternary fossil database): http://www.neotomadb.org/

NOAA Paleoclimatology: http://www.ncdc.noaa.gov/paleo/paleo.html

Paleobiology Database (search for fossil taxonomy, localities, or geologic formations): http://paleodb.org/cgi-bin/bridge.pl

Paleontology Portal (general fossil info based on geographic location or geologic time): http://www.paleoportal.org/

University of California-Berkeley Museum of Paleontology online paleontology exhibits: http://www.ucmp.berkeley.edu/

U.S. Geological Survey digital state geology maps: http://tin.er.usgs.gov/geology/state/

U.S. Geological Survey Geologic Terms Glossary: http://www.nature.nps.gov/geology/usgsnps/misc/glossaryAtoC.html#A

U.S. Geological Survey Paleontological Terms Glossary: http://geology.er.usgs.gov/paleo/glossary.shtml

U.S. Geological Survey National Geologic Map Database (NGMDB): http://ngmdb.usgs.gov/

U.S. Geological Survey Geologic Names Lexicon (GEOLEX; geologic unit nomenclature and summary): http://ngmdb.usgs.gov/Geolex/geolex_home.html

U.S. Geological Survey Geographic Names Information System (GNIS; search for place names and geographic features, and plot them on topographic maps or aerial photos): http://gnis.usgs.gov/

U.S. Geological Survey GeoPDFs (download searchable PDFs of any topographic map in the United States): http://store.usgs.gov (click on “Map Locator”).

U.S. Geological Survey Publications Warehouse (many USGS publications are available online): http://pubs.er.usgs.gov/

U.S. Geological Survey, description of physiographic provinces: http://tapestry.usgs.gov/Default.html

Arches National Park

Arches National Park (ARCH), Utah, was proclaimed a national monument on April 12, 1929 and was redesignated a national park November 12, 1971. ARCH is famous for its scenic geology, consisting of fins, pedestals, pinnacles, windows, and the greatest concentration of natural arches in the world (ARCH Figure 1). The boundaries of ARCH changed November 25, 1938; July 22, 1960; January 20, 1969; and October 30, 1998. It currently encompasses 30,966.13 ha (76,518.98 acres), all but 67.17 ha (165.97 acres) of which are under federal administration (ARCH Figure 2). ARCH is part of the Southeast Utah Group of the National Park Service (NPS), headquartered in Moab, Utah. The Southeast Utah Group also includes Canyonlands National Park (CANY), Hovenweep National Monument (HOVE), and Natural Bridges National Monument (NABR); summaries for these parks and monuments are also included in this report.

Geologic Background

ARCH is located in southern Grand County in eastern Utah. The Colorado River forms part of the eastern and southern boundaries. Much of northern ARCH is within the Salt Valley, a salt anticline formed by the movement of salt deposits at depth. Similar processes occurred at nearby CANY (see the CANY summary in this document for more information), about 18 km (11 miles) to the southwest. The city of Moab is across the Colorado River to the south of the park, and Canyon Rims BLM Recreation Area is just beyond Moab to the south.



ARCH Figure 1. Natural arches, windows, and bridges have a variety of shapes and sizes; the person in shadow near the bottom of this image shows the scale of this arch, part of Double Arch (JST).

Befitting a park created for its striking scenic geology, ARCH is discussed in a number of geologic publications. References that are useful for understanding the general geology of ARCH include: Dane (1935); Lohman (1975); Baars (1983); Doelling (1985, 1988, 2004, 2010); and Graham (2004). References concerning the paleontology of ARCH include: Dyer et al. (1983); Oviatt (1988); Agenbroad and Mead (1989); Lockley (1991); Mead et al. (1991); Sharpe (1991, 1993); Duffy (1993); Britt (1996); Engelmann (1995, 1999); Hasiotis (1996a, 1996b); Hasiotis and Demko (1996); Smith and Betancourt (1998); Engelmann and Fiorillo (2000); Santucci et al. (2000, 2001); Koch and Santucci (2002); White and Lockley (2002); Lockley et al. (2004); Foster (2005); Swanson et al. (2005); Wright et al. (2007); Lockley and Gierliński (2009); and Santucci and Kirkland (2010). The Geologic Resources Division of the NPS coordinated a geologic resource inventory scoping session for ARCH, CANY, HOVE, and NABR during May 1999 (NPS 1999), and produced a geologic resource inventory report for ARCH in 2004 (Graham 2004). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002). A paleontological survey was prepared for the park by Swanson et al. (2005).

The geologic record exposed at ARCH covers the Middle Pennsylvanian to the Late Cretaceous, some part of the pre-Quaternary Cenozoic, and the Quaternary (see the appendix for a geologic time scale). The oldest exposed rocks are more than 300 million years old and date to the middle and late Pennsylvanian. At this time, ARCH and much of southeastern Utah and southwestern Colorado were within the Paradox Basin, a large structural basin. The Pennsylvanian rocks are part of the basin fill. These rocks, belonging to the Hermosa Group, were deposited during a Paleozoic ice age (Rueger 1996). No glaciers were present near the Paradox Basin, but the rocks do preserve sea level fluctuations caused by glacial/interglacial cycles, which were in turn influenced by Milankovitch orbital cycles (Blakey 2008), as are documented for the recent glacial and present interglacial periods. Sea level fluctuations are well-documented in the Paradox Basin, which at the time was a deep basin rimmed by reefs during sea level highstands, with deposition of vast amounts of salt near the center of the basin during lowstands (Blakey 2008). The salt deposits have had important implications for the structural geology and modern topography of ARCH and the vicinity (Baars 1983; Doelling 2010). Soon after deposition, salt was squeezed upward, and eventually formed a series of salt walls at depth that were aligned over basement faults that permitted the Paradox Basin to subside (sink) (Doelling 1988). Sedimentary units which accumulated in the basin during the Pennsylvanian, Permian, and Triassic show a variety of effects from this diapirism (forcing of a more mobile material, in this case salt, through brittle surrounding rock). Their thicknesses vary greatly, with thin or absent deposits over the walls and thick deposits between

them, and the changing geometry of the depositional surface caused folding and tilting. Angular unconformities (gaps in time where tilting of the underlying rock during a depositional hiatus has resulted in differing attitudes of the underlying and overlying rock) resulted, even within formations; this can be seen at ARCH between the upper and lower sections of the Triassic-age Chinle Formation. The formation of the salt walls took place over 75 million years (Doelling 2010), between about 300 and 225 Ma (million years ago) (Doelling 1988).

At the beginning of the Permian, around 300 Ma, a period of mixed terrestrial and marine deposition represented by the Cutler Formation followed the shallow marine conditions. Tectonic activity continued during the deposition of the largely Permian-age Cutler Group, with the Paradox Basin sinking while the Uncompahgre Uplift formed to the east and provided sediment to fill the basin (Melton 1972; Blakey 2008; Baars 2010). The ARCH–CANY area at this time was approximately equatorial (Huntoon et al. 2002). Deposition occurred in coastal marine and terrestrial settings as a shallow sea transgressed (advanced over the land) during the Early Permian (Condon 1997; Baars 2000). Large coastal ergs (giant dune fields) developed on the down-wind sides of the sea (McKee 1979; Blakey 2008; Blakey and Ranney 2008) as it expanded and contracted several times (Huntoon and Kamola 1988). A gap or hiatus of at least 25 million years follows the Early Permian rocks at ARCH.

Triassic rocks at ARCH are principally represented by the Moenkopi Formation of the Early–Middle Triassic and the Chinle Formation of the Late Triassic. Southern Utah during the Early and Middle Triassic was a broad coastal plain that was occasionally flooded by a sea to the west (Stewart et al. 1972a; Blakey et al. 1993; Hintze and Kowallis 2009). The top of the Moenkopi Formation is separated from the Chinle Formation by millions of years, with intervening erosion (Pipiringos and O'Sullivan 1978). The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time (Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003).

Chinle Formation rocks show that the area gradually became drier over time as the supercontinent Pangaea moved north. During its deposition, the Four Corners area of the Southwest moved from near the equator to arid-zone latitudes. What had been a setting of rivers, lakes, and deltas turned arid (Tanner 2003). This is represented at ARCH by the first of a series of giant ergs, the Wingate Sandstone of the Glen Canyon Group (Blakey and Ranney 2008). The Wingate Sandstone erg of the Late Triassic and Early Jurassic was ended by a more humid period recorded in the Kayenta Formation. The Kayenta Formation represents a broad, gentle flood plain separated from the sea to the west (Blakey 1994; Peterson 1994). About 190 Ma, this setting was gradually replaced by the vast erg that became the Navajo Sandstone (Harshbarger et al. 1957; Dickinson and Gehrels 2003, 2009; Rahl et al. 2003; Lucas et al. 2005), the largest erg known from North America (Blakey 2008; Blakey and Ranney 2008). This Sahara-like desert was found throughout the Four Corners area into Nevada and Wyoming (Hintze and Kowallis 2009). At this time, the latitude of the area was only about 10° N, near the western edge of the supercontinent Pangaea (Loope et al. 2004a). Strong cross-equatorial winds blew from the northwest during the winter, driving dune movement, and monsoonal rains occurred during the summer (Loope et al. 2004a, 2004b).

Eventually, another shallow sea began to encroach on central Utah from the north and west, setting the stage for another series of transgression-regression cycles and their resulting shallow marine and coastal terrestrial deposition (Doelling 2010; Morris et al. 2010). The marine body was west of ARCH (Doelling 2010). It was active at about the same time as an outbreak of volcanic activity on the southern margin of the Colorado Plateau dated to approximately 170 Ma (Blakey and Parnell 1995). Aridity decreased through the Late Jurassic (Lockley and Wright 2003) as Pangaea continued moving north; southern Utah reached 22° N during this time (Loope 2008). The more humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the last geologic formation of the Jurassic in ARCH, the well-known Morrison Formation. By the time it was deposited, the area had moved out of the low subtropical latitudes (Lockley and Mickelson 1997). The Morrison Formation is interpreted as a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004). During deposition of the upper portion of the formation (the Brushy Basin Member), a vast basin is thought to have extended from near Grand Junction, Colorado, to Albuquerque, New Mexico. This basin was periodically occupied by a shallow saline lake which has been named Lake T'oo'dichi'; the setting is described as analogous to the modern Lake Eyre basin in Australia (Turner and Fishman 1991).

The Early Cretaceous is represented at ARCH by the Cedar Mountain Formation, which is similar to the Morrison Formation in lithologies and depositional settings (Doelling 2010). During the deposition of this formation, a transition began from internal drainage and low sedimentation, as seen in the upper Morrison Formation, to greater sedimentation in a foreland basin near a sea, as seen in the overlying Dakota Formation (Eberth et al. 2006).

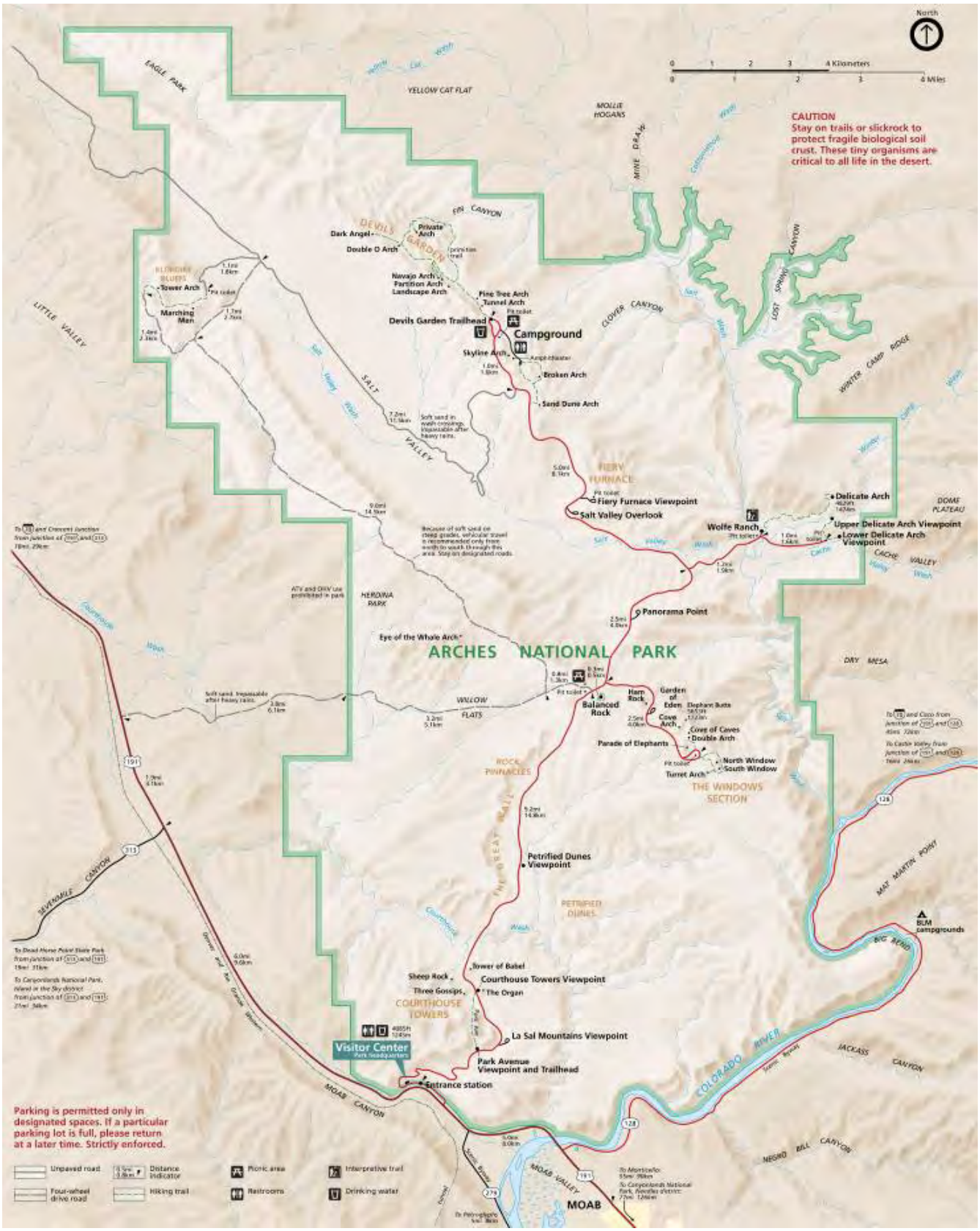
Terrestrial processes continued until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Formation signals the onset of several cycles when the Western Interior Seaway transgressed and regressed (withdrew) over the land (Gustason 1989; Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to

the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Marine deposition at ARCH from the Seaway is represented by the Mancos Shale (Doelling 2010).

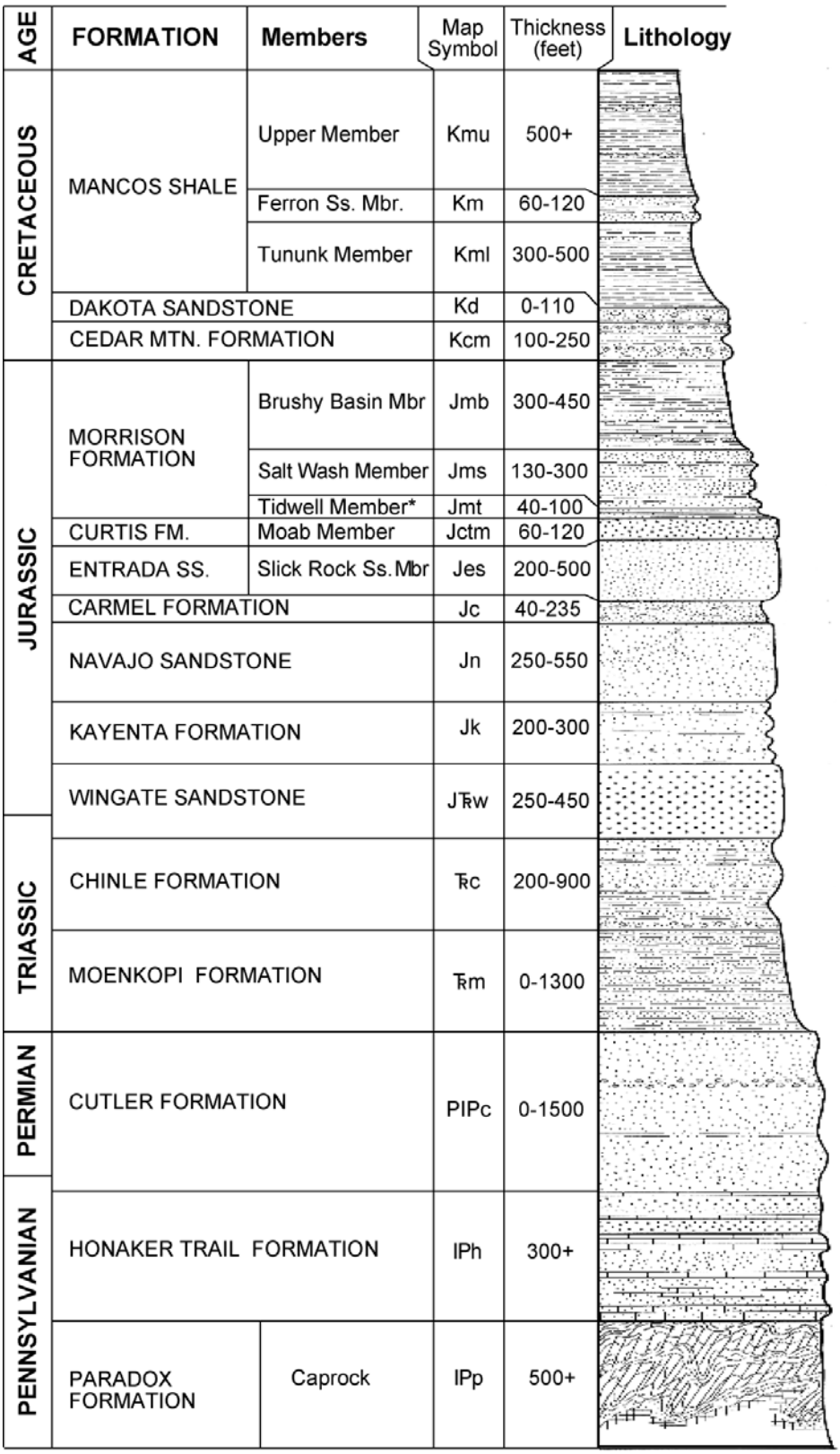
Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of a major mountain-building event known as the Laramide Orogeny (Cather 2003). This event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). The Paradox Basin was relatively unaffected by the Laramide Orogeny. Old structural features such as the San Rafael Swell and Uncompahgre Plateau underwent renewed uplift, but the great salt deposits may have buffered the rocks surrounding them (Baars 1983). Instead, anticlinal folding (convex folds) developed over salt walls, with associated faulting. The Salt Valley anticline, now collapsed, is an example of one of these salt-cored anticlines. Additional, more recent uplift and associated erosion have led to the collapse of the salt-cored anticlines by removing sedimentary layers that had previously sealed the salt walls, permitting fresh ground water to infiltrate and dissolve the salt (Doelling 2010). Extensive downcutting and canyon formation followed episodes of uplift (Doelling et al. 2010). In addition to periods of uplift, local igneous activity occurred on the Colorado Plateau during the late Eocene, Oligocene, and Miocene. Igneous bodies emplaced at depth formed the Abajo, Henry, and La Sal mountain ranges south of ARCH between about 31 and 23 Ma (Nelson et al. 1992; Nelson 1998; Nelson and Davidson 1998).

The long post-Cretaceous hiatus at ARCH ended by at least the Pliocene or Pleistocene (some previously-overlooked pre-Pleistocene Cenozoic deposits are present in Salt Valley; J. Kirkland, Utah Geological Survey state paleontologist, pers. comm., October 2011), assisted by the movement of salt. The Salt Valley anticline began to collapse by the Late Pliocene, leading to the formation of small depositional basins throughout the valley. Sediments in these basins were folded and faulted as salt rose or dissolved at depth (Oviatt 1988). Volcanic events that occurred in California and the Yellowstone area are recorded in the thick deposits of Salt Valley. There are two named and dated ash layers: the Bishop Ash from 730,000 years ago; and the Lava Creek B Ash from 610,000 years ago (Doelling 2010). Quaternary deposits at ARCH preserve evidence of the biota that lived there during the Ice Age. The megafaunal extinctions on the Colorado Plateau claimed such animals as the mammoth, the ground sloth *Nothrotheriops shastensis*, the shrub ox *Euceratherium collinum*, and the mountain goat *Oreamnos harringtoni*. Dated dung deposits from the Colorado Plateau indicate that they went extinct between 11,800 and 11,000 radiocarbon years ago (Mead and Agenbroad 1992). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 [http://calib.qub.ac.uk/calib/] where appropriate. The Colorado Plateau was cooler and more moist during the last glacial period, becoming more arid afterward. It has been most arid during the last 1,500 years (Elias et al. 1992). It had been thought that both megafauna and human cultural sites more than 6,000 years old were rare on the Colorado Plateau, but this is now known to be inaccurate (Agenbroad 1997). Humans have been present in the area for at least the last 10,000 years (Graham 2004), but most physical evidence for people at ARCH is much younger, dating from about the past 5,000 years (Lohman 1975).

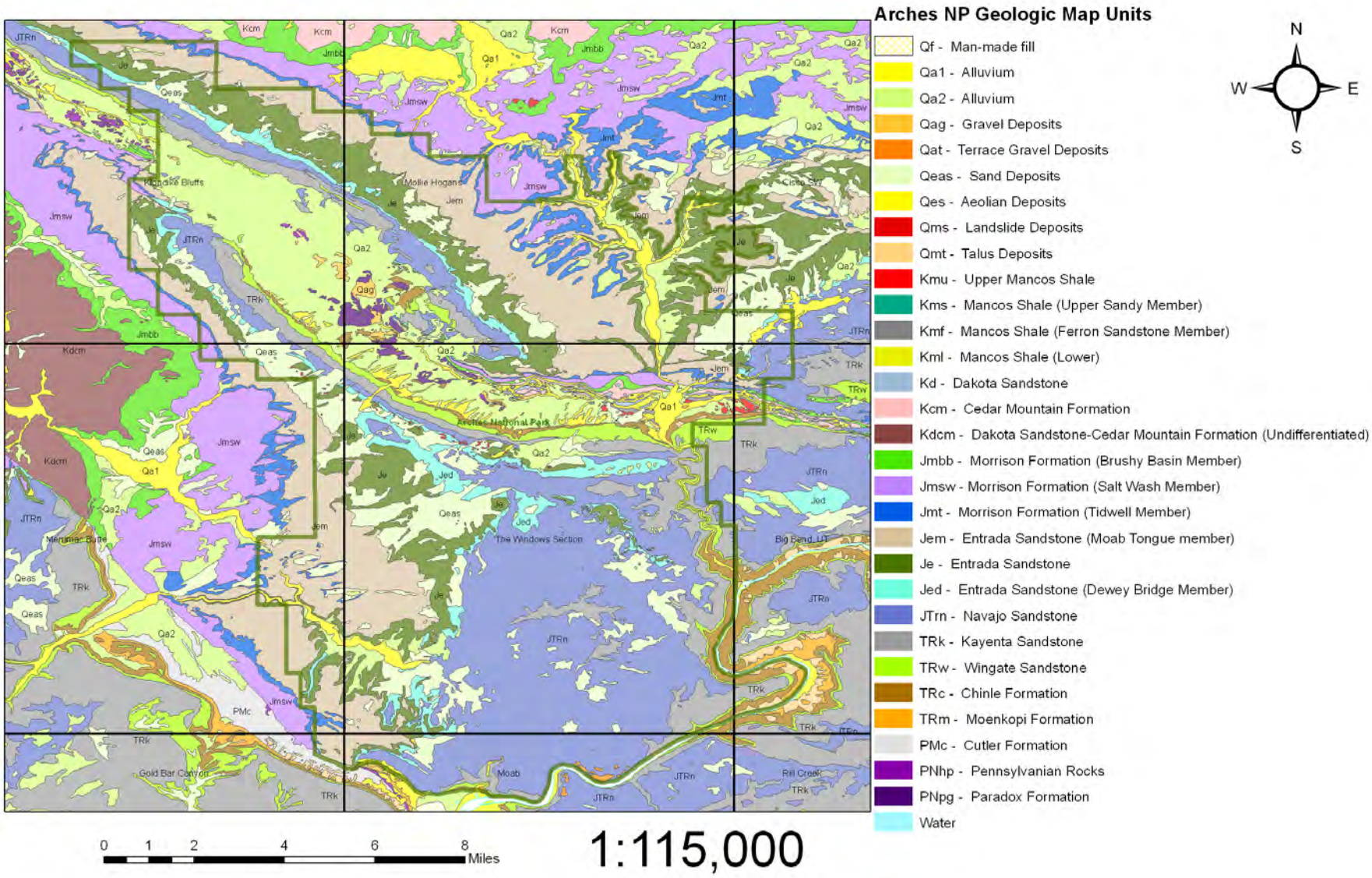
Rock units exposed within ARCH include, from oldest to youngest: the Paradox Formation and Honaker Trail Formation of the Hermosa Group (Middle–Upper Pennsylvanian); the Cutler Formation (Permian); the Moenkopi Formation (Lower–Middle Triassic); the Chinle Formation (Upper Triassic); the Wingate Sandstone of the Glen Canyon Group (Upper Triassic–Lower Jurassic); the Kayenta Formation and Navajo Sandstone of the Glen Canyon Group (Lower Jurassic); the Carmel Formation and Entrada Sandstone of the San Rafael Group (Middle Jurassic); the Curtis and Summerville formations of the San Rafael Group and the Morrison Formation (Upper Jurassic); the Cedar Mountain Formation (Lower Cretaceous); the Dakota Formation (Lower–Upper Cretaceous); the Mancos Shale (Upper Cretaceous); and remnants of Cenozoic rocks in Salt Valley (at least some known to be Eocene). Surficial units include a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene) (see ARCH Figure 3 for a stratigraphic column and ARCH Figure 4 for a geological map) (Doelling 1985, 2010; Graham 2004). The Cenozoic deposits of Salt Valley will be designated as a new unit in the future (J. Kirkland, pers. comm., October 2011). Paleontological resources at ARCH are known from the Paradox Formation (marine invertebrates), Honaker Trail Formation (marine invertebrates), Moenkopi Formation (reptile tracks), Chinle Formation (plant fossils, bones, and trace fossils), Wingate Sandstone (a tooth of a phytosaur, a crocodile-like reptile), Kayenta Formation (trace fossils and bone), Navajo Sandstone (trace fossils), Curtis Formation (dinosaur tracks), Morrison Formation (plant fossils, dinosaur bones, and various trace fossils), Cedar Mountain Formation (plant fossils, dinosaur bones, and various trace fossils), Dakota Formation (plant fossils), Mancos Shale (marine invertebrates) and Quaternary deposits (diatoms [algae that secrete silica "shells"], isolated mammal bones, and packrat middens). All of the other formations are potentially fossiliferous within the park (ARCH Table 1).



ARCH Figure 2. Geography of ARCH (NPS).



ARCH Figure 3: Stratigraphic column for ARCH (from Doelling [2010], used with permission of the Utah Geological Association).



ARCH Figure 4: Schematic geological map of ARCH. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1045870.

Formation	Age	Fossils Within ARCH	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Diatoms, root traces, ostracodes, a mammoth jaw bone, bison and sheep bones, and packrat middens	Alluvial, eolian, fluvial, and landslide deposits
Cenozoic remnants	Eocene	Bivalves and gastropods	Gastropods typical of the Green River Formation are present, so a lacustrine and fluvial system is likely
Mancos Shale	Late Cretaceous	The bivalve *Pycnodonte* and other marine invertebrates	Open marine, shallow marine, and coastal settings
Dakota Formation	Early–Late Cretaceous	Plant fossils	Terrestrial (especially fluvial) becoming shallow marine over time
Cedar Mountain Formation	Early Cretaceous	Sauropod dinosaur bones, dinosaur tracks, and possible feeding traces	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	Petrified wood, bivalves, invertebrate traces, and dinosaur bones (including from *Apatosaurus*)	Fluvial, floodplain, and lacustrine settings
Summerville Formation	Late Jurassic	None to date; invertebrate and vertebrate trace fossils are most likely	Coastal delta
Curtis Formation	Late Jurassic	Dinosaur tracks (mostly theropods)	Desert
Entrada Sandstone	Middle Jurassic	None to date; fossils are very rare in this formation in general	Desert with large eolian sand dunes
Carmel Formation	Middle Jurassic	None to date; pollen, spores, and vertebrate trace fossils are most likely	Marine–coastal terrestrial transitional settings
Navajo Formation	Early Jurassic	Invertebrate traces, burrows, and dinosaur tracks	Desert with large eolian sand dunes
Kayenta Formation	Early Jurassic	Invertebrate burrows, theropod bones, and dinosaur tracks	Fluvial, floodplain, and eolian settings
Wingate Sandstone	Late Triassic–Early Jurassic	A phytosaur tooth	Desert with large eolian sand dunes
Chinle Formation	Late Triassic	Leaves, stems, seeds, a possible conifer cone, *Sanmiguelia* foliage, root traces, burrows, bones, and theropod tracks	Fluvial, floodplain, lacustrine, and eolian settings, becoming drier over time
Moenkopi Formation	Early–Middle Triassic	*Chirotherium* reptile tracks	Coastal plain
Cutler Formation	Permian	None to date; plant fossils and marine invertebrates are most likely	Marine and arid coastal terrestrial settings

Formation	Age	Fossils Within ARCH	Depositional Environment
Honaker Trail Formation	Middle–early Late Pennsylvanian	Foraminifera, algae, sponges, corals, bryozoans, brachiopods, bivalves, cephalopods, gastropods, trilobites, crinoids, echinoids, and trace fossils (assemblage from area including land inside and outside of ARCH)	Open marine
Paradox Formation	Middle Pennsylvanian	Corals, bryozoans, brachiopods, trilobites, and crinoids	Restricted marine, cycling from normal marine water to evaporative conditions (a "dead sea")

ARCH Table 1. Concise tabulation of ARCH stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of ARCH present opportunities for education, interpretation, and continued or future scientific research in the park. Systematic exploration for fossils at ARCH did not begin until the 1990s, but reports of fossils have been made from the park as far back as 1933, when the discovery of dinosaur bones within the then-Arches National Monument was announced in a Moab newspaper. The area was known among paleontologists as a place to find dinosaur bones (W. Langston, University of Texas at Austin professor emeritus, pers. comm., February 2011). Wann Langston reports having seen a locality littered with bone fragments near the base of a mesa during a visit to ARCH perhaps 40 years ago; the locality was marked with a sign warning visitors to not remove bones. On a later visit, he found the locality had been stripped of fossils (W. Langston, pers. comm., February 2011). Several surveys have taken place in ARCH since 1995. New sites were found by George Engelmann (1994, 29 sites), Brooks Britt (1995, 7 sites), and two phases of a 2000 survey effort, begun that May by a joint NPS/Utah Geological Survey/Utah Friends of Paleontology group and continued in the summer by Brooke Swanson (31 sites) (Swanson et al. 2005). Further discoveries were made in 2011, including a number of firsts: the first fossils from the Moenkopi Formation (reptile tracks); the first fossils of the plant *Sanmiguelia* and the first bones from the Chinle Formation; the first fossils from the Wingate Sandstone (a phytosaur tooth, establishing a Triassic age for part of the formation); the discovery of dinosaur bones in the Kayenta Formation; and the first tracks from the Navajo Sandstone (J. Kirkland, pers. comm., October 2011). Given the vast areas of rock outcrops, diversity of fossiliferous rock units, and abundant fossil resources in outcrops of the same units in the immediate vicinity of ARCH, more fossil localities should be expected.

Hermosa Group: Paradox Formation (Middle Pennsylvanian)

The Paradox Formation is a thick heterogeneous unit. Salt is present in the formation at depth, with 29 documented cycles of salt deposition (Baars 1983), but because of its solubility it is not found where the formation is exposed at the surface. Instead, outcrops are of a residual layer called caprock. The Paradox Formation that outcrops in ARCH is composed of this caprock, with limestone, gypsum, and shale present. Only the upper part of the formation is exposed. The upper contact with the Honaker Trail Formation is a disconformity (Doelling 2010). In older references, the Paradox Formation and Honaker Trail Formation of ARCH are depicted as members of the Hermosa Formation (Lohman 1975). The Paradox Formation was deposited during the Middle Pennsylvanian, with the oldest exposed rocks in the ARCH–CANY area from the middle Middle Pennsylvanian (Baars 2010), approximately 307 Ma. The Paradox Formation's depositional area was a restricted basin with cyclic aridity, represented by cycles of rocks laid down under normal marine to more saline conditions (Wengerd 1962). It has been described as a "dead sea" (Baars 1983). While marine in origin, evaporites are common, and invertebrate fossils are rare (Rueger 1996). Because it was based on evaporitic beds, authors may choose to not map the Paradox Formation where such beds are not present, even if correlative layers are (Franczyk et al. 1995).

Fossils are known from the Paradox Formation of ARCH: rugose corals, bryozoans (moss animals), brachiopods (lamp shells), trilobites, and crinoids (sea lilies) have been reported from an exposure created by blasting for drainage (Swanson et al. 2005). Other types of fossils have been reported from the vicinity. Palynomorphs (organic microfossils such as pollen and spores) have been recovered from cores taken north, northeast, and east of CANY, representing arborescent and herbaceous coal swamp ferns and gymnosperms on the western shore of the seaway (Kosanke 1995; Rueger 1996), and microbial mound limestones are known in the CANY area (Fallin 1985). Other fossils known from the Paradox Formation include stromatolites (layered sedimentary structure formed by microbial trapping of sediment) (Chidsey and Eby 2000), algae, fusulinid foraminifera (foraminifera are amoeba-like organisms that form "shells," and fusulinids are large foraminifera from the late Paleozoic useful for correlating and dating beds), sponges, echinoids (sea

urchins), conodonts (extinct eel-like chordates), and fecal pellets (Ritter et al. 2002).

Hermosa Group: Honaker Trail Formation (Middle–lower Upper Pennsylvanian)

The Honaker Trail Formation of ARCH is mostly composed of limestone and sandstone, with some shale and siltstone. The limestone beds form ledges and cliffs, are gray and fossiliferous, and contain chert and clay minerals. The sandstone beds can be a variety of colors, and also form ledges and cliffs. Doelling (2010) reported only one outcropping area of the formation at ARCH, in the extreme southwest near the Visitor Center. This is continuous with an area outside of the park's boundaries (Doelling 2010), which has been described in detail by Melton (1972). The older but larger-scale map of Doelling (1985) also depicts a few small outcrops in Salt Valley. As noted, the Honaker Trail Formation of ARCH was described as a member of the Hermosa Formation in older references ("unnamed upper member" of Lohman [1975]). It was deposited during the Middle to early Late Pennsylvanian (Baars 2010), approximately 306 Ma. The upper contact is a disconformity with the Cutler Formation (Doelling 2010). It is a marine formation at ARCH, where offshore sand deposits intertongued with marine limestone. There was open circulation instead of the closed circulation of the Paradox Formation that led to salt and gypsum deposits (Doelling 2010), during this, the most extensive marine period in the Paradox Basin (Rigby et al. 1971). The climate was arid and seasonal (Melton 1972). Sediment from the Uncompahgre Uplift to the east was deposited as the Paradox Basin subsided (Melton 1972; Doelling 2010).

The Honaker Trail Formation and overlying units have gone through several interpretations (see the CANY summary in this document for more details). Briefly, the Hermosa Formation was raised to the Hermosa Group and divided into formations in 1958, including the Paradox Formation and overlying Honaker Trail Formation (Wengerd and Matheny 1958). The definition and extent of the main part of the Honaker Trail Formation is broadly accepted within the ARCH–CANY area, whereas the upper part of the formation and overlying strata have been controversial and the subject of several name changes. This interval had been known as the Halgaito Formation or the Rico Formation (an obsolete term of historical interest; Anderson et al. [2010]), but was redefined and named the Elephant Canyon Formation with the type locality at CANY's Elephant Canyon in 1962, with some of the lower portion assigned to the Honaker Trail Formation (Baars 1962). Because names and defined contacts differ significantly among workers, the relationships of the units are not one-to-one, so references using one system of nomenclature cannot necessarily be translated by substituting another. These considerations can make tracing references and correlating usages difficult.

While the Honaker Trail Formation is known to be fossiliferous, discerning what is known from it is obscured by the nomenclatural issues discussed above. This is more problematic for CANY, which has more extensive outcrops and different stratigraphy. In particular, it is not always possible to establish what unit is being discussed in references using "Hermosa Formation," such as those prior to 1958. Of interest for ARCH, McKnight (1940) mentioned the presence of foraminifera, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, trilobites, and crinoids in an area of Utah including part of CANY and farther north, and Rigby and Stokes (1971) described sponges from limestone ledges of the Hermosa Formation near Spanish Bottom in CANY. Geological data for this publication is minimal, and the geographical information could place the find in the Honaker Trail Formation or the Elephant Canyon Formation.

The Honaker Trail Formation is fossiliferous within ARCH, and includes marine invertebrates such as corals. Melton (1972) recovered algae, fusulinid foraminifera, sponges, corals, bryozoans, brachiopods, bivalves, cephalopods, gastropods, trilobites and other arthropods, crinoids, echinoids, and invertebrate trace fossils from several locations in the upper Honaker Trail Formation near the Visitor Center along the park's southern boundary. The primary site was just outside of the park to the south. The majority of the fossils were found in various limestone lithologies, with shales yielding bryozoans, brachiopods, bivalves, and gastropods, and sandstones only yielding trace fossils (Melton 1972). Bradyodont and cladodont shark teeth are also reported from across from the park entrance (Swanson et al. 2005). Doelling (2004) noted an area in the Klondike Bluffs with possible anomalous areas of the Honaker Trail Formation; ledges in these areas have what may be reworked Mississippian-age fossils. Fusulinids from the latest Pennsylvanian have been recovered just west of the park (Baars 1962; Doelling 1988). A similar invertebrate assemblage is known from CANY (Baars 1993, 2000; Santucci and Kirkland 2010). Plants have been recovered from a sandstone unit in the middle of the formation south of the Texas Gulf Sulfur Potash plant, about 12 km (7.5 miles) south of ARCH. Fossils include coal, petrified logs, and compressed leaves from at least 20 taxa of lepidodendrales (clubmoss-like trees), horsetails, ferns, and conifers. The depositional environment is interpreted as a swampy delta (Tidwell 1988).

Other Honaker Trail Formation fossils include palynomorphs (Kosanke 1995), root traces from eolian rocks (Loope 1994), other foraminifera, conodonts, fecal pellets (Ritter et al. 2002), borings (Loope 1994), and burrows (Loope and Kuntz 1987, as the upper member of the Hermosa Formation).

Moenkopi Formation (Lower–Middle Triassic)

The Moenkopi Formation is a heterogeneous unit present in several western states. In Utah, it was deposited on a broad coastal plain sloping gently toward a sea in the west, affected by occasional marine transgressions (Stewart et al. 1972a; Hintze and Kowallis 2009). It is up to 400 m (1,300 ft) thick in the area, but its thickness is quite variable due to the topography created

by the Paradox Formation salt bodies. The upper contact with the Chinle Formation is a disconformity (Doelling 2010). There are probably at least 10 million years of deposition missing between the two formations, a gap known as the Tr-3 unconformity (Biek et al. 2010). The Moenkopi Formation dates to the Early and early Middle Triassic (Heckert et al. 2005a), between approximately 250 and 240 Ma. Within ARCH, it crops out mostly along Big Bend, with a few exposures in Salt Valley. Typical exposures include steep lower and upper slopes separated by a ledgy interval, with an upper thin ledge at the top in the Big Bend area. A thin conglomeratic bed is locally present at the base of the formation (Doelling 2010). Depositional environments include fluvial, mudflat, sabkha, and shallow marine settings, in the vicinity of a shallow sea to the west (Stewart et al. 1972a; Doelling 2010), which did not reach as far as ARCH (Lucas et al. 1997a).

The Moenkopi Formation has been divided into several members, depending on location. At ARCH and Moab, four members are recognized. From oldest to youngest, they are the Tenderfoot, Ali Baba, Sewemup, and Pariott members (Shoemaker and Newman 1959; Stewart et al. 1972a; Doelling 2010). The Tenderfoot and Sewemup members are the lower and upper steep slope-forming members mentioned above, and the Ali Baba and Pariott members are the ledgy intervals. The Tenderfoot Member is composed of interbedded chocolate brown silty sandstone, sandy mudstone, siltstone, and shale (Doelling 2010). In part, it correlates to the Hoskinnini Member found elsewhere in central and southern Utah (Stewart et al. 1972a). The beds become progressively thicker leading into the Ali Baba Member, which is similar but has thicker ledge-forming beds. Unlike the lower transition, the transition from the Ali Baba to the Sewemup Member is abrupt. This member is a lighter shade of brown. The Pariott Member is found only in the Big Bend exposures. It is composed of red-brown to lavender sandstone interbedded with brown, orange-brown, and red siltstone, mudstone, and shale (Doelling 2010).

The Moenkopi Formation of ARCH is fossiliferous. During field work in 2011, examples of the reptile track genus *Chirotherium*, an index fossil for the Moenkopi Formation, were discovered in the park in rocks formerly identified as part of the Chinle Formation (ARCH Figure 5) (J. Kirkland, pers. comm., October 2011). Previously, the Moenkopi Formation was not known to be fossiliferous at ARCH, and the members present at ARCH are not known for their fossils. The most likely fossils from these members would probably be terrestrial trace fossils, such as root traces, invertebrate burrows, and tetrapod tracks. The Moenkopi Formation is extensively fossiliferous as near as the CANY area (Santucci and Kirkland 2010). As noted above, a report of a fossil locality in the Sewemup Member a little over 3 km (2 miles) northwest of ARCH in Salt Valley with marine fossils (Shoemaker and Newman 1959; Stewart et al. 1972a) has turned out to be a misidentified Cedar Mountain Formation locality (Lucas et al. 1997a).



ARCH Figure 5. Chirotherium *tracks in the Moenkopi Formation (photo courtesy Jim Kirkland, Utah Geological Survey).*

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location (Stewart et al. 1972b). At ARCH, the formation has not been formally divided. Historically, it has been correlated to the Church Rock Member (Stewart et al. 1972b; Lupe 1979). It may include, from oldest to youngest, the Petrified Forest, Owl Rock, and Church Rock members (G. Willis, Utah Geological Survey mapping program manager, pers. comm., January 2012). Outcrops can be found in Salt Valley and along the Colorado River canyon. The formation is divided by an internal disconformity within the park. The lower Chinle Formation is composed of mottled, variegated, and interbedded sandstone, conglomerate, and siltstone, and mudstone, with the coarser beds forming ledges and the finer beds forming slopes (Doelling 2010). The mottled beds are thought to represent paleosols (ancient soil horizons), and includes root casts (Doelling and Ross 1998; Doelling 2010). The upper Chinle Formation has a ledge-forming interval bracketed by slope-forming intervals. The slope-forming intervals are composed of interbedded siltstone, mudstone, and sandstone, with thin ledges of coarser material. The upper slope-forming interval also includes some thicker sandstone beds, thought to represent eolian (wind-blown) deposits in advance of the erg that formed the Wingate Sandstone. The middle ledgy interval is composed of conglomeratic sandstone and sandstone; the uppermost part of this interval is sometimes identified as the Black Ledge (Doelling 2010), a widespread informal bed (Stewart et al. 1972b). Finer-grained intervals in the formation are interpreted

as floodplain deposits, and the coarser beds are interpreted as stream channels. The formation ranges from 60 to 245 m (200 to 800 ft) thick in the ARCH area (Doelling 2010). The upper contact with the Wingate Sandstone is generally gradational (Tanner 2003). The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) because this division is known to be in the overlying Wingate Sandstone (Lucas et al. 1997b, 2006a).

The Church Rock Member is a siltstone-sandstone unit, with stream (Blakey and Gubitosa 1983), lake margin, and eolian deposition, with eolian more prominent toward the top as the unit transitions in the Wingate Sandstone (Dubiel 1989). Deposition during the Church Rock Member shows the continuation of aridification begun in the underlying Owl Rock Member (Tanner 2003), as the dunes of the Wingate Sandstone approached from the northwest (Blakey and Gubitosa 1983).

The Chinle Formation is fossiliferous at ARCH. Seven localities were found by Brooke Swanson in the summer of 2000: locality BAS-17 has invertebrate burrow casts; BAS-18 has small leaves, stems, seeds, and a possible conifer cone; BAS-19 is an overhang site with burrows; BAS-20 has burrow casts large enough to have come from vertebrates (greater than 9 cm [3.5 in] in diameter); BAS-21 has spiral invertebrate burrows; BAS-22 has branching invertebrate burrows; and BAS-23 has arcuate invertebrate burrows (Swanson et al. 2005; invertebrate burrows illustrated in ARCH Figure 6). As mentioned, root casts can be found in the lower Chinle Formation of the ARCH area (Doelling 2010). Lockley and Gierliński (2009; they identify the unit as the Rock Point Formation of the Chinle Group) reported on two Moab-area track sites, one from a recessed roadcut on Highway 191 that is partially within ARCH. This site, UCD 13 2004 (University of Colorado at Denver Dinosaur Tracks Museum), has been known for over twenty years. Forty complete or partial small theropod tracks (bipedal predatory dinosaurs) assigned to the track genus *Grallator* were found on a single fallen slab, now cataloged by ARCH as ARCH 351. Matrimony Spring, a site across the Colorado River from ARCH, has approximately 50 *Grallator* footprints. Other track types found in the upper Rock Point Member include the reptilian *Rhynchosauroides*, enigmatic *Evazoum*, and sauropod ("brontosaur") *Eosauropus*. Lockley and Gierliński (2009) also noted that there are many track sites in the region. Similar tracks are known from the upper Chinle Formation at CANY (Lucas et al. 1995; Santucci et al. 1998). Several additions to the Chinle assemblage of ARCH were discovered in 2011, including the plant *Sanmiguelia* (ARCH Figure 7) and the first bones from the formation in the park, as well as more tracks (J. Kirkland, pers. comm., October 2011).



ARCH Figure 6. Invertebrate burrows from the Chinle Formation (VLS).

There are other reports of Chinle Formation fossils in the Moab area. Scott (1982) included a palynomorph locality from the Church Rock Member near ARCH. Cleveland and Buchanan (2005) reported three localities with the crayfish burrow *Camborygma eumekenomos* from around Moab. Bones, phytosaur teeth, and numerous theropod tracks have been found just west of ARCH, and abundant crayfish burrows have been found southwest of the park along the Colorado River below a within-formation angular unconformity (Santucci and Kirkland 2010). Finds from the uppermost members of the Chinle Formation in San Juan County to the south include ferns and other plants, bivalves, gastropods, ostracodes (seed shrimp), conchostracans (clam shrimp), fish, tetrapod bone fragments, and invertebrate and reptile tracks (Milner 2006; Spears et al. 2006). Among these is a significant fish site in Lisbon Valley (Spears et al. 2006).



ARCH Figure 7. Sanmiguelia *foliage from the Chinle Formation (photo courtesy Jim Kirkland, Utah Geological Survey).*

The various members of the Chinle Formation have similar types of fossils, but differ in the genera and species present and which are most commonly found. Plants are represented by charophyte algae (Lucas and Keitzke 1993), diverse palynomorphs (Gottesfeld 1972; Fisher and Dunay 1984; Parker 2004), and leaves, wood, and other remains of lycopods, horsetails, ferns and fern-like plants, cycads, bennettitales, ginkgoes, conifers, and taxa of uncertain classifications (Ash 1972, 1989, 2003, 2005a; Stewart et al. 1972b; Parker and Ash 2004). Fungal remains are also present (Ash 1972; Stewart et al. 1972b). Invertebrates are represented by tentaculites (enigmatic possible mollusks, misidentified as worm fossils [Ash 2005b]), bivalves, gastropods, ostracodes, conchostracans (Breed 1972; Stewart et al. 1972b), crabs (Rinehart et al. 2003), crayfish (Miller and Ash 1988), insects (Breed 1972; Stewart et al. 1972b), and various trace fossils (Caster 1944; Breed 1972; Gillette et al. 2003; Demko et al. 2005), including borings in wood (Walker 1938; Tapanila and Roberts 2012) and a variety of feeding traces on plants (Ash 1996). Reworked Paleozoic fossils, primarily of invertebrates, are occasionally found in the lowest part of the formation, and include foraminifera, algae, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, ostracodes, and crinoids (McKee 1937; Stewart et al. 1972b).

The Chinle Formation's vertebrate assemblage has been summarized in a number of documents, including Stewart et al. (1972b), Colbert (1972), Parrish (1989), Long and Murry (1995), Heckert et al. (2005a, 2005b), Irmis (2005a), Parker (2005, 2009), Lucas et al. (2007), Loughney et al. (2011), and Parker and Martz (2011). Vertebrates include: freshwater sharks; ray-finned fish; lungfish; coelacanths; several types of enigmatic tetrapods known from isolated bones and teeth; metoposaur amphibians; procolophonids (stocky lizard-like herbivorous and insectivorous reptiles); sphenodontians (relatives of the modern tuatara); trilophosaurids (reptiles that resembled large herbivorous lizards); tanystropheids (long-necked piscivorous reptiles); the unusual aquatic reptile *Vancleavea*; drepanosaurids (beaked, possibly aquatic chameleon-like reptiles); several forms of crocodile-like phytosaurs, known under a variety of names; several forms of herbivorous armored aetosaurs, also known under a variety of names; the small unusual armored *Revueltosaurus*; "rauisuchian" crocodile relatives including poposaurids and rauisuchids, which resembled large carnivorous dinosaurs, and shuvosaurids (toothless forms that resembled ostrich-mimic dinosaurs); "sphenosuchians" (small terrestrial crocodile relatives); lagerpetonids (small bipedal dinosaur relatives); silesaurids (quadrupedal dinosaur relatives); several early dinosaur taxa; and dicynodonts (a type of non-mammalian therapsid, or "mammal-like reptile"). Lungfish, metoposaurs, aetosaurs, and phytosaurs are the most common forms. Vertebrate trace fossils include coprolites (fossil dung) (Hunt et al. 1998), swim traces, and footprints possibly attributable to sphenodontians, tanystropheids, phytosaurs, aetosaurs, carnivorous and herbivorous dinosaurs, and mammal relatives (Hunt and Lucas 2007). Bones accumulated in floodplains, bogs, small ponds, and in or near channels (Parrish 1989).

Glen Canyon Group: Wingate Sandstone (Upper Triassic–Lower Jurassic)

The Wingate Sandstone is an eolian sandstone unit that spans the Triassic–Jurassic boundary (Lucas et al. 2006a). Limestone lenses are also present (Harshbarger et al. 1957). At ARCH, it forms stained red-brown sandstone cliffs. It is exposed in two bands flanking Salt Valley, along the lower Salt Wash, and in the Colorado River Canyon. It ranges from 75 to 135 m (250 to 450 ft) in thickness in the area, averaging 90 m (300 ft). The contact with the overlying Kayenta Formation is often difficult to identify, and could be conformable or disconformable; if disconformable, the gap in time was not great. The Wingate Sandstone is interpreted as eolian dune deposits from an erg (Doelling 2010); other depositional settings such as fluvial (Baars 2010) and sabkha (Huntoon and Dubiel 2000) are found elsewhere. Wind blew from the northwest (Patterson et al. 1989). After much debate concerning the age of the Wingate Sandstone and other Glen Canyon Group units, it was thought that the Wingate Sandstone was entirely Jurassic in age, but the discovery of a skull of the phytosaur *Redondasaurus* at the base of the formation in Lisbon Valley south of ARCH shows that at least some of the unit is Late Triassic in age (Lucas et al. 1997b). Trace fossils suggest that the Triassic–Jurassic boundary is in the middle to upper 75 m (245 ft) of the formation (Lucas et al. 2006a).

The Wingate Sandstone has had a complicated history (Colbert and Mook 1951). It was originally defined to contain much the same rocks as what was later named the Entrada Sandstone, but the Entrada Sandstone eventually took over most of these rocks, including the original type section for the Wingate Sandstone (at Fort Wingate, New Mexico). Furthermore, the Wingate Sandstone was later divided into the lower Rock Point Member

and the upper Lukachukai Member, but where the Rock Point Member has been assigned to the Chinle Formation (as discussed above), the Lukachukai Member is abandoned as redundant (Dubiel 1989).

The Wingate Sandstone is fossiliferous at ARCH; a phytosaur tooth was discovered in the formation in 2011, corroborating a Late Triassic age for part of the unit (J. Kirkland, pers. comm., October 2011). In general, fossils are uncommon in the Wingate Sandstone, with the phytosaur skull mentioned above being the only body fossil of note, but trace fossils are known from multiple levels in Utah, Arizona, and Colorado. Trace producers include invertebrates (Odier et al. 2004), mammal-like animals, nondinosaurian reptiles, and theropod, prosauropod, and possible sauropod dinosaurs (Odier et al. 2004; Lucas et al. 2006a, 2006b). Mammal-like tracks are often found nearly alone and are confined to dunes (Odier et al. 2004). There appears to be distinct Late Triassic and Early Jurassic assemblages, with mammal-like tracks limited to the Late Triassic and prosauropod tracks appearing in the Early Jurassic (Lucas et al. 2006a, 2006b). Several track sites are known from the lower Wingate Sandstone around Moab and ARCH. Two sites south of Moab in Kane Springs Canyon have tracks of *Grallator*, the reptilian *Brachychirotherium*, and *Eosauropus*. Two sites near Corral Canyon about 3.5 km (2.2 miles) west of ARCH have *Grallator* and *Brachychirotherium* (Lucas et al. 2006a).

Glen Canyon Group: Kayenta Formation (Lower Jurassic)

The Kayenta Formation is mostly made up of fine- to coarse-grained sandstone with some small amounts of interbedded shale and siltstone, and rare limestone and conglomerate (Vandiver 1936). At ARCH, it is a heterogeneous formation, typically red to purple in color. Most of it is fluvial sandstone lenses, with some eolian sandstone, and conglomerate, siltstone, and shale. Coarser-grained beds form ledges. Outcrops are widespread along Salt Valley and in the eastern part of the park. The formation is 60 to 90 m (200 to 300 ft) thick. Its contact with the Navajo Sandstone is intertonguing and can be difficult to place (Doelling 2010). The intertonguing contact represents a time when Navajo Sandstone dunes advanced over the rivers of the Kayenta Formation (Harshbarger et al. 1957). While dating of Glen Canyon Group units has proven difficult to establish, the age of the Kayenta Formation is now thought to be middle Early Jurassic, between approximately 197 and 190 Ma (Lucas et al. 2005). The Kayenta Formation was formed by shifting freshwater braided and meandering streams (Colbert 1974); floodplain deposits are also known (Lucas et al. 2005).

The Kayenta Formation is known for its dinosaur tracks, and a few are known at ARCH. Dinosaur tracks have been reported from two sites (BAS-10 and 11). Invertebrate burrow fossils have also been found at two sites in the park (BAS-9 and 12; the latter may be from the Navajo Sandstone). The sites are playa deposits, which may also yield tree stumps (Swanson et al. 2005). Two theropod tracks have been found in the park's Petrified Dunes area (Santucci and Kirkland 2010), which may be the same as BAS-10 and 11. The first dinosaur bones from this formation at ARCH were discovered in 2011, and belonged to a moderately-sized theropod comparable in scale to *Dilophosaurus* (J. Kirkland, pers. comm., October 2011), known from several specimens found in the Kayenta Formation of Arizona. Bivalves have been found from this formation in the Moab area, and dinosaur tracks and freshwater bivalves have been found from the Kayenta–Navajo transitional zone to the northwest of ARCH in the San Rafael Swell (Baker 1946). Additionally, reed impressions are known from the CANY area (Rigby et al. 1971).

Vertebrate body fossils of the Kayenta Formation have been described in many publications, but most are from a silty facies found to the south in Arizona (see Lucas et al. [2005] or the Navajo National Monument summary of the Southern Colorado Plateau Network report for a partial list). Vertebrate body fossils of the sandy facies of the Kayenta Formation, which would be typical of ARCH, include fish scales, crocodylomorphs (crocodiles and their relatives), and tritylodontid synapsids (early mammal relatives) such as *Kayentatherium* (Clark and Fastovsky 1986). Kayenta Formation vertebrate trace fossils include coprolites (Luttrell and Morales 1993), therapsid tracks, lizard-like tracks (Hamblin and Foster 2000), and dinosaur tracks such as *Grallator* and *Eubrontes* (Lucas and Tanner 2007). The relevant ichnofossil taxonomy can be complicated. A recent summary for the Chinle Formation and Glen Canyon Group can be found in Lucas et al. (2006b). Kayenta Formation fossils that are not vertebrates or vertebrate traces include microbial ("algal") limestone (Luttrell and Morales 1993), petrified wood (Jenkins and Shubin 1998), plant impressions (Rigby et al. 1971), freshwater bivalves and gastropods (Harshbarger et al. 1957), ostracodes (Lucas and Tanner 2007), and invertebrate trails and burrows (Luttrell and Morales 1993).

Glen Canyon Group: Navajo Sandstone (Lower Jurassic)

The Navajo Sandstone is a fine- to medium-grained quartz sandstone noted for its prominent cross-bedding formed by dune progression. There are also lenticular beds of cherty limestone (Harshbarger et al. 1957). At ARCH, the formation is composed of orange to light gray sandstone, with prominent cross-bedding. Localized limestone is also found, grading into reddish siltstone. The Navajo Formation forms the majority of the outcrops in the southeastern quarter of ARCH, and is 60 to 170 m (200 to 550 ft) thick. The upper contact with the Carmel Formation is a disconformity, which explains some of the variability in thickness (Doelling 2010). The bulk of the Navajo Sandstone probably dates to the middle and late Early Jurassic (Lucas et al. 2005), between approximately 190 and 175 Ma.

The Navajo Sandstone represents the largest erg known from North America (Blakey 2008). As discussed in the "Kayenta Formation" section, the upper Kayenta Formation intertongues with the lower Navajo Sandstone as the dunes of the latter

overrode the streams of the former. Recent research on geochronology of detrital zircon grains in the Navajo Sandstone suggests that most of the sand was eroded from the ancestral Appalachian Mountains, transported westward by a continental-scale river system to the western shore of Jurassic North America, and was then blown southward and incorporated into the Navajo–Nugget–Aztec dune field (Dickinson and Gehrels 2003, 2009; Rahl et al. 2003). Preservation of this tremendous thickness of wind-blown sand was made possible because of basin subsidence associated with Early Jurassic compressional deformation near the west margin of North America. This deformation caused the continental interior to flex downward, creating accommodation space for sediment accumulation (Allen et al. 2000).The Navajo Sandstone formed in a tropical arid setting with eolian, playa, and sabkha (salt flat) deposits (Mahaney et al. 2004). Wetter restricted interdunes are also known, and are characterized by mudstone with planar bedding, dolomitic limestone, and sandstone (Irmis 2005b). Interdune deposits are well-known from the Moab area, ranging from fluvial sandstones to lake limestones. During periods of increased precipitation, vegetation spread (Wilkens 2008). Interdune deposits are present within ARCH as well (Doelling 2010). Fossils are mostly associated with interdunes (Loope et al. 2004b). As time went on during the deposition of the Navajo Sandstone, the climate became drier (Irmis 2005b), although there were several periods of increased precipitation with durations on the order of tens of thousands of years (Milan et al. 2008).

The Navajo Sandstone is not extensively fossiliferous, but has yielded invertebrate traces from playa deposits at ARCH (locality BAS-16; Swanson et al. 2005). Burrows were reported at ARCH by Freeman and Visher (1975) and Picard (1975). Dinosaur tracks were first identified from this formation in the park in 2011 (J. Kirkland, pers. comm., October 2011). There is also evidence for microbial ("algal") marsh fossils in the ARCH–CANY area (Baars 1983). Interdune deposits outside of ARCH in the Moab area have yielded noteworthy fossils, and given the proximity, similar fossils may be found in ARCH. Moab-area finds are unusual when viewed against Navajo Sandstone fossils from elsewhere, as they are dominated by plant fossils from lake settings. Fossil trees are particularly well known. Some are *in situ* stumps, while others were knocked down and have been moved (Loope 1979). *In situ* trees may have been killed by rising water tables, while those that were moved may have been knocked down by dunes that collapsed from heavy rains. Some lived for decades and attained diameters of nearly 1 m (3 ft) (Parrish and Falcon-Lang 2007). The Moab interdune localities also preserve invertebrate burrows (Riese et al. 2011), unusual complex burrows possibly made by mammals and/or mammal relatives (Lucas et al. 2006c; Riese et al. 2011), and a variety of plant fossils besides trees, including root traces (Riese et al. 2011), plant hash, and impressions and natural casts and molds of plant material (Wilkens 2008). When precipitation increased, dry uplands had cycadeoids, while araucarian conifers grew along lake margins (Wilkens 2008).

To date, fossil organisms and traces represented in the Navajo Sandstone over its entire depositional area include stromatolites, root traces (Irmis 2005b), pollen, spores, horsetails (*Equisetum*), ferns (Loope et al. 2004b), cycadeoids, conifers (Wilkens 2008), petrified wood (Irmis 2005b), possible fungal filaments (Mahaney et al. 2004), worm burrows (Stokes 1978), insect burrows, scorpion tracks (Loope et al. 2004b), unionid bivalves (Wilkens 2008), ostracodes, conchostracans (Harshbarger et al. 1957), a crocodylomorph, the small bipedal carnivorous theropod dinosaur *Segisaurus* (Irmis 2005b), at least two types of sauropodomorph dinosaurs ("prosauropods") including *Seitaad* (Sertich and Loewen 2010), a tritylodontid synapsid, and vertebrate trace fossils (Irmis 2005b). The assemblage may be a persistent remnant of the Kayenta Formation assemblage (Winkler et al. 1991). Sandstone dune deposits have mostly yielded vertebrates, while limestone interdune beds have mostly yielded plants and invertebrates (Stokes 1991), although arthropod tracks (*Octopodichnus*) are known from the dunes (Hunt and Lucas 2006). Interdune fossils include stromatolites, root casts, horsetails, ostracodes, and insect traces, with locally common petrified wood (Irmis 2005b).

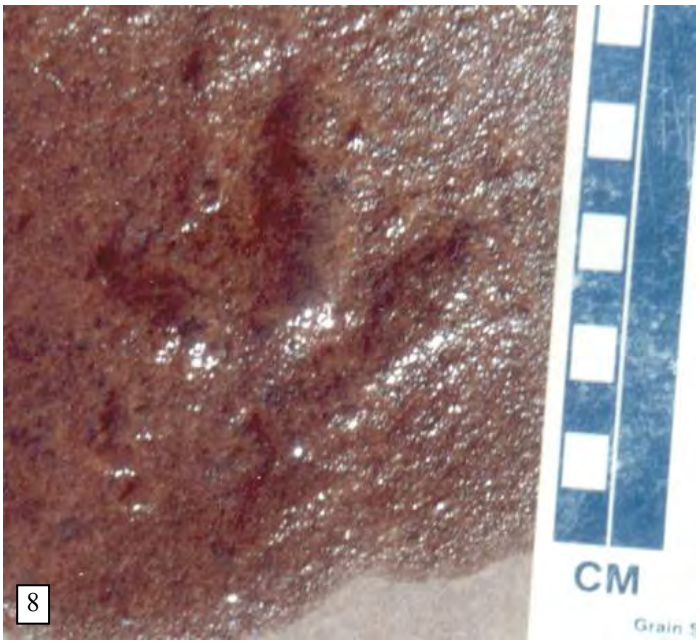
Vertebrate trace fossils are much more common than body fossils in the Navajo Sandstone, and have been extensively described. More than 60 dinosaur track sites are known in the Navajo Sandstone across its depositional area (Loope 2006; Milan et al. 2008). The various vertebrate trace fossils have gone under different names and interpretations, depending on the author. Recent works (Irmis 2005b; Hunt and Lucas 2006; Loope 2006) seem to have settled on seven or eight primary ichnogenera: *Brasilichnium* (synapsid, but see Milan et al. 2008 for an interpretation as a sauropodomorph), *Batrachopus* (crocodylomorph), *Grallator* (small theropod), *Anchisauripus* (medium theropod), *Eubrontes* (large theropod), *Navahopus* (small ?sauropodomorph), *Otozoum* (larger sauropodomorph), and *Anomoepus* (ornithischian or "bird-hipped" dinosaur) (Irmis 2005b). A few oddities are also known, including lizard-like tracks (*Lacertipus*), a track once thought to be pterosaurian in origin ("*Pteraichnus*"), bird-like *Trisauropodiscus* tracks (Irmis 2005b), undescribed large ornithopod (beaked herbivorous dinosaur) tracks (Santucci and Kirkland 2010), and large vertebrate burrows thought to have been made by synapsids (Lucas et al. 2006c). So far, large traces (*Anchisauripus*, *Eubrontes*, and *Otozoum*) are only known from interdunes, with smaller traces (*Batrachopus*, *Brasilichnium*, and *Grallator*) the only tracks found in dune sands (Loope 2006), although this differentiation may be due to preservational bias (Irmis 2005b). In general, the vertebrate trace fossils are consistent with the known body fossils.

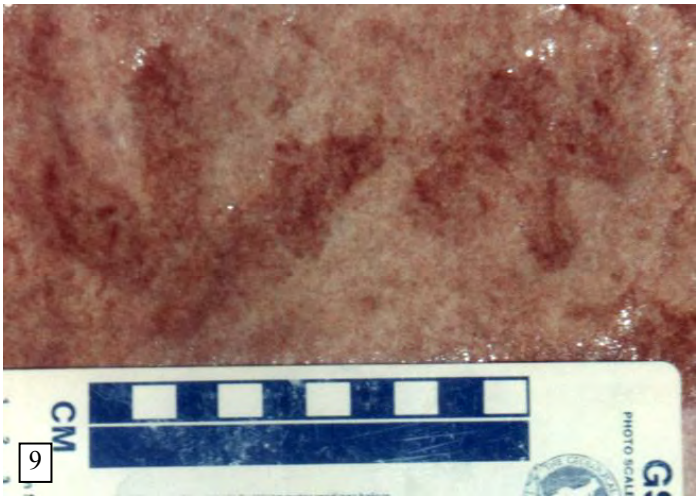
San Rafael Group: Curtis Formation (Upper Jurassic)

The Curtis Formation is exposed widely in the western, northwestern, and northeastern sections of ARCH. It is mostly made up of the resistant Moab Member (or Tongue), formerly of the Entrada Sandstone (see the Entrada Sandstone section of this

report for more details on the stratigraphic nomenclature). The Moab Member is composed of white- to gray-weathering sandstone that forms cliffs and generally resembles the Navajo Sandstone. It is 20 to 35 m (60 to 120 ft) thick in the area. There are also brown silty sandstones found beneath the Moab Member in the northern part of the park, which are more like typical marine Curtis Formation beds found to the west; this interval is up to 8 m (25 ft) thick (Doelling 2010). The Curtis Formation is generally marine, but the sea did not reach ARCH, and the Moab Member is interpreted as a desert (Swanson et al. 2005). The formation was previously thought to be late Middle Jurassic in age (Lucas and Anderson 1997), but is now known to date to the early Late Jurassic (Oxfordian stage), sometime between approximately 161 and 156 Ma (Wilcox 2007; Hintze and Kowallis 2009).

The Moab Member is extensively fossiliferous at and near ARCH. Five sites have been reported from within ARCH (BAS-3 through 7); all had theropod tracks (ARCH Figures 8 and 9), and one (BAS-3) had quadruped tracks, possibly from a sauropod (Swanson et al. 2005). The uppermost part of the unit, at the contact with the overlying Summerville Formation, hosts part of an extensive megatracksite (Lockley 1991; Lockley et al. 2007). This megatracksite is sometimes reported as from the Summerville Formation instead. Practically all of the tracks are of a large theropod (*Megalosauripus*) (Lockley and Hunt 1995), although another type of large theropod track (*Therangospodus*) has also been identified (Lockley et al. 1998a). Smaller theropod tracks have been found lower in the Moab Member (Lockley and Hunt 1995), including from at least eight locations just west of ARCH (Lockley et al. 2007). Tracks in the Moab Member near Moab have been known since 1936 (Lockley 1991); footprints from near Courthouse Wash west of ARCH were described by McKnight (1940), and area tracks were examined by track specialist Roland T. Bird in 1944, but the tracks were largely forgotten for many years (Lockley 1991). McKnight's tracks are examples of the smaller dinosaur tracks lower than the megatracksite horizon (Lockley et al. 2007). The megatracksite horizon is interpreted as a coastal plain (Lockley 1991). Horizontal burrows from 3 to 7 cm (1 to 3 in) in diameter are known from the upper Moab Member as well (Swanson et al. 2005). Invertebrate traces have been found in the main body of the Moab Member (Lockley et al. 2007). The Moab Member has produced few other described fossils over its depositional area. The only body fossil appears to be a poorly preserved partial skeleton of the crocodylomorph *Entradasuchus* from the Moab area (Hunt and Lockley 1995). The Curtis Formation also has a marine fossil assemblage, better known in northern Utah (Hoggan 1970; omitted here for lack of marine deposition – see the Dinosaur National Monument summary in this report for more details).





ARCH Figure 8. Dinosaur track from the Moab Member of the Curtis Formation (VLS).
ARCH Figure 9. Two dinosaur tracks from the Moab Member (VLS).

Morrison Formation (Upper Jurassic)

The Morrison Formation is a heterogeneous and variegated unit dated to the middle Late Jurassic, between about 155 and 148 Ma (Foster 2007). The formation is divisible into several members. From oldest to youngest, the Tidwell, Salt Wash, and Brushy Basin members are reported from ARCH (although see the Summerville Formation section below for discussion of alternative interpretations of the Tidwell Member). The Morrison Formation outcrops near the northeastern, northwestern, and southwestern boundaries of ARCH, and on the north side of the lower Salt Valley. The Tidwell Member is mapped with the underlying Summerville Formation. It is

composed of siltstone with discontinuous beds of limestone; the limestone beds are most abundant near the base and top of the member. It is 8 to 15 m (25 to 50 ft) thick, and is interpreted as fluvial overbank and lake deposits. The Tidwell Member grades into and in some cases intertongues with the overlying Salt Wash Member. The Salt Wash Member is about 60 to 75% red and green muddy siltstone, and 25 to 40% ledge- and bench-forming sandstone lenses. It is interpreted as representing an anastomosing braided river system, with sandstones as channels and siltstones as overbank deposits. Within ARCH, the member is 40 to 90 m (130 to 300 ft) thick, averaging 55 m (180 ft). The overlying Brushy Basin Member is about 75% silty and clayey mudstone, with some sandstone and conglomerate lenses mostly found near the base. The member forms steep colorful slopes. Deposition occurred on a floodplain and in the Lake T'oo'dichi' basin; bright green mudstones are thought to represent lake deposits, while maroon, variegated and banded deposits are interpreted as terrestrial. Much of the mudstone probably represents decomposed fine volcanic debris. The Brushy Basin Member is 90 to 135 m (300 to 450 ft) thick at ARCH. Its upper contact with the Cedar Mountain Formation is a disconformity (Doelling 2010). The climate of the Morrison Formation is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004).

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Brachiosaurus*, *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks (Turner and Peterson 1999; Foster 2007). The formation is extensively fossiliferous at ARCH, with fossils known from the Salt Wash and Brushy Basin members within the park. Fifteen sites were reported for the Salt Wash Member by Swanson et al. (2005), yielding silicified wood and wood fragments, unionid bivalves, the invertebrate trace fossil *Scoyenia*, other invertebrate traces, dinosaur bones, and possible vertebrate traces. Cavities can be found in this member that correspond to now-lost bones (Swanson et al. 2005). Doelling (2010) noted the presence of dinosaur bone and petrified wood fragments in the Salt Wash Member of the ARCH area. Nineteen sites were reported for the Brushy Basin Member at ARCH, yielding silicified wood, *Scoyenia*, other invertebrate traces, and dinosaur bones (particularly those of sauropods, but also possibly including bones of ankylosaurs, which are rare in the Morrison Formation) (Swanson et al. 2005). Bones of the iconic sauropod *Apatosaurus* were found near the top of the member in 2000 (ARCH Figure 10) (Swanson et al. 2005), by the joint NPS/Utah Geological Survey/Utah Friends of Paleontology survey. This specimen was described by Foster (2005), and may be the youngest known Morrison Formation dinosaur from the Colorado Plateau (Santucci and Kirkland 2010). Two sites with bone from unspecified Morrison Formation units have also been reported from ARCH (Swanson et al. 2005). Hasiotis (1996a) briefly reported the presence of invertebrate traces from the Morrison Formation in and around ARCH. Hasiotis (1996b) reported that termite nests were present in the formation in and round the park, and Hasiotis and Demko (1996) found ant nests in and around the park as well. Additionally, Engelmann (1999) and Engelmann and Fiorillo (2000) drew on information from ARCH's Morrison Formation fossil sites.



ARCH Figure 10. Apatosaurus *caudal vertebrae at ARCH (photo courtesy John Foster, Museum of Western Colorado).*

A number of fossils have also been found in the Morrison Formation in the immediate vicinity of ARCH. Large petrified logs have been found at the contact with the Summerville Formation west of ARCH (Santucci and Kirkland 2010). A species of fern, *Osmundacaulis wadei*, was named from stems found in the Little Valley north of Moab (Tidwell and Rushforth 1970). Engelmann and Hasiotis (1999) reported on deep possible sauropod tracks from near Dalton Wells about 5 km (3 miles) west of ARCH, near the top of the lower Brushy Basin Member. Another horizon below the deep tracks preserves root traces, termite nests, and other trace fossils (Engelmann and Hasiotis 1999). Smith and Lisak (2001) described an *Allosaurus* skull with exaggerated nasal ornamentation from the Salt Valley outside of the park. Dayvault and Hatch (2003) reported on the presence of unusual plant fossils believed to represent new

growth from conifers and other plants, usually found in the lower Brushy Basin Member; these are common in the Moab area. Cycads and bennettitales (cycadeoids) are known from the Brushy Basin Member of the ARCH–Moab area as well (Dayvault and Hatch 2005). Foster and Lockley (2006) described several dinosaur track localities from the Moab area, including one on Copper Ridge near Highway 191 with limping theropod tracks, and another near ARCH and the Moab airport with possible stegosaur tracks.

The Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites, fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans, ostracodes, and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods, gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006); however, some of the traces, such as those attributed to anthozoans, brachiopods, horseshoe crabs, and echinoderms are only known from marine-influenced units that are not found at CURE (Hasiotis 2004). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians, lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods, diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

Cedar Mountain Formation (Lower Cretaceous)

The Cedar Mountain Formation is similar to the underlying Brushy Basin Member of the Morrison Formation, but not as brightly colored and with less clay. The base of the formation is marked by a cliff-forming sandstone, with the section above mostly composed of silty mudstone with sandstone intervals. It outcrops in the Cache and Salt valleys of ARCH. It can be up to 75 m (250 ft) thick in the area, but is usually 30 to 45 m (100 to 150 ft) thick. The upper contact with the Dakota Formation is a disconformity. Generally, the depositional setting is thought to have been similar to the underlying Brushy Basin Member (Doelling 2010). The climate became more humid over time, with high temperatures (Garrison et al. 2007). The Cedar Mountain Formation has been divided into several members. Of these, the Yellow Cat, Poison Strip, and Ruby Ranch members are present at ARCH (Swanson et al. 2005). There may be a higher member equivalent to the Mussentuchit Member (present to the west; see the CARE summary) in the area as well (Kirkland et al. 1997); a femur possibly from a hadrosaurid (duckbilled dinosaur) was recovered from a site in these rocks about 1.6 km (1 mile) west of ARCH (Galton and Jensen 1978; Kirkland et al. 1997).

The Cedar Mountain Formation is the lateral equivalent of the Burro Canyon Formation, as found at Colorado National Monument (COLM) and HOVE. Generally, the Burro Canyon Formation is present east of the Colorado River, and the Cedar Mountain Formation is present west of it (Dickinson and Gehrels 2008). The two intertongue in places (Dayvault and Hatch 2005).

The dating of the Cedar Mountain Formation is complex. The upper contact is established at approximately 97 Ma (Britt et al. 2009), but the internal dating of the various members and the lower contact are uncertain. For example, the Mussentuchit Member was thought to have begun deposition at approximately 98.37 ± 0.07 Ma (Kirkland and Madsen 2007). Then several quarries were found to be placed incorrectly in the Ruby Ranch Member and were actually from the Mussentuchit Member, including the Price River Two Quarry, which was dated to between 116 and 109 Ma by detrital zircons (Carpenter et al. 2008; although note that Kirkland and Madsen [2007] suggest that this age is too old and derives from older reworked zircons). At the base of the formation, the uppermost Yellow Cat Member dates to 124 Ma, in the early Aptian stage of the Early Cretaceous, but the age of the base of the member is uncertain; McDonald et al. (2010) suggested an early Barremian age (approximately 130 Ma) or older. Additionally, there are reports of basal Cretaceous ostracodes in the Yellow Cat Member of the San Rafael Swell (Sames et al. 2010). Internal stratigraphy can

also be problematic, as several fossiliferous quarries have been transferred from one member to another (Kirkland and Madsen 2007; Carpenter et al. 2008). Together, these suggest that the basal age and stratigraphic relationships of the formation are still incompletely known.

For many years, the fossils of the Cedar Mountain Formation were poorly known. This began to change with a series of publications on fossil plants in the late 1970s and early 1980s (Tidwell et al. 1976; Thayn et al. 1983, 1985; Thayn and Tidwell 1984; Tschudy et al. 1984; Tidwell and Thayn 1985). Then, during the later 1980s and 1990s, rich vertebrate localities were found at several levels in the formation, and it became an important stratigraphic unit for the study of Cretaceous vertebrates, fitting between the well-studied Morrison Formation of the Upper Jurassic and various Upper Cretaceous formations of the American West. Numerous fossiliferous sites have been found near ARCH; Kirkland and Madsen (2007) provide a useful summary. Swanson et al. (2005) reported fossil localities within ARCH from all three members present at the park. There are also two sites within ARCH in the Cedar Mountain Formation with fossils from unspecified members: a trace fossil locality possibly from the formation (L-GFE-94-12); and a site with dinosaur bone fragments (L-GFE-94-17) (Swanson et al. 2005). The tree fern *Tempskya* has been reported from the Moab area (Tidwell and Hebbert 1992). Doelling (2010) noted that isolated invertebrate fossils, such as gastropods and ostracodes, may be found in this formation in the area. Finally, a locality near ARCH, described by Shoemaker and Newman (1959) and Stewart et al. (1972a) as having marine fossils from the Sewemup Member of the Moenkopi Formation (see below), has been shown to be a misidentified Cedar Mountain Formation site. The locality, a little over 3 km (2 miles) northwest of ARCH in Salt Valley, has yielded freshwater invertebrates (charophytes, bivalves, gastropods, and ostracodes) (Lucas et al. 1997a). Swanson et al. (2005) misidentified this site as being within ARCH.

Cedar Mountain Formation (Lower Cretaceous): Yellow Cat Member

The Yellow Cat Member is composed of interbedded mudstone and some sandstone and limestone (Kirkland et al. 1997, 1999). It is similar to the Morrison Formation, and part of it had been included in the Brushy Basin Member. However, it is drabber and less variegated in color (Kirkland et al. 1999). It is interpreted as having had a semi-arid monsoonal climate, like the Morrison Formation but wetter (Kirkland et al. 1997). The general depositional settings were alluvial and lacustrine (Eberth et al. 2006; Britt et al. 2009). The upper contact with the Poison Strip Member has been described as a disconformity, but the two intertongue in at least some areas (Eberth et al. 2006). In the Green River area 40 km (25 miles) northwest of ARCH, there is a lower part to the Yellow Cat Member that is not present at ARCH or its immediate vicinity (Kirkland and Madsen 2007). It is best known from the Crystal Geyser Quarry southeast of Green River, which has yielded abundant fossils of the unusual theropod *Falcarius*, as well as an unnamed ankylosaur (Kirkland et al. 2005a).

The Yellow Cat Member has one published fossil locality within ARCH: site A-8, which has sauropod bones (Swanson et al. 2005). Doelling (2010) reported that white petrified wood is locally abundant in this unit in the ARCH area. Several notable sites have been found in this member just outside of ARCH. Among the best known are the Dalton Wells Quarry and the Gaston Quarry. The Dalton Wells Quarry is about 5 km (3 miles) west of ARCH (Galton and Jensen 1978). It is one of the best Early Cretaceous dinosaur sites in the world (Eberth et al. 2006). Fossils from the Dalton Wells Quarry were first described by Galton and Jensen (1978), who named *Iguanodon ottingeri* from a jaw fragment with teeth which had been collected in 1968 by Lin Ottinger. The quarry has been extensively described by Eberth et al. (2006) and Britt et al. (2009). It is composed of four stacked bonebeds which represent debris flows. The flows occurred in the open air and probably did not transport material over long distances. The bonebeds accumulated in an area of lakes, ponds, and wetlands. Over time, large lakes formed near the bonebeds, and eventually a lake was present over the site. The animals may have been victims of drought-related mortality (Eberth et al. 2006). Approximately 4,200 bones have been recovered from this site, representing about 5% of the total bones present. Almost all of the fossils represent dinosaurs. At least 67 individuals of eight dinosaur genera have been recovered from the quarry: small theropod *Nedcolbertia* (at least two); large dromaeosaurid *Utahraptor* (at least nine); a possible ornithomimid (ostrich mimic; at least two); an unnamed *Camarasaurus*-like sauropod (at least fifteen); sauropod *Venenosaurus* (at least three); indeterminate sauropods (at least twenty-one); polacanthine ankylosaur (armored dinosaur) *Gastonia* (at least nine); and an unnamed large iguanodont (at least four). Most of the dinosaurs were subadults. Other vertebrates are represented by eight bones from turtles, choristoderes (extinct aquatic reptiles), crocodiles or crocodile relatives, and a possible pterosaur. Fossil wood and invertebrate burrows are also known from the site, and many of the bones have been modified by insects. Paleozoic marine fossils have been found in reworked chert pieces at the site, and include bryozoans, brachiopods, and crinoids (Britt et al. 2009). Theropod and sauropod tracks are known from the vicinity (Lockley et al. 1999).

North of ARCH the Gaston Quarry has yielded *Utahraptor* (Kirkland et al. 1997), *Gastonia* (Kirkland 1998), and an iguanodont (Kirkland et al. 1997), and dinosaur tracks were found nearby (Lockley et al. 1999; Kirkland et al. 2005b). The specimens for which *Utahraptor* and *Gastonia* were named came from the Gaston Quarry (Kirkland et al. 2005b). Another quarry, the Stikes Quarry, was found north of ARCH in 2004. It has yielded theropod and iguanodont bones, including possible skull bones of *Utahraptor* (Kirkland et al. 2011). There is also a unique trace fossil site near ARCH. Bird tracks were found in the uppermost part of the Yellow Cat Member just outside of the

park in 2005 (Wright et al. 2007); at the time of discovery, they were the oldest bird tracks known from North America (Wright et al. 2006, 2007).

Overall, the non-vertebrate fossil assemblage of the Yellow Cat Member exclusive of the lower unit found at Green River includes microbial limestones (Kirkland et al. 1999), palynomorphs (Sames et al. 2010), charophytes (Eberth et al. 2006; Sames et al. 2010), petrified wood (Doelling 2010), root casts (McDonald et al. 2010), gastropods (Kirkland et al. 1999), ostracodes (Sames et al. 2010), and invertebrate trace fossils (McDonald et al. 2010). Reworked Paleozoic marine fossils may also be found, as mentioned above (Britt et al. 2009). Non-dinosaurian vertebrates include hybodont sharks, ganoid fish (Kirkland et al. 1998), bowfins, semionotid fish, lungfish, turtles, the sphenodont *Toxolophosaurus* (Kirkland et al. 1997, 1999), choristoderes (Britt et al. 2006), crocodiles or crocodile relatives (Kirkland et al. 1997, 1999), possible pterosaurs (Britt et al. 2009), mammals (McDonald et al. 2010), and shark coprolites (Kirkland et al. 1998). The dinosaurian assemblage is by far the best-known paleontological aspect of the Yellow Cat Member. Represented are several theropods (*Nedcolbertia*, *Utahraptor*, a possible ornithomimid, an unnamed large theropod, and at least one other taxon) (Kirkland et al. 2005b), several sauropods (an unnamed *Camarasaurus*-like form, the brachiosaurid *Cedarosaurus*, an unnamed titanosaur [Kirkland et al. 2005b], and *Venenosaurus* [Britt et al. 2009]), the armored dinosaur *Gastonia* (Kirkland et al. 2005b), a hypsilophodont-grade ornithopod (McDonald et al. 2010), several iguanodonts (at least *Cedrorestes* [Gilpin et al. 2007], *Hippodraco* [McDonald et al. 2010], *Iguanodon ottingeri*, and an unnamed large sail-backed form from the Dalton Wells Quarry that may be the same as the latter [Kirkland et al. 1997, 1999]), and tracks of theropods, sauropods, possible ornithopods (Lockley et al. 1999), and birds (Wright et al. 2007). This dinosaur assemblage is similar to dinosaur assemblages of similar age in Europe (Kirkland et al. 1997).

Cedar Mountain Formation (Lower Cretaceous): Poison Strip Member

The Poison Strip Member is a cliff-forming sandstone unit interpreted as a meandering river complex (Kirkland et al. 1997, 1999). It was previously known as the "middle sandstone" (Young 1960; Kirkland et al. 2005b). Accelerating deposition caused stacking of channel deposits (Eberth et al. 2006). The Poison Strip Member has one published fossil locality within ARCH: site A-7, which has yielded a large log and a smaller log, the latter possibly burned (Swanson et al. 2005). Cycads and bennettitales are known from this member near ARCH, including one locality just west of the park that has yielded a new species of the bennettitale *Monanthesia* (Dayvault and Hatch 2005). A petrified forest of cycads and conifers is known in this member east of the Gaston Quarry (Santucci and Kirkland 2010). A partial skeleton of an armored dinosaur was found about 5 km (3 miles) west of ARCH (Bodily 1969; Kirkland et al. 1997), and bones of another were found 8 km (5 miles) to the southwest of this site (Bodily 1969). The identification of this dinosaur has changed over the years; the most recent identification is as a large unnamed polacanthine (Carpenter et al. 2008).

The fossil assemblage of the Poison Strip Member is not yet as well-known or diverse as that of the Yellow Cat Member. Non-vertebrate fossils include palynomorphs (Eberth et al. 2006), root traces (DiCroce and Carpenter 2001), cycads, cycadeoids, conifers (Kirkland et al. 1999), and invertebrate trace fossils (McDonald et al. 2010). Vertebrates represented include theropods (DiCroce and Carpenter 2001), the sauropod *Venenosaurus* (Tidwell et al. 2001), a large unnamed polacanthine (Carpenter et al. 2008), and the ornithopod *Planicoxa* (DiCroce and Carpenter 2001). *Venenosaurus* and *Planicoxa* were both described from Tony's Bone Bed, about 1.6 km (1 mile) east of the Gaston Quarry (Kirkland et al. 2005b). Dinosaurs from this member are similar to those of the Yellow Cat Member (Kirkland and Madsen 2007).

Cedar Mountain Formation (Lower Cretaceous): Ruby Ranch Member

The Ruby Ranch Member is mostly composed of mudstone, with minor sandstone and limestone (Kirkland et al. 1999). It is notable for its abundant carbonate nodules, representing paleosols and ephemeral ponds. The surrounding sandstones were deposited by low-sinuosity rivers (Kirkland and Madsen 2007). The prevailing climate of the member is thought to have been semi-arid and monsoonal, like the Morrison Formation (Kirkland et al. 1999).

The Ruby Ranch Member is fossiliferous within ARCH (ARCH Figures 11, 12, and 13). An important track site was discovered in this member in the park (Lockley et al. 2004), during the 2000 survey. The tracks were found in sandstone beds in the lower Ruby Ranch Member, south of Delicate Arch and north of its associated parking lot. There are two major track assemblages at this site, and a third level above them with indistinct tracks. The lower assemblage has two types of theropod tracks and faint sauropod tracks, as well as unusual markings (Lockley et al. 2004) (ARCH Figure 13). Wright et al. (2007) suggested that the unusual markings could be invertebrate traces, pterosaur feeding traces, or inorganic markings. The upper assemblage is a trampled layer with numerous tracks that can be assigned to seven distinct track makers: three types of three-toed theropod tracks, including gracile and robust forms; two-toed theropod tracks that are of the right size to belong to *Utahraptor*; sauropod tracks; probable ornithopod tracks; and probable ankylosaurian tracks (Lockley et al. 2004). Swanson et al. (2005) reported site A-4 (apparently the lower assemblage), which yielded stromatolite-like traces, possible pterosaur traces, and trackways from small theropods. They also mentioned the presence of other trace fossils, although it is not clear if these are all from ARCH, as at least one (the ornithopod track from Lockley et al. [1999]) is from outside the park (near the type locality for the member).







ARCH Figure 11. Dinosaur tracks from the Ruby Ranch Member of the Cedar Mountain Formation (VLS).
ARCH Figure 12. Dromaeosaur-like track from the Ruby Ranch Member (VLS).
ARCH Figure 13. Possible feeding traces from the Ruby Ranch Member (VLS).

Overall, the Ruby Ranch Member's fossil assemblage includes pollen (Kirkland et al. 1999), root traces, arthropod trace fossils (Tremain 2003), hybodont sharks (Kirkland et al. 1999), the sphenodont *Toxolophosaurus*, crocodylomorphs, several types of theropods (an unnamed carnosaur, *Acrocanthosaurus*, and dromaeosaurids), brachiosaurid sauropods, nodosaurids, a new species of the polacanthid *Gastonia*, the long-tailed ornithopod *Tenontosaurus* (mostly known as a prey item of *Deinonychus*) (Kirkland and Madsen 2007), and tracks of theropods, sauropods, ankylosaurians, and ornithopods (Lockley et al. 2004). Additionally, a site in Utah near the border with Colorado (the Hotel Mesa Quarry) has yielded fossils of sharks, bony fish, lungfish, various crocodile relatives, theropods, the sauropod *Brontomerus*, and ornithopods (Taylor et al. 2011). It has been attributed to both the Burro Canyon Formation (Kirkland and Madsen 2007) and the Ruby Ranch Member of the Cedar Mountain Formation; by geography, it is within the Burro Canyon Formation (Taylor et al. 2011). The dinosaurs of the Ruby Ranch Member are similar to those of the Arundel Formation of Maryland, Cloverly Formation of Montana and Wyoming, and Trinity Group of Oklahoma and Texas (Kirkland et al. 1999). However, with the reassessment of quarry stratigraphic position it is not clear how many of these fossils are actually from the member. There is at least a large unnamed nodosaurid present in the middle of the unit (Carpenter et al. 2008), and the unnamed carnosaur and brachiosaurid ("*Astrodon*") are known from the Long Walk Quarry low in the member (Kirkland and Madsen 2007).

Dakota Formation (Lower–Upper Cretaceous)

The Dakota Formation (also Sandstone or Group) is a heterogeneous rock unit with outcrops as distant as Minnesota; because it was named in Nebraska from rocks on the opposite

shore of the Western Interior Seaway, its use in areas on the western side of the seaway may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). Within ARCH, it is only exposed in Cache and Salt valleys, where it forms an orange- or brown-colored ridge. The thickness of the formation within the park is typically 6 to 25 m (20 to 80 ft). Composition ranges from conglomeratic sandstone, to claystone and shale, to limestone; the most prominent lithology in the park is conglomeratic sandstone (Doelling 2010). To the east, the Dakota Formation includes a basal conglomerate, followed by a coaly mudstone interval, and finally a sandstone interval at the top of the unit (Doelling 2010). This conglomerate-coaly shale-sandstone division is common for the formation throughout the Colorado Plateau. These divisions reflect environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma (see Titus et al. [2005] for discussion of the age). It grades into the overlying Mancos Shale over a narrow interval at ARCH (Doelling 2010).

The Dakota Formation is fossiliferous within ARCH. The 2000 paleontological survey identified one fossiliferous Dakota Formation locality within the park, with plant or woody tissue (Site A-6; Swanson et al. 2005). Over its entire depositional area, including outside of ARCH, the fossils of the Dakota Formation are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. The best-known terrestrial fossils are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Formation (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Formation of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Formation for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). The floral assemblage in Utah includes coal (Molenaar and Cobban 1991), palynomorphs (including moss spores, fern spores, and gymnosperm and angiosperm pollen; Nichols 1997), the tree fern *Tempskya* (Ash and Read 1976), *Equisetum*, possible gymnosperms, and angiosperms (Tidwell et al. 1976). Ferns have been recovered from this formation in Westwater, 45 km (27 miles) northeast of ARCH (Tidwell et al. 1967; Rushforth 1969). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Formation rocks in Utah reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals; locations included areas near Bryce Canyon National Park (BRCA; see the BRCA summary in this report for more details) (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Formation marine fossils include algae, foraminifera (Tibert et al. 2003), dinoflagellate cysts (a type of plankton) (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Formation may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera, corals, bryozoans, brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Mancos Shale (Upper Cretaceous)

The Mancos Shale is a widespread unit that marks the advances and retreats of the Western Interior Seaway. The formation is divisible into a number of members depending on location. At ARCH, it is divided from oldest to youngest into the Tununk, Ferron Sandstone, and Blue Gate members. It is primarily a gray slope-forming shale unit in the park, with some sandstone ledges. Outcrops are found in the Cache and Salt valleys (Doelling 2010). As described in the Henry Mountains to the southwest, the Mancos Shale includes parts of four transgressive-regressive cycles. The Tununk and Ferron members represent most of the cycle began during the deposition of the Dakota Sandstone, the Blue Gate Shale includes two cycles, and the final cycle is present in rocks not found at ARCH (Peterson et al. 1980).

The Tununk Member of ARCH is a shale unit between 30 and 120 m (100 and 400 ft) thick. It grades into the overlying Ferron Sandstone Member (Doelling 2010). An altered ash layer present in the Tununk Member has been dated to 93.46 ± 0.6 Ma and 93.25 ± 0.55 Ma (Kirkland and Madsen 2007). Volcanic ash may have provided nutrients for organisms (Bagshaw 1977). Water depth during deposition in the area was 90 to 180 m (300 to 600 ft), with normal marine conditions (Peterson et al. 1980).

Deposition occurred in advance of deltas, with several minor sea-level fluctuations (Sethi and Leithold 1997).

The fossils of the Tununk Member are largely derived from marine mollusks. It has yielded coccoliths (microscopic calcium carbonate plates from algae) from fecal pellets (Leithold 1993), foraminifera, bivalves, ammonites (Peterson et al. 1980), gastropods (Santucci and Kirkland 2010), ostracodes (Peterson et al. 1980), burrows (Molenaar and Cobban 1991), shark teeth (Santucci and Kirkland 2010), and scales (Morris et al. 2010) and teeth of bony fish (Bagshaw 1977). This member is fossiliferous at ARCH: the May 2000 paleontological survey reported a locality (Site A-5, identified as the lower member of the Mancos Shale) with fossils of the bivalve *Pycnodonte newberryi unbonata* (Swanson et al. 2005). *Pycnodonte* shells are abundant in the basal Tununk Member at ARCH (Santucci and Kirkland 2010). The interval just above the Dakota Formation is known for fossil-rich beds in the ARCH area (Doelling 2010).

The Ferron Sandstone Member of ARCH is a brown-gray ledge-forming sandstone unit between 25 and 35 m (90 and 120 ft) thick. It is interpreted as having been deposited under a shallow shoaling sea (Doelling 2010). It is a sandstone tongue into the marine Mancos Shale (Dickinson and Gehrels 2008), representing the culmination of a regressive phase (Molenaar and Cobban 1991). At some locations it is divisible into smaller units (Cotter 1975). A variety of deltaic and coastal depositional settings are known. Two large deltas were present in central Utah: the northern Vernal Delta and the southern Last Chance Delta (Cotter 1975). Five principal deltaic cycles are known from the unit, as well as multiple ash beds (Crowley et al. 1994). To the north in the Uinta Basin, the Ferron Sandstone dates to approximately 89.7 Ma (Molenaar and Cobban 1991). Its importance to the fossil fuel community has made it exceptionally well-known (Morris et al. 2010).

The fossils of the Ferron Sandstone include both marine and terrestrial taxa. The assemblage includes organic debris (Cotter 1975), coal, palynomorphs (representing acritarchs [miscellaneous organic microfossils], dinoflagellates, green algae, sphagnum mosses, ferns, gymnosperms, and angiosperms) (Crowley et al. 1994), plant fragments, seed cones, leaves (Ursek 1979), ferns, gymnosperms, angiosperms (Young 1987), angiosperm logs (Ursek 1979), petrified wood (Smith et al. 1963), root traces (Ryer 1981), bryozoans (Hill 1982), bivalves, ammonites (Molenaar and Cobban 1991), gastropods (Ryer 1981), echinoids (Hill 1982), marine invertebrate burrows and trails (Cotter 1973), possible cephalopod coprolites (Ursek 1979), sharks, rays (Becker et al. 2010), bony fish (Becker and Linteris 2009), and ornithopod dinosaur tracks (Jones 2001). Marine fossils are found throughout the member; coal and plant fragments appear in the upper part (Peterson et al. 1980). Fossiliferous zones are present at the top of this member in the vicinity of ARCH (Doelling 2010), and bivalves, ammonites, gastropods, and shark teeth have been recovered from the area, including bivalves and ammonites from near the airport (Doelling 1988).

The Blue Gate Shale Member of ARCH is a shaly unit less than 150 m (500 ft) thick, generally similar to the Tununk Member (Doelling 2010). Another open marine unit, it is interpreted as having been deposited under depths of 60 to 120 m (200 to 400 ft) (Peterson et al. 1980). Fossils of the Blue Gate Shale mostly come from marine invertebrates. The unit has yielded algal biscuits (lumps of calcium carbonate produced by algae) (Cross 1976), bivalves, ammonite and belemnite cephalopods, gastropods (Eaton 1990), trace fossils (Riemersma and Chan 1991), and shark teeth (Santucci and Kirkland 2010).

Cenozoic remnants (Eocene, at least in part)

Isolated sections of Cenozoic rocks have been found in Salt Valley (J. Kirkland, pers. comm., October 2011). These rocks include gastropods typical of the Green River Formation better known to the north, in northeastern Utah, southwestern Wyoming, and northwestern Colorado (see the Fossil Butte National Monument [FOBU] summary in this document for an extended discussion of the Green River Formation). The rocks were first discussed as gravel deposits, by Dyer (1983), Dyer et al. (1983), and Oviatt (1988). These authors noted the presence of bivalve and gastropod fossils in the eroded limestone and chert clasts. Oviatt (1988) proposed that the gravels represent now-eroded lake-margin facies of the Eocene-age Green River Formation.

Quaternary sediments (Pleistocene–Holocene)

The Quaternary deposits of ARCH are a combination of stream alluvium, terrace gravels from former drainages, eolian sand, and landslide deposits. Grain sizes range from gravel to clay. Deposits range in thickness; most alluvial and eolian deposits are less than 7.6 m (25 ft) thick, but can be much thicker, for example in the northern Moab Valley and areas where salt is undergoing dissolution at depth, leading to subsidence (Doelling 2010). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). ARCH has a variety of Quaternary fossil resources, including examples of both types mentioned above. Summaries of Quaternary fossil sites in southern Utah and the vicinity include Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008).

The best-known Quaternary fossil site of ARCH is Bison Alcove. Bison Alcove is an alcove site from an elevation of 1,317 m (4,321 ft), found at the contact of the Carmel Formation and the Entrada Sandstone. The site has remains of large and small mammals, including two individuals of *Ovis* (bighorn sheep) and the namesake bison (Mead et al. 1991; Sharpe 1993), and abundant packrat middens (Sharpe 1993). Megafaunal

remains date to 405 ± 60 radiocarbon years before present (530–320 calendar years before present) and 355 ± 60 radiocarbon years before present (510–300 calendar years before present) (Sharpe 1993). The bison remains may derive from packrats taking small bones from the carcass of an animal killed or scavenged nearby by humans. Humans themselves have also been in the alcove, leaving behind an artifact composed of a hornsheath and twig joined by resin (Mead et al. 1991). The alcove has also been host to raptor roosts (Sharpe 1993).

Packrat middens from Bison Alcove have also been published. Packrat middens are collections of plant material and food waste constructed by packrats (*Neotoma* spp.) and cemented by their viscous urine. They can be well-preserved in dry caves, and illustrate the environment of the builder's foraging range. Middens are important tools for reconstructing the ecology and climate of the late Pleistocene and Holocene of the southwestern United States (Strickland et al. 2001). Dated Bison Alcove middens range in age from at least 20,050 ± 160 radiocarbon years before present (24,390–23,490 calendar years before present) to modern (Smith and Betancourt 1998). The primary references are Sharpe (1991, 1993) and Smith and Betancourt (1998); information about these middens is also included in Mead et al. (1991) and the broader reviews Anderson et al. (2000) and Thompson and Anderson (2000). The online U.S. Geological Survey/National Oceanic and Atmospheric Administration Packrat Midden Database (http://esp.cr.usgs.gov/data/midden/) has entries for 17 middens from Bison Alcove.

Sharpe (1993) analyzed 11 midden samples, producing dates on conifer needles, packrat dung, and keratin from the megafauna. Plant remains from the middens suggest that more mesophytic vegetation dominated the Bison Alcove area between 20,000 and 12,500 radiocarbon years before present, with *Pinus flexilis* (limber pine) and *Pseudotsuga menziesii* (Douglas fir) abundant. These two taxa are no longer growing at ARCH. They were probably widely dispersed among a scrubby understory. Bison Alcove full glacial and late glacial taxa are also found at Cowboy Cave near CANY (see below). Modern vegetation was present near Bison Alcove before 2,660 radiocarbon years before present, with conditions possibly slightly more moist than present (Sharpe 1993). Smith and Betancourt (1998) used packrat dung from various sites, including 13 Bison Alcove middens, as a proxy for the size of the packrats that produced them, to examine how the animals responded to temperature change. They found that the rodents were larger during the Late Pleistocene and declined to modern sizes rapidly, by the early Holocene, as the climate warmed (Smith and Betancourt 1998).

Another alcove in ARCH was the source of mammoth fossils (Santucci et al. 2001). At this alcove, in the area of Lower Courthouse Wash, a partial mandible of a Columbian mammoth (*Mammuthus columbi*) has been collected (ARCH Figure 14). Bone fragments were still at the site as of 2005 (Swanson et al. 2005). The specimen was found in the 1950s by ARCH Chief Ranger Lloyd Pierson, and is informally known as "Woody" (Santucci and Kirkland 2010). It may be the same as a mammoth cited by Agenbroad and Mead (1989) as from "Arches". This specimen is dated to 17,200 ± 800 radiocarbon years before present (22,340–18,830 calendar years before present) (Agenbroad and Mead 1989). Santucci et al. (2001) presented an overview and cited selected examples of National Park Service fossils found in caves, including at ARCH.



ARCH Figure 14. Mammoth mandible from a cave in Lower Courthouse Wash (VLS).

Other Quaternary fossils from ARCH include diatoms, poor root traces, and ostracodes found in marly beds in Salt Valley; these beds are associated with the dated Middle Pleistocene ashes mentioned earlier (Oviatt 1988). They were first thought to be possibly of Late Miocene age (Dyer et al. 1983), but their association with the volcanic material places them in the Pleistocene (Oviatt 1988).

Similar Quaternary fossils are found at a handful of sites within about 100 km (60 miles) of ARCH. Packrat middens have been described from a few sites at CANY (Elias et al. 1992; Anderson et al. 2000; Coats et al. 2008). CANY also has yielded pollen (Reheis et al. 2005) and a variety of small fossils used for sediment dating, such as charcoal, seeds, and wood (Biggar and Adams 1987). Cowboy Cave, approximately 60 km (37 miles) southwest of ARCH and just west of CANY, is one of the best-known Pleistocene–Holocene sites in the region. The contents of the cave were extensively described by Jennings (1980). It is notable for having both extensive Pleistocene material and four distinct Holocene cultural layers above, spanning at least the last 13,000 years (humans present only since approximately 9,000 years ago). In the Pleistocene layer, various types of mammal dung and hair were found with packrat middens and a handful of bones, mostly bison, but also two juvenile mammoth tusks and a possible elk bone (Barnett and Coulam 1980; Hansen 1980; Jennings 1980; Spaulding and Peterson 1980). The cultural levels show human use of stone, bones, hair, hides, shells, plant parts, and other local organic materials to make tools,

ornaments, bags, and other artifacts (Hewitt 1980; Hull 1980a, 1980b; Janetski 1980; Lucius 1980; Weder 1980). Prehistoric dung at the cave mostly came from bison, but also includes feces from Shasta ground sloths (*Nothrotheriops shastensis*), elephants, horses, elk, elk or camel (Mead and Agenbroad 1992), rabbits, rodents (Hansen 1980), and humans (Hogan 1980).

COLM, over 72 km (45 miles) northeast of ARCH, has a diverse Quaternary fossil record that is of interest for comparative purposes. In 1965 a mastodon tooth was reportedly found in a canyon there, but has been lost (Scott et al. 2001a). Bone and plant specimens have been collected from a fissure; nine vertebrate taxa were tentatively identified, including eagle, wolf, and bighorn sheep (Trujillo et al. 2004). Middens are known from the park (Anderson et al. 2000; Scott et al. 2001b; Trujillo et al. 2004). COLM also has charcoal in valley fill that span almost the entire Holocene in age (Scott et al. 2001b).

Finally, there are a few isolated reports of mammoths, from Fruita near COLM, Bartlett Wash less than 5 km (3 miles) west of ARCH, and Professor Valley east of ARCH (Agenbroad and Mead 1989).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within ARCH. However, they are known to preserve fossils elsewhere, and future field investigations within the park may recover fossils from one or more of them.

Cutler Formation (Permian)

The Cutler Formation is mapped as being present in a narrow area of southwestern ARCH (Doelling 1985, 2010). Doelling et al. (1994) mapped it as largely an unnamed lower Cutler Formation member, with possibly some rocks of an unnamed upper Cutler Formation member present in the extreme southwest. The lower Cutler Formation member of the ARCH area is a red and maroon cliff-forming sandstone and conglomerate unit with some sandy mudstone and thin gray fossiliferous limestone beds. It is between 55 and 65 m (180 and 220 ft) thick, and is described as representing part of the Elephant Canyon Formation interval (Doelling et al. 1994) or what had been called the Rico Formation (Doelling [2010]; the lithological issues were briefly discussed above). The upper member is similar to the lower member, but lacks limestone beds. The sandstone beds represent both fluvial and eolian deposition (Doelling et al. 1994). The Cutler Formation itself is raised to a group with multiple named members elsewhere, such as at CANY (Baars 2010), but these divisions are not present at ARCH. The upper contact with the Moenkopi Formation is a disconformity (Doelling 2010).

The Elephant Canyon Formation was deposited during the Late Pennsylvanian and Early Permian (Sanderson and Verville 1990), sometime between 305 and 280 Ma. The setting for this interval is thought to have been a coastal plain to shallow marine complex in an arid area (Terrell 1972; Condon 1997), similar to the modern Colorado River delta in the Gulf of California. There was a combination of dunes, channels, mudflats, tidal flats, supratidal areas, prodeltas, bays, bars, and open marine areas that formed in repetitive cycles as a given area shifted from marine to nonmarine and back (Terrell 1972). The Cutler Formation of the ARCH area is interpreted as alluvial fan deposits shed from the adjacent Uncompahgre Uplift (Doelling 1985, 1988).

The Cutler Formation is not known to be fossiliferous at ARCH, but there is some potential, given the presence of fossiliferous beds in the lower member in the vicinity. Fossils reported from the Elephant Canyon Formation in general include wood fragments, microbial mats, foraminifera, corals, bryozoans, brachiopods, bivalves, gastropods, cephalopods, trilobites, echinoderms, invertebrate burrows (Terrell 1972), and palaeoniscid fish (Sumida et al. 1999). They are most abundant in the limestones (Campbell 1987). Fossils reported from the former Rico Formation include root traces, leaf impressions, logs (Loope 1984), foraminifera, corals, bryozoans, brachiopods, bivalves, scaphopods (tusk shells), gastropods, cephalopods, ostracodes, trilobites, crinoids (McKnight 1940), and fish (Lohman 1974). Few fossils have been found in undivided Cutler Formation rocks. Wood is known in the general CANY area (Rigby et al. 1971); and fossil logs and leaves have been found just east of CANY in undivided rocks (Baker 1933). Otherwise, the nearest significant finds from undivided Cutler Group rocks come from the Placerville area of western Colorado, about 140 km (85 miles) southeast of ARCH (Baird 1965; Lewis and Vaughn 1965; Laurin 1994). Finds here include labyrinthodont amphibians, early amniotes and amniote relatives, pelycosaurs (“mammal-like reptiles”) (Lewis and Vaughn 1965), and footprints of labyrinthodonts and amniote relatives (Baird 1965).

San Rafael Group: Carmel Formation (Middle Jurassic)

The Carmel Formation is a marine and terrestrial Middle Jurassic unit. The transition between these general depositional settings is near ARCH: west of CANY, it is a marine unit, but it becomes terrestrial to the east (Blakey et al. 1983). This interval was initially mapped as the Carmel Formation (Dane 1935), but was renamed the Dewey Bridge Member of the Entrada Sandstone (Wright et al. 1962) because it does not contain limestone typical of the Carmel Formation in areas to the west.

While it is true that the interval is sandy, similar to the overlying Slick Rock Member of the Entrada Sandstone, it is lithologically distinct, and has been found to be stratigraphically continuous with the marine Carmel Formation. Therefore, it has been returned to the Carmel Formation (Doelling 2010). Many older references (Lohman 1975; Baars 1983; Doelling 1985, 1988) use the Dewey Bridge Member of the Entrada Sandstone for this interval. The Carmel Formation dates to the Middle Jurassic, with an age of approximately 169 to 162 Ma based on radiometric dates from central and southwestern Utah (Sprinkel et al. 2011).

Scattered outcrops of the Dewey Bridge Member of the Carmel Formation are found throughout ARCH. It is divisible into informal lower and upper parts (members of Doelling [2010]) in the park. The lower part is composed of yellow-gray resistant sandstone beds, with some pink to red-brown sandstone beds interbedded. It is 5 to 26 m (15 to 85 ft) thick in the ARCH area. The upper part is mostly composed of red-brown silty sandstone, and forms slopes between the more resistant lower part and the Entrada Sandstone. It is 18 to 48 m (60 to 157 ft) thick in the ARCH area. Bedding is commonly distorted or convoluted due to soft-sediment deformation, which also often affected the basal part of the overlying Entrada Sandstone. The Carmel–Entrada contact is sharp to irregular. The depositional setting of the Carmel Formation in the ARCH area is described as broad tidal flats with the Carmel Sea to the west (Doelling 2010).

The Carmel Formation is not known to be fossiliferous at ARCH (Swanson et al. 2005). Most fossils described from the formation as a whole are marine, but the Dewey Bridge Member of ARCH is terrestrial. Terrestrial fossils described from the Carmel Formation include spores, pollen (Sprinkel et al. 2011), and tracks of crocodylomorphs and dinosaurs (Lockley and Hunt 1995), from northeastern Utah (Santucci and Kirkland 2010).

San Rafael Group: Entrada Sandstone (Middle Jurassic)

The Entrada Sandstone is dominated by pale reddish-orange to reddish-brown sandstone, with some silt (Harshbarger et al. 1957). It is widely exposed in southwestern and northeastern ARCH (Doelling 2010). Previously, the Entrada Sandstone of ARCH was divided into three members. From oldest to youngest, these were the Dewey Bridge, Slick Rock, and Moab members (Wright et al. 1962; Lohman 1975; Baars 1983; Doelling 1985, 1988). More recently, the Dewey Bridge Member has been reassigned to the Carmel Formation, and the Moab Member (or Tongue) has been reassigned to the Curtis Formation, leaving only the Slick Rock Member in the Entrada Sandstone (Doelling 2010). The Slick Rock Member hosts many of the famous geologic features of ARCH. For example, the natural arches usually formed at its lower or upper contacts. It is a red-orange or brown sandstone, with both cross-bedding typical of eolian deposition and horizontal bedding. It is usually 60 to 105 m (200 to 350 ft) thick in the area, and is interpreted as representing a dune field in a back-beach. The upper contact with the Curtis Formation is disconformable (Doelling 2010). The Entrada Sandstone lacks age-diagnostic fossils and is regarded as late Middle Jurassic in age (Lucas et al. 1985). The climate was semi-arid to arid (Ekdale and Picard 1985) and hot, with sparse vegetation. The Entrada dunes made up the last major erg of the area, and formed when southern Utah was about 22° N (Loope 2008).

The Entrada Sandstone is not known to be fossiliferous at ARCH (Swanson et al. 2005), and it is very sparsely fossiliferous overall. Many reports of fossils from the Entrada Sandstone are from the Moab Member, now in the Curtis Formation. These include the well-known megatracksite of the Moab area (Lockley 1991; Lockley et al. 2007) and the crocodylomorph *Entradasuchus* (Hunt and Lockley 1995). Britt (1996) reported two dinosaur trace fossil sites in this formation within ARCH, but these are from the Moab Member as well (Swanson et al. 2005). Trace fossils have been reported from the Escalante Member in south-central Utah, including rare root traces, insect burrows (Loope 2008), vertebrate burrows, traces of sand-swimming vertebrates (Graverson et al. 2007), and theropod (large *Megalosauripus* and smaller *Therangospodus*) and sauropod (cf. *Brontopodus*) dinosaur tracks (Foster et al. 2000).

San Rafael Group: Summerville Formation (Upper Jurassic)

The Summerville Formation is thin at ARCH (about 2 to 6 m [6 to 20 ft] thick) and was mapped with the overlying Tidwell Member of the Morrison Formation by Doelling (2010). Historically, the Summerville–Tidwell interval of the ARCH area has been regarded as only the Summerville Formation (Dane 1935; Baars 1983) or only the Tidwell Member (Doelling 1985, 1988). The Tidwell Member is sometimes transferred to the Summerville Formation, making it the uppermost member (Lucas and Anderson 1997), but this has not been widely adopted. The Summerville Formation is composed of tan to brown ledge-ledge forming sandstone and slope-forming red sandy siltstone in the ARCH area, with a basal zone of reworked yellow-gray sandstone. The upper contact with the Tidwell Member is a disconformity. Most outcrops of the Summerville-Tidwell interval are found near the northeastern, northwestern, or southwestern boundaries, or along the north side of the lower Salt Valley (Doelling 2010).

In the past, the Summerville Formation was usually thought to have been deposited during the late Middle Jurassic (Steiner 2003), but like the underlying Curtis Formation, it is now known to date to the Oxfordian stage of the early Late Jurassic (Wilcox 2007; Hintze and Kowallis 2009). Deposition was affected by occasional flooding from the regressing Sundance Sea (Petersen and Pack 1982). Aridity decreased during deposition (Lockley and Wright 2003), as Pangaea continued to move north (Lockley and Mickelson 1997). At ARCH, the

Summerville Formation is regarded as representing marginal marine delta deposits (McKnight 1940; Doelling 2010).

The Summerville Formation does not appear to be fossiliferous at ARCH, unless the Moab Member megatracksite horizon is actually from the base of the Summerville Formation instead (Lockley 1991; Lockley and Mickelson 1997). Locally, dinosaur tracks can be found in the reworked basal sandstone zone (Lockley 1991; Doelling 2010); the presence of these tracks was noted by Swanson et al. (2005), but they did not cite any specific instances within ARCH. The Summerville Formation fossil assemblage is limited, and includes plant debris (Petersen and Pack 1982), a single sauropod tail vertebra (Lucas and Heckert 2000), invertebrate traces (Lockley and Wright 2003), and a variety of vertebrate traces including pterosaur tracks (Lockley et al. 1995; Lockley and Wright 2003; Mickelson et al. 2004), theropod tracks (Lockley et al. 1996), and sauropod tracks from near the top of the Summerville Formation at Moab (Lockley 1991). Invertebrate and pterosaur trace fossils have been found near Moab in the Summerville Formation (Mitchell and Lockley 2005; Lockley et al. 2007). If the Tidwell Member is included in the Summerville Formation, the fossil assemblage is broadened to include root traces (Kjemperud et al. 2008), algal and microbial laminations, charophytes, gastropods, ostracodes (Scott et al. 2001b), burrows (Kjemperud et al. 2008), horseshoe crab traces (Scott et al. 2001a), a lungfish tooth plate (Imhof and Trujillo 2005), indeterminate reptile bones (the putative pterosaur *Utahdactylus*; Bennett 2007), the poorly-known sauropod *Dystrophaeus* (Gillette 1996; Lucas and Anderson 1997), and pterosaur and dinosaur tracks (Lockley et al. 1998b).

Park Collections

There are fossils from ARCH held in the National Park System and in outside institutions. Swanson et al. (2005) reported 136 specimens in the NPS system from ARCH, which are curated at the Southeast Utah Group Headquarters in Moab. No fossils from ARCH were on display within the park as of 2005, but a few fossil specimens collected elsewhere are available as part of a "touch and feel" exhibit at the Visitor Center (Paleozoic invertebrates, a petrified tree trunk, a cross-section of dinosaur bone, and a theropod track) (Swanson et al. 2005). The *Grallator* slab discussed in Lockley and Gierliński (2009) is currently in the ARCH maintenance yard, but is not being curated as the park lacks the facilities for such a large specimen (V. Webster, Southeast Utah Group museum curator, pers. comm., December 2010).

Several other institutions have ARCH specimens, or representations of them, in their collections. Over 900 packrat midden specimens collected from ARCH were originally held at Northern Arizona University (Flagstaff, AZ) (Swanson et al. 2005). These specimens have since been moved to the Museum of Northern Arizona (Flagstaff, AZ), and in the process, four additional uncataloged specimens were found and added to the park catalog (V. Webster, pers. comm., December 2010). Tracings of the *Grallator* track slab from the Chinle Formation site on Highway 191 are in the University of Colorado at Denver Dinosaur Tracks Museum, and a few track specimens in their collections are from this site (CU 147, 148.1, 148.2, and 148.4) (Lockley and Gierliński 2009). Molds and casts from the Ruby Ranch Member tracksite near Delicate Arch are also in the Dinosaur Tracks Museum (Lockley et al. 2004). The specimens mentioned by Swanson et al. (2005) in the New Mexico Museum of Natural History and Science are not from ARCH, pertaining to the misidentified Moenkopi Formation site outside of the park as discussed above.

Cultural Resource Contexts

Humans have been present in the ARCH area for thousands of years (Lohman 1975; Graham 2004), so it would not be surprising to find diverse paleontological resources with or as cultural resources. For example, as discussed above, Bison Alcove contains both paleontological and cultural resources, although the human presence was fairly recent (Mead et al. 1991). CANY and Cowboy Cave provide further examples for potential finds, such as a recent discovery of an antelope hide bag. This bag, dating to between 770 and 970 AD, was recovered in 2005 from a dune at the Horseshoe Canyon Unit. Among its contents were marsh-elder and other seeds, three leather pouches, a pressure flaker made from an antler, a chert biface, and chert flakes (Geib and Robins no date). As noted earlier, nearby Cowboy Cave was occupied by humans over multiple intervals spanning approximately 9,000 years, with numerous artifacts made from the contemporaneous biota. Stokes (1972) interpreted three petroglyphs in southeastern Utah, including one near Moab, as representing proboscideans (possibly mastodons), but such identifications are extremely controversial. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- Swanson et al. (2005) provided extensive recommendations for the paleontological resources of ARCH, divided into interpretive actions, management, protection, research, and collections.

- The park should consider future field inventories for paleontological resources to more fully document *in situ* occurrences of fossils. The park may consider a formal site documentation and condition assessment for significant fossil localities. Monitoring of significant sites should be undertaken at least once a year in the future. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- Within the NCPN, the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and increase understanding of this important fossil-bearing unit. A similar USGS–NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossil theft and vandalism are of concern at ARCH; for example, Swanson et al. (2005) mentioned the theft of sauropod bones from near Wolfe Ranch, and Wann Langston reported fossil theft from a marked site as described above. Parks such as Petrified Forest National Park have investigated casual collecting of fossils and the social science driving the behavior. Contact the Geologic Resources Division for further information or assistance.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- If a fossil packrat midden is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers. Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Black Canyon of the Gunnison National Park

Black Canyon of the Gunnison National Park (BLCA), Colorado, protects a portion of the Black Canyon, a steep-walled chasm cut by the trapped Gunnison River (BLCA Figure 1). BLCA was proclaimed March 2, 1933 as a national monument, and was redesignated a national park October 21, 1999. Its boundaries have changed a number of times: May 16, 1938; October 28, 1939; April 13, 1960; July 13, 1984; October 21, 1999, and November 11, 2003. It presently encompasses 13,334.40 ha (32,950.03 acres), all but 890 ha (2,200 acres) of which are under federal administration (BLCA Figure 2). A wilderness area was designated in BLCA October 20, 1976; it presently encompasses 6,313 ha (15,599 acres) of the park. BLCA is administered with Curecanti National Recreation Area (CURE), which adjoins it on the southeast.

Geologic Background

BLCA occupies part of northeastern Montrose County in western Colorado. The central Black Canyon of the Gunnison passes through the park. This canyon begins upstream in adjoining CURE and continues downstream through Gunnison Gorge National Conservation Area, administered by the Bureau of Land Management. To the northeast are the West Elk Mountains, which include the Gunnison National Forest. Physiographically, the Black Canyon crosses the Rocky Mountain and Colorado Plateau provinces, with BLCA within the Colorado Plateau province and CURE within the Rocky Mountain province (Hansen 1981). The city of Montrose is about 12 km (7 miles) southwest of the western terminus of BLCA, the city of Delta is about 28 km (17 miles) northwest of the western terminus, and the city of Gunnison is about 60 km (37 miles) east of the eastern terminus. Documents concerning the general geology of BLCA include Hansen (1965, 1967, 1968, 1971, 1981, 1987a, 1987b), Heise (1999), Kellogg et al. (2004b), and Thornberry-Ehrlich (2005). Documents including information about the paleontology of BLCA include Fiorillo (1996), Scott et al. (2001a), Trujillo (2001), Koch and Santucci (2002), Koch (2005), Kenworthy (2006), Koch et al. (2006), Merewether et al. (2006), and Olstad (2011). The Geologic Resources Division of the National Park Service (NPS) coordinated a geologic resource inventory scoping session for BLCA and CURE (Heise 1999) and produced a geologic resource evaluation report for the two units in 2005 (Thornberry-Ehrlich 2005). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The oldest rocks exposed at BLCA are metamorphosed sedimentary rocks that are at least as old as the Paleoproterozoic (see the appendix for a geologic time scale), predating 1,700 Ma (million years ago) (Hansen 1981). During this time, the area was covered by a thick sequence of sediments (Hansen and Peterman 1968), possibly as part of an arc or back-arc basin (Jessup et al. 2006). The geographic area consisted of a series of island arcs similar to the present-day Banda Sea in Indonesia, colliding with the growing proto-North American continent around 1,700 Ma (Jessup et al. 2005). During this process, igneous rocks intruded the crust at depth, and additional crustal fragments accreted onto the continent, deforming the original rocks of the Black Canyon region. Tectonic activity paused, then resumed around 1,430 Ma and lasted for about 30 million years, producing another episode of deformation (Jessup et al. 2006). Diabase dikes intruded around 510 Ma (Hansen 1981), but otherwise the Neoproterozoic and Paleozoic are not represented by rocks in the Black Canyon. It is known that the Black Canyon area was part of the ancestral Uncompahgre uplift around 300 Ma (Kellogg 2004), which would have led to erosion of overlying rocks and precluded deposition until the uplift in turn was eroded.



BLCA Figure 1. A view into the steep-walled Black Canyon of the Gunnison (VLS).

There follows a hiatus in the rock record at BLCA until the Middle Jurassic. The dune sandstones of the Entrada Sandstone are the oldest such rocks in the Black Canyon area. This formation transitions into the shallow marine setting represented by the lower Wanakah Formation (Hansen 1965; Kellogg 2004), which in turn was replaced by a short-lived dune field (Kellogg 2004). Aridity decreased through the Late Jurassic (Lockley and Wright 2003). The more humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the last geologic formation of the Jurassic in BLCA, the well-known Morrison Formation. The Morrison Formation is interpreted as a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004); however, a narrow marine embayment (the Black Canyon Embayment) persisted into early Morrison time in the Black Canyon area, permitting marine influence (O'Sullivan 1992). During deposition of the upper portion of the formation (the Brushy Basin Member), a vast basin is thought to have extended from near Grand Junction, Colorado to Albuquerque, New Mexico. This basin was periodically occupied by a shallow saline lake which has been named Lake T'oo'dichi'; the setting is described as analogous to the modern Lake Eyre basin in Australia (Turner and Fishman 1991). Another fluvial and floodplain unit, the Burro Canyon Formation, followed the Morrison Formation during the Early Cretaceous (Hansen 1965).

Terrestrial processes continued until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Sandstone signals the onset of several cycles when the Western Interior Seaway transgressed (advanced) and regressed (withdrew) over the land (Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Marine deposition at BLCA from the Seaway is represented by the Mancos Shale (Hansen 1965). The Black Canyon area was under water for almost the entire Late Cretaceous, only becoming emergent again in the last few million years of the period (Lillegraven and Ostresh 1990; Franczyk et al. 1992). The sea was not very deep; depth during the most extensive marine transgression was on the order of 90 to 180 m (300 to 600 ft) in western Colorado and eastern Utah (Hill 1982). The Sevier Orogeny (mountain-building event) pushed up rocks on the west of the Colorado Plateau in central Utah, providing source rocks for the thick Late Cretaceous sedimentary formations of the Plateau (Dickinson and Gehrels 2008). At times, such as approximately 94 Ma, volcanoes were active to the west, leading to the deposition of ash beds (Bagshaw 1977).

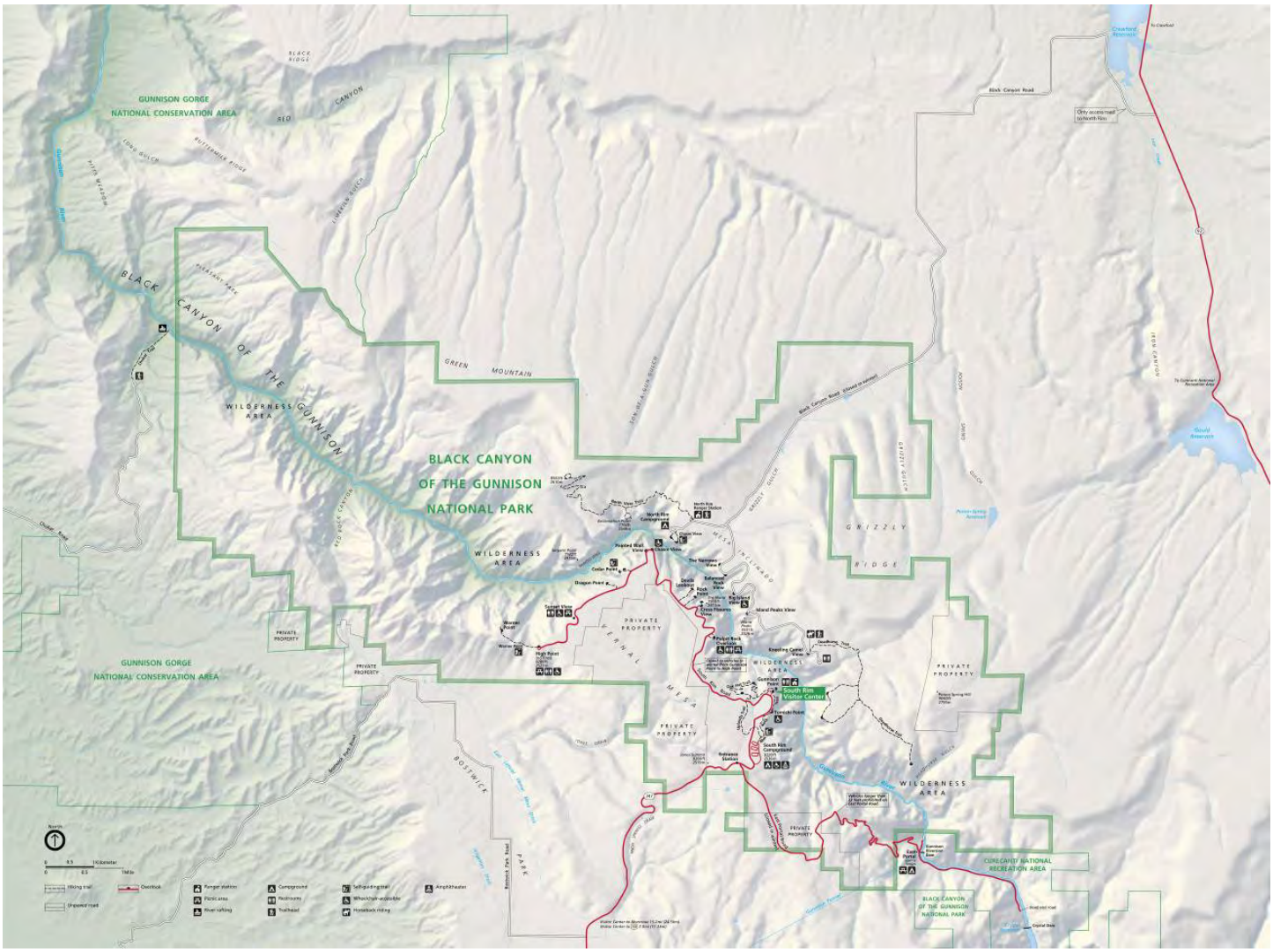
Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of a major mountain-building event known as the Laramide Orogeny (Cather 2003). This event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). The Gunnison uplift, which includes the Black Canyon, is a Laramide structure, although its movements were based on reactivated faults from the Proterozoic tectonic activity (Hansen 1981, 1987a). By the late Eocene, the Gunnison uplift had eroded back into its surroundings (Hansen 1987a).

In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent local igneous activity during the late Eocene, Oligocene, and Miocene. Several formations of volcanic ash, known as tuffs, are present within and near BLCA, erupted from volcanic calderas to the south. They form the rims of the Black Canyon within neighboring CURE. The oldest volcanic rocks belong to the West Elk Breccia, derived from an eruption that took place around 32 Ma (Hansen 1981) in the West Elk Mountains to the north. A series of tuffs followed from the San Juan Mountains between approximately 29 and 27 Ma (Olson et al. 1968; Steven and Lipman 1976; Hansen 1981). Volcanic activity dwindled after this time, persisting at a low level. For example, the Mesa Falls Ash, which erupted about 1.2 Ma, and the Lava Creek B Ash, which erupted about 640,000 years ago, are present in the Black Canyon (Sandoval and Karlstrom 2006).

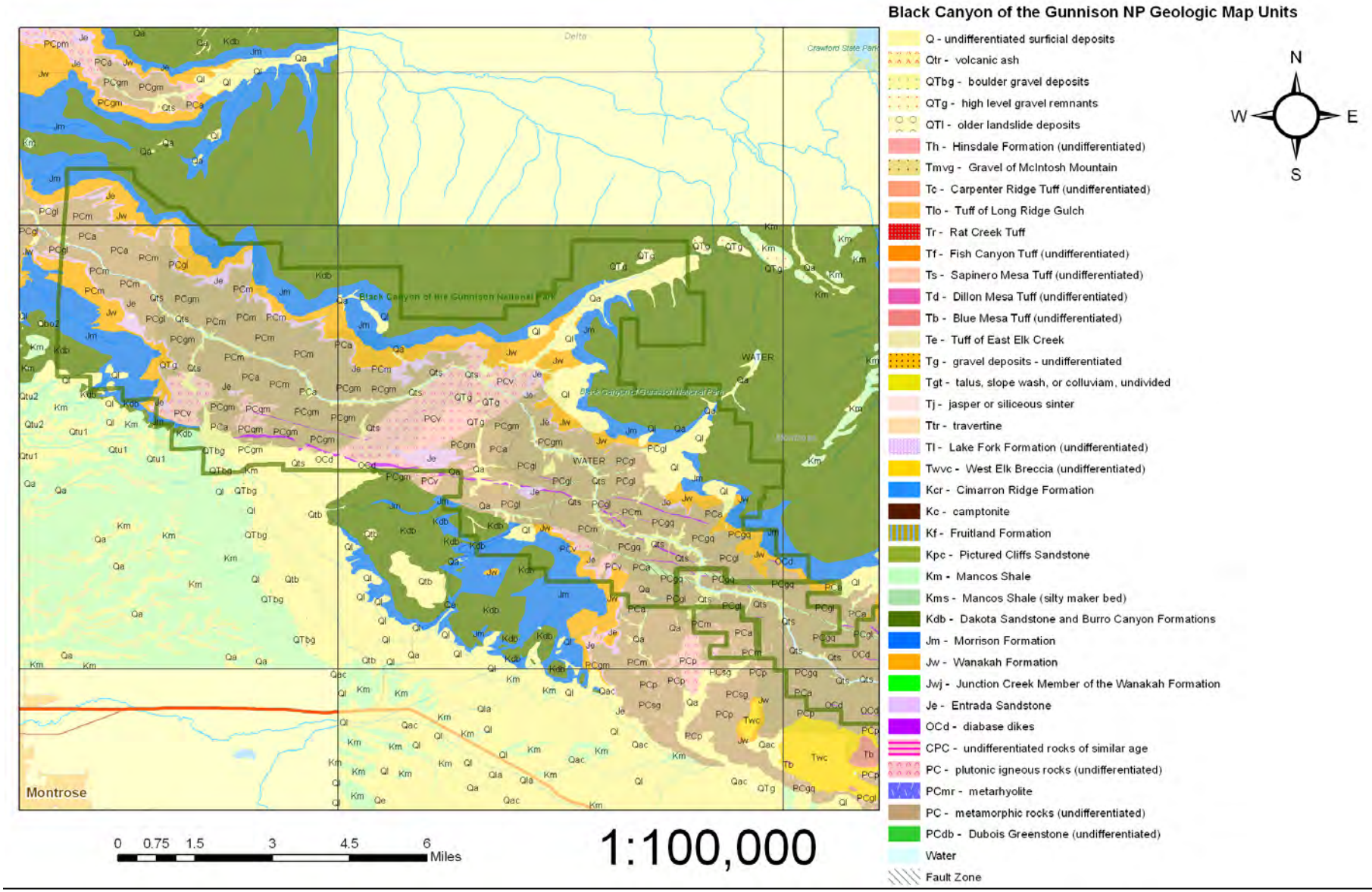
Glaciation occurred in the southern Rocky Mountains during the Pleistocene, which began approximately 2.6 Ma. There is evidence for glaciation on Grand Mesa to the northwest (Henderson 1923). The most recent local episode of widespread glaciation, the Pinedale Glaciation, ended around 14,000 years ago. Glaciation resulted in the formation of terraces on the Gunnison River because of increased sediment load from glaciers eroding mountains in the river highlands (Kellogg 2004). The river began downcutting in the area about 2.5 Ma (Kellogg 2004). Its incongruous presence in the hard crystalline Proterozoic rocks of the Black Canyon is due to being trapped there: it became confined between the thick deposits of volcanic rocks from the West Elk and San Juan volcanic centers and was further restricted by a syncline (concave fold), the axis of which was oriented favorably to the drainage (Hansen 1987a). By the time it had eroded through the Tertiary and Mesozoic rocks and reached the hard crystalline basement, it could not move laterally and was limited to downcutting, which produced the steep walls of the Black Canyon (Kellogg 2004). People arrived in the nearby Uncompahgre River valley by at least between 11,000 and 10,000 years ago, based on the presence of Clovis and Folsom points (Kellogg et al. 2004a).

Rock units exposed at BLCA include, from oldest to youngest: a variety of metamorphic and igneous rocks (Paleoproterozoic); diabase dikes (Cambrian or Ordovician); the Entrada Sandstone and Wanakah Formation of the San Rafael Group (Middle

Jurassic); the Morrison Formation (Upper Jurassic); the Burro Canyon Formation (Lower Cretaceous); the Dakota Sandstone (Lower–Upper Cretaceous); and the Mancos Shale (Upper Cretaceous). Surficial units include a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene) (see BLCA Figure 3 for a geological map) (Hansen 1968, 1971; Kellogg et al. 2004b). Cretaceous rocks younger than the Mancos Shale are present near BLCA to the south, in limited exposures (Dickinson 1965; Hansen 1971). The Wanakah Formation, Morrison Formation, Dakota Sandstone, and Mancos Shale are known to be fossiliferous at BLCA, and the other Mesozoic units are potentially fossiliferous, as well as the Quaternary deposits. The Palroproterozoic rocks are unlikely candidates for yielding fossils, being either igneous in origin or heavily metamorphosed, and dating from a time before the evolution of hard parts (BLCA Table 1). The stratigraphic interval including the Wanakah Formation and the lower Morrison Formation has been of some dispute concerning the placement of unit contacts and the names used. This document will use the nomenclature of Thornberry-Ehrlich (2005), established from the maps cited above; alternative interpretations will be noted in the text.



BLCA Figure 2. Geography of BLCA (NPS).



BLCA Figure 3. Schematic geological map of BLCA. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1038963.

Formation	Age	Fossils Within BLCA	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	None to date; packrat middens and isolated bones of large mammals are most likely	Alluvial, eolian, fluvial, and landslide deposits
Mancos Shale	Late Cretaceous	Marine bivalves and other marine invertebrates	Open marine
Dakota Sandstone	Early–Late Cretaceous	Petrified wood, other plant fragments, leaf impressions, root traces, and invertebrate burrows	Terrestrial (especially fluvial) becoming shallow marine over time
Burro Canyon Formation	Early Cretaceous	None to date; fossils known elsewhere include plant fossils, freshwater invertebrates, and freshwater and terrestrial vertebrates	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	A leaf impression, root traces, bivalve molds, invertebrate burrows, and bone fragments	Fluvial, floodplain, and lacustrine settings
Wanakah Formation	Middle Jurassic	Plant molds, invertebrate burrows, and possible sponge spicules and vertebrate tracks; microbial ("algal") limestone is common in the Black Canyon of the Gunnison but not yet confirmed from BLCA	Open marine? to closed marine to eolian sand dunes over time
Entrada Sandstone	Middle Jurassic	None to date; fossils are very rare in this formation in general	Desert with large eolian sand dunes
Diabase dikes	Cambrian or Ordovician	Unfossiliferous (intrusive igneous rocks)	Not applicable (igneous rocks that solidified at depth)
Igneous and metamorphic rocks	At least Paleoproterozoic	None to date; fossils are extremely unlikely because of lithology and metamorphism	Not applicable (igneous rocks that solidified at depth and strongly metamorphosed sedimentary rocks)

BLCA Table 1. Concise tabulation of BLCA stratigraphy, fossils, and depositional setting; details and references can be found in the text. Igneous and metamorphic rocks are included here for completeness, but are not discussed in detail in the text because of the improbability of fossils.

Sources of Paleontological Resources

The fossils of BLCA present opportunities for education, interpretation, and continued or future scientific research in the park. Fossil research in the Black Canyon began in the 1990s with the work of Anthony Fiorillo and Cathleen May in CURE, who found trace fossils and dinosaur bones. There have been several paleontological surveys of BLCA and CURE since then, including in 2000, 2001, and 2002 by Kelli Trujillo, and in 2005 and 2011 by the National Park Service. The most recent survey reported 15 fossil sites in BLCA (Koch 2005; Koch et al. 2006). All 15 sites are on the North Rim (Koch 2005). The 2011 survey reported an additional 23 sites in BLCA (Olstad 2011), including 10 on the North Rim and 13 on the South Rim in the Wilderness Area between Red Rock Canyon and the Chukar Trail. The Dakota Sandstone is to date the most fossiliferous unit at BLCA, yielding 23 sites (Kenworthy 2006; Olstad 2011).

San Rafael Group: Wanakah Formation (Middle Jurassic)

The Wanakah Formation and overlying lower Morrison Formation are a problematic stratigraphic interval on the Black Canyon. The issue revolves around what should be included in the two formations. Historically, authors identified Jurassic eolian (wind-deposited) sandstone below the Morrison Formation as the Junction Creek Sandstone, sometimes as a member of the Wanakah Formation; see for example the geologic maps of Hansen (1971), Hedlund and Olson (1973), and Hedlund (1974). Research published in the 1990s instead

included the "Junction Creek Sandstone" of the Black Canyon and the overlying beds of the Wanakah Formation in the Morrison Formation, as the Tidwell Member (O'Sullivan 1992). This has been followed by other workers in the area (Kellogg 2004; Kellogg et al. 2004b). As noted above, for consistency with Thornberry-Ehrlich (2005), the older terminology is used here. The situation is further complicated by Lucas et al. (2006a)'s contention that the Summerville and Wanakah formations are synonymous; other workers, including Condon (1992), O'Sullivan (1992, 2004) and O'Sullivan et al. (2006), have found the Wanakah Formation to be older than the Summerville Formation.

The Wanakah Formation is divisible into multiple members at some locations (Goldman and Spencer 1941; Steele 1985; O'Sullivan 1992, 2004). Using the older definition, the Wanakah Formation of the Black Canyon includes up to four divisions. The basal part is the Pony Express Limestone Member, a discontinuous gray limestone up to 2 m (7 ft) thick. It is only present in the canyon west of Dead Horse Mesa (Hansen 1971), and is well-exposed on the northeast rim in the park (Hansen 1965). This member represents deposition in the northern part of a basin that extended into northern New Mexico, where the equivalent rocks are known as the Todilto Member (O'Sullivan 1992). The Pony Express Limestone Member is followed by a section composed of gypsum and gypsum-rich sandstone and mudstone, up to 44 m (145 ft) thick in BLCA. Above this is the Junction Creek Sandstone Member, a discontinuous gray, yellow, or pink dune sandstone up to 27 m (90 ft) thick. Finally, there is an upper part composed of gray silty mudstone and microbial chert-rich limestone, between 8 and 18 m (25 and 60 ft) thick (Hansen 1971). Under the more recent definition, the Junction Creek Sandstone and the overlying beds are the Tidwell Member of the Morrison Formation, and the Junction Creek Sandstone is renamed the informal "beds at Smith Fork" (Kellogg 2004). The Pony Express Limestone Member has been interpreted as unusual lake or marine deposits, with a marine origin more likely (O'Sullivan 1992; O'Sullivan et al. 2006). The gypsiferous Wanakah Formation beds above the limestone are interpreted as deposition from a shallow closed seaway. The hot climate promoted evaporation and the formation of gypsum beds. The overlying sandstone beds are dune sands deposited on the floor of the basin following the seaway drying up (Kellogg 2004). The formation dates to the end of the Middle Jurassic (Callovian stage) (Scott et al. 2001b), about 165 to 160 Ma.

The Wanakah Formation is not known for its fossils. Koch (2005) reported three sites in this formation at BLCA along the North Rim Road; these sites had invertebrate burrows. She also reported a fossil plant mold discovered as float on Wanakah Formation rocks, but noted that it was more likely to have been eroded from the Morrison Formation or Dakota Sandstone. Olstad (2011) reported five sites in this formation at BLCA, including plant molds on Green Mountain on the North Rim and invertebrate traces, a concentration of unidentified sponge-spicule-like features, and two possible vertebrate tracksites in the Red Rocks Wilderness on the South Rim. The microbial limestone of the upper section is a commonly-noted feature of this formation on the Black Canyon (Hansen 1965, 1968; Kellogg 2004). The limestone can be identified by wavy microbial structures (Kellogg 2004). The presence of microbial limestone has not been confirmed within BLCA, and this section of the formation is among the beds that have commonly been assigned to the lower Morrison Formation in recent years. Hansen (1965) also reported bivalves and gastropods in the microbial limestone. The Pony Express Limestone Member is not known to be fossiliferous in the Black Canyon, and is sparsely fossiliferous overall. It and its equivalent Todilto Limestone Member have yielded aquatic insects (O'Sullivan et al. 2006), ostracodes (seed shrimp) (Steele 1985; O'Sullivan et al. 2006) and fish fragments (Cross and Larsen 1915; Steele 1985; O'Sullivan et al. 2006). Reports from other areas of fossils in the upper section include root traces (Scott et al. 2001b), possible sponge spicules, bivalves, gastropods, ostracodes, indeterminate fossils, possible worm tubes and other invertebrate trace fossils (O'Sullivan et al. 2006), and possible pterosaur (flying reptile) feeding traces (Trujillo et al. 2004).

For reference, Summerville Formation fossils include plant debris (Petersen and Pack 1982), a single sauropod ("brontosaur") tail vertebra (Lucas and Heckert 2000), invertebrate traces (Lockley and Wright 2003), and a variety of vertebrate traces including pterosaur tracks (Lockley et al. 1995; Lockley and Wright 2003; Mickelson et al. 2004), tracks of theropods (carnivorous dinosaurs) (Lockley et al. 1996), and sauropod tracks from near the top of the Summerville Formation at Moab (Lockley 1991). If the Tidwell Member is included in the Summerville Formation, as suggested by Lucas et al. (2006a), its fossil assemblage is broadened to include root traces (Kjemperud et al. 2008), algal and microbial laminations, charophytes (green algae), gastropods, ostracodes (Scott et al. 2001b), burrows (Kjemperud et al. 2008), horseshoe crab traces (Scott et al. 2001a), a lungfish tooth plate (Imhof and Trujillo 2005), indeterminate reptile bones (the putative pterosaur *Utahdactylus*; Bennett 2007), the poorly-known sauropod *Dystrophaeus* (Gillette 1996; Lucas and Anderson 1997), and pterosaur and dinosaur tracks (Lockley et al. 1998).

Morrison Formation (Upper Jurassic)

The Morrison Formation is a heterogeneous and variegated unit dated to the middle Late Jurassic, between about 155 and 148 Ma (Foster 2007). The formation is divisible into several members. At BLCA, these include from oldest to youngest at least the Salt Wash and Brushy Basin members (Hansen 1965, 1981). As discussed above, the Tidwell Member is also reported as present in recent reports on the geology of the Black Canyon (O'Sullivan 1992; Kellogg 2004), due to a different interpretation of the underlying upper Wanakah Formation. The Salt Wash Member of the area is mostly composed of gray cliff-forming sandstone and red, gray, and green shale, between 35 and 52 m (120 and 170 ft) thick (Kellogg 2004). It often has

three sandstone ledges, and becomes thicker down the canyon (Hansen 1965). The Brushy Basin Member is mostly composed of variegated shale and yellow sandstone (Kellogg 2004), up to 105 m (350 ft) thick (Hansen 1981). The Salt Wash Member is interpreted as primarily fluvial in origin, while the Brushy Basin Member derives from the shallow Lake T'oo'dichi' basin and associated drainage (Foster 2003). The climate is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004). The upper contact with the Burro Canyon Formation is a disconformity in the Black Canyon area (Kellogg 2004).

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Brachiosaurus*, *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks (Turner and Peterson 1999; Foster 2007). Dinosaur bones were first found in the Morrison Formation of the Black Canyon by uranium prospectors in the early 1950s (Jenkins 2004). Fossils from the BLCA–CURE area were not studied in depth until the 1990s, when sites were located in CURE (Koch et al. 2006). Koch (2005) reported two fossil sites in Morrison Formation rocks at BLCA, yielding root traces, bivalve molds, invertebrate burrows, and a bone fragment. The most significant find in the Morrison Formation of BLCA is a "badly weathered" sauropod bone fragment at North Rim (Scott et al. 2001a). BLCA's invertebrate trace fossils were also briefly mentioned by Koch et al. (2006). Olstad (2011) reported five fossil sites in the Morrison Formation at BLCA, including invertebrate burrows and an impression of an angiosperm leaf on Grizzly Ridge on the North Rim, and invertebrate burrows, root traces, and a bone scatter in the Red Rocks Wilderness on the South Rim. There is also a site in the Salt Wash Member just outside of the park that has yielded a fragment of theropod tail vertebra, a sauropod rib, and other dinosaurian bone fragments (Anonymous 1990; Scott et al. 2001a). Neighboring CURE has a more extensive Morrison Formation fossil record at this time, including petrified wood (Kenworthy 2006), a mold of a conifer cone (Olstad 2011), rhizoliths (root traces), conchostracans (clam shrimp), unionid clam burrows, crayfish burrows, termite nests, other invertebrate burrows (Fiorillo 1999), isolated turtle and crocodile relative bones, teeth and bones of at least six dinosaur genera (*Allosaurus*, *Apatosaurus, Camarasaurus*, *Ceratosaurus*, *Stegosaurus*, and *Diplodocus* or *Barosaurus*), possible mammal bones, and large vertebrate burrows (Koch et al. 2006). The Morrison Formation finds in CURE and BLCA are of added interest because of the paucity of Morrison Formation fossils in this area of Colorado (Turner and Peterson 1999; Foster 2003). The only other significant find from the vicinity is another partial *Apatosaurus* skeleton, in this case from low in the formation about 10 km (6 miles) east of Gunnison (Bartleson and Jensen 1988). Both BLCA and CURE were part of the Morrison Formation Extinct Ecosystems Project, a cooperative synoptic survey of this well-known formation (Turner and Peterson 1998; Koch et al. 2006).

The Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites, fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans, ostracodes, and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods (lamp shells), gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006); however, some of the traces, such as those attributed to anthozoans, brachiopods, horseshoe crabs, and echinoderms are only known from marine-influenced units that are not found at BLCA (Hasiotis 2004). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians (tuatara), lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods, diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

Dakota Sandstone (Lower–Upper Cretaceous)

The Dakota Sandstone (also Formation or Group) is a heterogeneous rock unit with outcrops as distant as Minnesota; because it was named in Nebraska from rocks on the opposite

shore of the Western Interior Seaway, its use in areas on the western side of the seaway may be inappropriate (Young 1959; Witzke and Ludvigson 1994; Titus et al. 2005). In the Black Canyon, it is composed of sandstone and carbonaceous shale, with minor conglomeratic sandstone, and coal beds near the base (Hansen 1971). It can be 20 to 34 m (65 to 110 ft) thick in the area, and interfingers with the overlying Mancos Shale (Kellogg 2004). More resistant than the Mancos Shale, it caps hills, forms parts of the canyon rims, and produces hogback ridges (Hansen 1965).

On the Colorado Plateau it is common for the Dakota Sandstone to have conglomerates followed by coaly shale and then sandstone at the top. These divisions reflect environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma (see Titus et al. [2005] for discussion of the age). Hansen (1981) gave an age of approximately 105 Ma for the Dakota Sandstone of the Black Canyon.

The fossils of the Dakota Sandstone are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. It is the most fossiliferous unit at BLCA: 23 of the park's 38 fossil sites are in the Dakota Sandstone along the North Rim and in the far southwestern corner. These sites have yielded fossil wood (BLCA Figure 4), other plant fragments, leaf impressions (BLCA Figure 5), root traces, and invertebrate burrows (BLCA Figure 6) (Koch 2005; Kenworthy 2006; Olstad 2011). An example of these sites is Green Mountain Wood, where vehicular traffic and ungulate trampling have exposed petrified wood (BLCA Figure 4). At least one wood specimen from this site is from an early angiosperm (Kenworthy 2006). Leaf impressions found on Grizzly Ridge are from the genus *Salix* (willow) (Olstad 2011). Similar fossils have been found in the vicinity. Four sites at CURE have yielded plant fragments, wood, a plant mold, shell molds, and invertebrate burrows (Koch 2005). Coal beds, leaf and twig impressions, and invertebrate trace fossils are present in exposures near the lower Black Canyon (Kellogg 2004). Petrified wood may be found as well (Jenkins 2004). John Foster (Museum of Western Colorado curator of paleontology, pers. comm., November 2011) reported observing petrified logs up to 5 to 6 m (15 to 20 ft) long in this formation on a ranch outside the North Rim. Plant fossils are also known from the Ouray area 55 km (34 miles) to the south (Burbank 1930), and ammonites have been found near Delta (Young 1959).



BLCA Figure 4. Petrified wood from the Dakota Sandstone at Green Mountain (BLCA locality 5MN7666) (JPK).



BLCA Figure 5. A Dakota Sandstone leaf impression (BLCA locality 5MN9732) (photo courtesy Forest Frost, BLCA/CURE).



BLCA Figure 6. Examples of the invertebrate trace fossil Diplocraterion *in the Dakota Sandstone (BLCA locality 5MN9716) (photo courtesy Forest Frost, BLCA/CURE).*

The best-known terrestrial fossils from the Dakota Sandstone are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Sandstone (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Sandstone of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Sandstone for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (horsetail) (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Sandstone rocks in Utah (including from near Bryce Canyon National Park) reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Sandstone marine fossils include algae, foraminifera (another type of "shell"-forming protist; Tibert et al. 2003), dinoflagellate cysts (a type of plankton) (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils (such as marks left by the shells of dead ammonites striking the seafloor; Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Sandstone may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera (large foraminifera from the late Paleozoic useful for correlating and dating beds), corals, bryozoans (moss animals), brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Mancos Shale (Upper Cretaceous)

The Mancos Shale is a marine unit largely made up of shale with some sandy, silty, or carbonate-rich layers. It is composed of dark gray silty clay-rich shale with sandstone lenses and siltstone concretions in the Black Canyon. Although the upper portion is not preserved, it is still quite thick in the area, up to perhaps 670 m (2,200 ft) (Hansen 1971). This maximum thickness is found at Cathedral Peak (Hansen 1971). Easily eroded, it forms drab badlands of rounded hills in the area (Kellogg 2004). If the Black Canyon section was complete, the Mancos Shale would span on the order of 25 to 30 million years, given the length of time the area was beneath the Western Interior Seaway (Franczyk et al. 1992). Fossils from BLCA and Gunnison Gorge National Conservation Area show that the section only includes the period from about 100 to 85 Ma (Merewether et al. 2006; Ball et al. 2010). Deposition occurred as mud on the floor of the shallow Western Interior Seaway (Hansen 1965).

The Mancos Shale is a widespread formation in the Colorado Plateau and its internal subdivisions differ by location. It is not usually divided on the Black Canyon (Hansen 1971, 1981; Hedlund and Olson 1973; Hedlund 1974; Franczyk et al. 1992; Kellogg 2004; F. Frost [CURE/BLCA archeologist] and A. Mims [CURE/BLCA visual information specialist], pers. comm., February 2011), although a few recent documents have applied the terminology originated in the Mesa Verde area (Merewether et al. 2006; Ball et al. 2010). The members these authors use, from oldest to youngest, are the Graneros Member, the Bridge Creek Limestone Member, the Fairpoint Member, the Blue Hill Member, the Juana Lopez Member, the Montezuma Valley Member, and the Niobrara Member. Additional details on these members can be found in Leckie et al. (1997).

The Mancos Shale is fossiliferous within BLCA, having yielded marine invertebrates (Scott et al. 2001a; Merewether et al. 2006). Merewether et al. (2006) included a measured section from a site at Red Rock in southwestern BLCA (SE1/4 SW1/4 Section 29, T50N R8W). It was measured by E. A. Merewether and D. A Sawyer on September 25, 2003. Bivalve taxa collected from this site include *Inoceramus* sp. and *Pseudoperna congesta*, corresponding to the Bridge Creek Limestone Member, *Inoceramus dimidus* and *Lopha lugubris*, corresponding to the Juana Lopez Member, and *Pycnodonte* (Merewether et al. 2006). Just outside of the park to the south is U.S. Geological Survey site D6855, where fossils of the straight ammonite *Baculites obtusus* have been found (Cobban 1973; Merewether et al. 2006).

Diverse marine fossils are known from the Mancos Shale across the Colorado Plateau, with the most abundant being ammonites and bivalves. Marine microorganisms and invertebrates are represented by acritarchs (a "wastebasket" classification for organic microfossils that cannot otherwise be classified), dinoflagellates (Crowley et al. 1994), nannoplankton (Cohen 1959), radiolarians (Olesen 1991), foraminifera (Bagshaw 1977; Leckie et al. 1997), corals (Ball et al. 2010), brachiopods (Leckie et al. 1997), bryozoans (Hill 1982), ammonites, bivalves (Merewether et al. 2006; Ball et al. 2010), gastropods (Leckie et al. 1997), crustaceans (Kues 1980; Scott et al. 2001a),

ostracodes (Bagshaw 1977), crinoids (sea lilies) (Leckie et al. 1997), echinoids (sea urchins) (Hill 1982), and worm tubes and other invertebrate trace fossils (Leckie et al. 1997). Marine vertebrates are represented by sharks and other cartilaginous fish (Ekren and Houser 1965; Leckie et al. 1997), teeth, scales, and bones of bony fish (Leckie et al. 1997), turtles (Irby 1995), plesiosaurs (Carter 1991; Spielmann and Lucas 2006), mosasaurs (Kass 1999; Spielmann and Lucas 2006), and sea birds (Lucas and Sullivan 1982). Terrestrial fossils are present near the western margin of the continental seaway, and include root traces, coal, leaves (Hill 1982), ferns, gymnosperms, angiosperms (Young 1987), terrestrial microfossils (spores, and pollen from gymnosperms and angiosperms) (Crowley et al. 1994), crocodilian bones (Irby 1995), dinosaur bones (Hill 1982) and at least one partial dinosaur skeleton (Lucas et al. 2006b). Hundreds of palynomorph (organic microfossil) taxa are known. Thompson (1972) reported 220 taxa of acritarchs, pollen, spores, and dinoflagellates and other algae from the Mancos Shale in the Mesa Verde area, and Cushman (1994, 1996) found 118 taxa of terrestrial plants and 146 taxa of marine palynomorphs in western Colorado.

Fossils are common in the Mancos Shale of western Colorado. Plant fragments, bivalve shells, invertebrate burrows, and various fish remains including scales, fins, and gill rakers have been found within the Mancos Shale of CURE (Koch 2005; Olstad 2011). Plant remains, a coral, ammonites, bivalves, burrows, fecal pellets, a shark tooth, bones, and fish scales have been recovered from a core taken from western Gunnison Gorge National Conservation Area near Olathe, over 12 km (7 miles) west of BLCA (Ball et al. 2010). Trujillo (2001) reported observing diverse fossils from the Mancos Shale in the Juana Lopez Member of the Gunnison area, including ammonites, bivalves, and isolated shark teeth and tetrapod limb bones. Lime-rich beds and concretions are fossiliferous in the area (Hansen 1965). A species of mosasaur (*Prognathodon stadtmani*) was named from a partial skeleton discovered near Cedaredge, about 35 km (22 miles) northwest of western BLCA (Kass 1999). Mollusks have been found at a number of sites within a similar radius (Cobban 1973; Weimer 1982; Merewether et al. 2006).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within BLCA. However, they are known to preserve fossils elsewhere, and future field investigations within the park may recover fossils from one or more of them.

San Rafael Group: Entrada Sandstone (Middle Jurassic)

The Entrada Sandstone of the Black Canyon region is a yellow and pink fine-grained sandstone with a basal conglomeratic section derived of weathered fragments of the Proterozoic basement rocks. The formation forms cliffs and "large rounded forms". It is 15 to 30 m (50 to 100 ft) thick and is well-exposed in the western part of the canyon (Kellogg 2004), but does not extend into CURE (Hansen 1971). Most of the formation has bedding associated with eolian dunes, but the uppermost part is thin-bedded, suggestive of shallow water deposition that occurred as the area transitioned into the shallow marine deposition of the overlying Wanakah Formation (Hansen 1965; Kellogg 2004). The Entrada Sandstone is divided into multiple members depending on location, but this has not been done at BLCA.

The Entrada Sandstone lacks age-diagnostic fossils and is regarded as late Middle Jurassic in age (Lucas et al. 1985), approximately 165 Ma. The climate was semi-arid to arid (Ekdale and Picard 1985) and hot, with sparse vegetation. The Entrada dunes made up the last major erg of the area, and formed when Utah was about 22° N (Loope 2008). The Entrada Sandstone system is perhaps analogous to the Baja California coast at Guerrero Negro (Scott et al. 2001b).

The Entrada Sandstone is not known to be fossiliferous at BLCA (Trujillo 2001; F. Frost and A. Mims, pers. comm., February 2011; Olstad 2011). Trujillo (2001) regarded it as likely to eventually yield vertebrate tracks within the park. The formation is very sparsely fossiliferous overall. Many reports of fossils from the Entrada Sandstone are from the Moab Member, which is now regarded as part of the Curtis Formation (Doelling 2010). These include the well-known megatracksite of the Moab area (Lockley 1991; Lockley et al. 2007) and the crocodylomorph *Entradasuchus* (Hunt and Lockley 1995). Trace fossils have been reported from the Escalante Member in south-central Utah, including rare root traces, insect burrows (Loope 2008), vertebrate burrows, traces of sand-swimming vertebrates (Graverson et al. 2007), and theropod (large *Megalosauripus* and smaller *Therangospodus*) and sauropod (cf. *Brontopodus*) dinosaur tracks (Foster et al. 2000). Hoggan (1970) reported bivalves and crinoids in marine facies of the Entrada Sandstone in the Vernal area.

Burro Canyon Formation (Lower Cretaceous)

The Burro Canyon Formation of BLCA and its surroundings is poorly exposed and difficult to distinguish from the overlying Dakota Sandstone (Hansen 1965), with which it is often mapped. The main road into BLCA climbs up on the Burro Canyon Formation, and it is exposed 30 to 65 m (100 to 200 ft) below the rims of Grizzly Ridge and Fruitland Mesa (Jenkins 2004). It is mostly composed of cliff-forming light gray conglomeratic sandstone and pebble conglomerate; some shale beds are also present (Hansen 1971). Chert pebbles are common, especially near the top of the unit (Kellogg 2004). It is

less than 35 m (115 ft) thick (Hansen 1981). In this area, it is interpreted as the deposits of meandering streams flowing across broad alluvial plains (Hansen 1965). Sediments were deposited under a semi-arid climate, probably reworking material from the Morrison Formation (Walters et al. 1987). The predominant paleoflow was to the north (Craig 1981). The upper contact with the Dakota Sandstone is conformable (Harris et al. 2004).

The age of the Burro Canyon Formation is uncertain. Hansen (1981) gave an age of about 110 Ma for the Burro Canyon Formation in the Black Canyon. It is usually thought of as having been deposited between the Barremian stage of the Early Cretaceous (130 to 125 Ma) and the onset of Dakota Sandstone sedimentation around 100 Ma. However, the base of the formation has been reported to intertongue with the Brushy Basin Member of the Morrison Formation in places, which would imply a Late Jurassic age for part of the formation (Ekren and Houser 1959; Dickinson and Gehrels 2008). It should be noted that the Burro Canyon Formation and Brushy Basin Member can be difficult to distinguish. Mudstones of the Burro Canyon Formation weather as fragments, while Brushy Basin Member mudstones form a hard frothy crust (Ekren and Houser 1959). The Burro Canyon Formation and Cedar Mountain Formation of Utah are lateral equivalents. Generally, the Burro Canyon Formation is present east of the Colorado River, and the Cedar Mountain Formation is present west of it (Dickinson and Gehrels 2008).

The Burro Canyon Formation is not known to be fossiliferous within BLCA, but has some potential (Trujillo 2001). In general, the fossils of the Burro Canyon Formation are not well known at this time. The current fossil assemblage from the formation over its entire area of outcrops includes organic debris, lake-dwelling algae, palynomorphs (organic microfossils) including conifer and angiosperm pollen (Tschudy et al. 1984), fern leaves (Simmons 1957), *Tempskya* (Dayvault and Hatch 2005), cycadeoids (Tidwell et al. 1976), conifers (Simmons 1957), early flowers (Fishell 1981; Armstrong and McReynolds 1987), plant fragments (Stokes 1952), wood fragments (Tschudy et al. 1984), freshwater bivalves (Reeside 1957), gastropods, freshwater ostracodes, ganoid fish scales (Simmons 1957), and dinosaur bones (Trujillo et al. 2004), including a theropod femur and claw (Armstrong 2010). Additionally, a site in Utah near the border with Colorado (the Hotel Mesa Quarry) has yielded fossils of sharks, bony fish, lungfish, various crocodile relatives, theropods, the sauropod *Brontomerus*, and ornithopods (beaked bipedal dinosaurs) (Taylor et al. 2011). It has been attributed to both the Burro Canyon Formation (Kirkland and Madsen 2007) and the Ruby Ranch Member of the Cedar Mountain Formation; by geography, it is within the Burro Canyon Formation (Taylor et al. 2011).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary sediments of BLCA are a heterogeneous group of deposits left by water, wind, and gravity movements, scattered throughout the park (Hansen 1971; Hedlund and Olson 1973; Hedlund 1974). These deposits are not known to be fossiliferous at BLCA, but are potentially fossiliferous. Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Quaternary fossils are not common within about 100 km (60 miles) of BLCA, although there are sites near the park.

The best-known local Quaternary site is Haystack Cave, a cavity in the Sapinero Mesa Tuff in an area about 40 km (25 miles) east of BLCA and about 1 km (less than a mile) north of CURE's Blue Mesa Reservoir (Emslie 1986). Sediments in the cave represent both the late Pleistocene and the Holocene, with human artifacts in the upper deposits. Bones from the cave have been dated to between 14,935 ± 610 radiocarbon years before present (19,365–16,880 calendar years before present) and 12,154 ± 1,700 years before present (18,640–10,135 calendar years before present) (Emslie 1986). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate. A wide variety of animal remains have been reported, including bones of frogs, lizards, raptors, ground birds, perching birds, rodents, pikas, rabbits, the extinct American cheetah *Miracinonyx*, mountain lions, coyotes, foxes, badgers, skunks, weasels, bears, horses, deer, pronghorns, bighorn sheep, bison, the extinct muskox *Bootherium*, and the extinct shrub ox *Euceratherium* (Emslie 1986; Scott et al. 2001a). The accumulation probably comes from a combination of denning carnivores, rodent (particularly packrat) nesting, and birds roosting. The site has been damaged by vandals (Emslie 1986).

Recently, a second extensive cave assemblage has been found near Crested Butte, 65 km (40 miles) northeast of BLCA. This site, Crescent Creek Cave, includes the only known high-elevation fauna that spans the last glacial maximum, with material dated from 43,330 to 1,120 radiocarbon years before present. Bones of frogs, snakes, ducks, ground birds, perching birds, bats, rodents, pikas, rodents, dogs, weasels, skunks, bears, horses, bison, bighorn sheep, and *Euceratherium* have been found here. Holocene-age shrews and voles have also been found at the nearby Signature Cave (Emslie 2011), and Pleistocene shrews are reported from Cement Creek Cave, another Crested Butte-area cave (Emslie 2002; Emslie et al. 2005). There are multiple reports of Quaternary fossils from Colorado National Monument (COLM) 85 km (53 miles) to the northwest, including a mastodon tooth (Scott et al. 2001a), bone and plant specimens from a fissure (Trujillo et al. 2004), packrat middens (Anderson et al. 2000; Scott et al. 2001a; Trujillo et al. 2004), and charcoal from throughout the Holocene (Scott et al. 2001b). Another recent discovery, albeit somewhat farther away, is the Snowmastodon site, over 85 km (53 miles) northeast of BLCA outside of Snowmass Village. This site,

discovered in 2010 and presently within Ziegler Reservoir, yielded several thousand bones deposited in a lake setting between 150,000 and about 45,000 years ago. More than 30 vertebrate taxa are represented, including fish, frogs, salamanders, snakes, lizards, birds, ground sloths, bats, rodents, rabbits, weasels or minks, otters, Columbian mammoths, mastodons, horses, deer, bison, camels, and others (Stucky et al. 2011a, 2011b). Non-vertebrates are represented by pollen, seeds, cones, leaves, wood (which sometimes bears beaver chewing marks), gastropods, ostracodes, and beetles and other insects (Anonymous 2011; Stucky et al. 2011b).

Paleoecological materials are somewhat more abundant in the area. Packrat middens have been found in the Black Canyon; Anderson et al. (2000) reported middens dated between 5,500 and 6,000 calendar years before present. More detailed information was not provided, but the U.S. Geological Survey/National Oceanic and Atmospheric Administration packrat midden database (http://esp.cr.usgs.gov/data/midden/) has a record of one midden cited to Anderson et al. (2000), with locality information that places it within CURE; it is not clear where this locality information came from. Alison Mims and Forest Frost (pers. comm., February 2011) do not know of any middens from BLCA or CURE. Emslie et al. (2005) reported on 17 middens collected in the upper Gunnison Basin in 1996, with four midden sites (Dillon Pinnacles Site 1, Red Creek Site 2, Two Ravens Cave 2, and West of Haystack Cave) from lands adjacent to CURE. The Gunnison Basin middens are all younger than 3,500 radiocarbon years before present, which may be due to preservation factors (Emslie et al. 2005). Ancient pollen has been reported in lake sites in the mountains to the north (Fall 1988, 1997; Thompson et al. 2000) and south (Maher 1961; Carrara et al. 1984; Thompson et al. 2000), and Holocene-age peat records have been recovered from fens on Grand Mesa (Sullivan and Gilmore 2006), but such sites are unlikely to be found in BLCA given its geography. Packrat middens, which can be found in rockshelters and alcoves, are more likely Quaternary fossils, as well as isolated resistant materials such as wood or teeth in surficial deposits.

Another source of Quaternary fossils in the Black Canyon area is from archeological sites. Several such sites are known. An example is the Kezar Basin site just south of CURE, which was used periodically between 8,800 and 5,800 years before present. It appears to have been a game processing site, and pelvic and shoulder girdle elements from deer and pronghorns have been recovered from it. Other local sites are younger (Stiger 2001).

Park Collections

Three paleontological specimens from BLCA are included in the park's museum collections at the Cimarron facility under the accession number BLCA 074. They are: a plant/tree mold (collection number BLCA 1795) from locality 5MN7669, collected by Alison Koch in June 2005; a sandstone block containing Early Cretaceous leaf impressions (BLCA 1796) from locality 5MN9717, collected by Tyra Olstad in July 2011; and three Wanakah limestone blocks containing unidentified sponge-spicule- or larva-like features (BLCA 1797a-c) collected by Tyra Olstad in August 2011. The fossils from Red Rock described by Merewether et al. (2006) are U.S. Geological Survey specimens, and are housed in the Survey's collections in Denver. Their catalog numbers are D14149 and D14150 (Merewether et al. 2006).

Cultural Resource Contexts

There are not yet any records of fossils found in cultural resource contexts at BLCA, but there is the potential for such finds. As noted earlier, people have been present in the area for millennia, having arrived in the Uncompahgre River valley between 11,000 and 10,000 years ago at least (Kellogg et al. 2004a). A concentration of Paleo-Indian sites (12,000 to 7,500 years ago) may exist along the Gunnison River east of Blue Mesa Dam, and the present area of the Blue Mesa Reservoir was "a high-use base camp area for aboriginal hunting and its associated activities (tool production and food processing)" (Moeller et al. 1993). Animal remains have been found at some of these sites, such as the Kezar Basin site mentioned above, and in some cases date to at least the Middle Holocene (Stiger 2001). Tools also represent a potential crossover between cultural and paleontological resources. It is not uncommon for people to use fossiliferous materials in tools. In the Black Canyon region, people used chert from the microbial limestone of the upper Wanakah Formation (or Tidwell Member of the Morrison Formation) for tools (Kellogg 2004). Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division produced a geologic resource evaluation report for BLCA and CURE in 2005, recommending that paleontological research projects be encouraged at the two units (Thornberry-Ehrlich 2005). This recommendation was echoed by Koch et al. (2006). Additional recommendations were provided by Koch (2005) and Kenworthy (2006).

- BLCA reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 15 localities. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009), including some discussion of sites within neighboring CURE. Fossil theft and vandalism are potential threats as well; for example, nearby Haystack Cave has been vandalized (Emslie 1986).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Black Canyon of the Gunnison National Park Digital Geologic Map
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Bryce Canyon National Park

Bryce Canyon National Park (BRCA), Utah, was established to preserve its scenic geologic resources, including colorful pinnacles, walls, and spires in horseshoe-shaped amphitheaters (BRCA Figure 1). Many of these features are formed from rocks of the Claron Formation, discussed below. BRCA was initially proclaimed as a national monument June 8, 1923. It was redesignated as Utah National Park on June 7, 1924, and became Bryce Canyon National Park February 25, 1928. Its boundaries have changed several times: May 12, 1928; June 13, 1930; January 5, 1931; February 17, 1931; May 4, 1931; and March 7, 1942. BRCA currently encompasses 14,501.94 ha (35,835.08 acres), all but 1.01 ha (2.50 acres) of which are under federal administration (BRCA Figure 2).

Geologic Background

BRCA is located is southwestern Utah, spanning parts of southwestern Garfield County and northwestern Kane County. It is bordered on the east and south by Grand Staircase–Escalante National Monument (GSENM), and on the west and northeast by two separate parcels of Dixie National Forest. BRCA occupies part of the eastern Paunsaugunt Plateau. The Pink Cliffs, formed in the Claron Formation, are oriented north to south through BRCA. The White Cliffs, another prominent member of the series of cliffs in the BRCA–Grand Staircase area, are found to the south. Several of southwestern Utah's High Plateaus are in the vicinity, with the Sevier Plateau to the north, the Aquarius and Awapa plateaus to the northeast, the Kaiparowits Plateau to the east, and the Markagunt Plateau to the west. Structurally, BRCA is near a pair of large Basin and Range faults: the Paunsaugunt Fault runs through extreme eastern BRCA, and the Sevier Fault is found to the west (Davis and Pollock 2010). The East Fork of the Sevier River flows near the western border of BRCA. Cedar City is 71 km (44 miles) to the west, and St. George is over 120 km (75 miles) to the southwest.



BRCA Figure 1. Colorful, deeply eroded landscape of BRCA (VLS).

The geology and paleontology of BRCA are discussed in a number of documents, including: Gregory (1940, 1951); Spitznas (1949); Vokes (1952); Brox (1961); Marine (1963); Davis and Krantz (1986); Lundin (1989); Bowers (1990); Eaton (1994, 1999, 2005); Eaton et al. (1998, 2009); Munk (1998); National Park Service (NPS) (1999); Ott (1999); Johnson et al. (2000); Tilton (2001); Koch and Santucci (2002); Pollock and Davis (2000); Sable and Hereford (2004); Goodin (2005); Hoffman (2005); Thornberry-Ehrlich (2005); Kenworthy (2006); Emerson et al. (2007); Williams and Lohrengel (2007); Doelling (2008); Pearce (2009); Davis and Pollock (2010); Santucci and Kirkland (2010); Williamson et al. (2011); and Biek et al. (in prep.). The Geologic Resources Division of the NPS coordinated a geologic resources inventory scoping session for BRCA during July 1999 (NPS 1999), and produced a geologic resources evaluation report for the park in 2005 (Thornberry-Ehrlich 2005). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The oldest rocks exposed at BRCA were deposited around 100 Ma (million years ago), shortly before a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the area (see the appendix for a geologic time scale). These rocks, part of the Dakota Formation, include the onset of several cycles when the Western Interior Seaway transgressed (advanced) and regressed (retreated) over the land (Gustason 1989; Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Because of BRCA's location at the western margin of the Seaway, only one large transgression-regression cycle is represented by a dominantly marine unit (the Tropic Shale) at the park. The Tropic Shale was deposited during the greatest transgression of the Seaway (Albright et al. 2007a). At the time it was forming, the western margin of the seaway was near Cedar City and the Pine Valley Mountains, just 70 km (40 miles) west of BRCA (Gustavson 1989; Eaton et al. 2001; Tibert et al. 2003). Coincidentally, this transgression also occurred at about the same time (~94 Ma) as an ocean anoxic event that is related to significant marine extinctions (Elder 1991; Kirkland 1996). At the same time the Seaway was present, volcanoes in an island arc near the present California–Nevada border introduced large amount of ash, which over time have altered to bentonitic clays. The Seaway retreated from BRCA following the deposition of the Tropic Shale (Davis and Pollock 2010). The shallow sea was replaced by beach and

coastal plain environments. Sediments deposited in these environments became the transitional Straight Cliffs Formation (Peterson 1969a; Anderson et al. 2010; Davis and Pollock 2010) and the fully terrestrial Wahweap Formation (Titus et al. 2005; Davis and Pollock 2010).

The Seaway fully retreated from south-central Utah during the Sevier Orogeny (mountain building event), about 71 Ma (Roberts 2005; Roberts et al. 2005). Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of the Laramide Orogeny (Cather 2003). This mountain building event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). The nearest such feature is the Kaibab uplift to the east and southeast of the park (Davis and Pollock 2010). Tectonic activity led to three periods of faulting at BRCA: between about 90 and 50 Ma, related to the Laramide Orogeny; between about 25 and 20 Ma, related to the growth and collapse of the southern Marysvale Volcanic Field (see below); and from about 15 Ma to the present, related to east-west Basin and Range extension (Davis and Pollock 2010). The modern topography of the BRCA area formed in response to increased erosion following recent uplifts and reorganization of the Colorado River drainage system (Davis and Pollock 2010).

BRCA has a hiatus of uncertain duration, possibly as much as 20 million years, between the Kaiparowits Formation and the next youngest rocks (the Claron Formation), during which deposition, if any, was not preserved. Much of the exposed rock at BRCA belongs to the Paleocene–Eocene-age Claron Formation. This formation was deposited as part of a depositional system that transitioned from predominantly fluvial to predominantly lacustrine (Davis and Pollock 2010), and is related to a series of lakes that formed in central and eastern Utah during the Paleogene (Davis et al. 2009). Erosion and weathering of the colorful Claron Formation produced the park's outstanding scenery. Younger rocks and sediments at BRCA are limited to isolated exposures of gravelly units and Quaternary sediments (Davis and Pollock 2010).

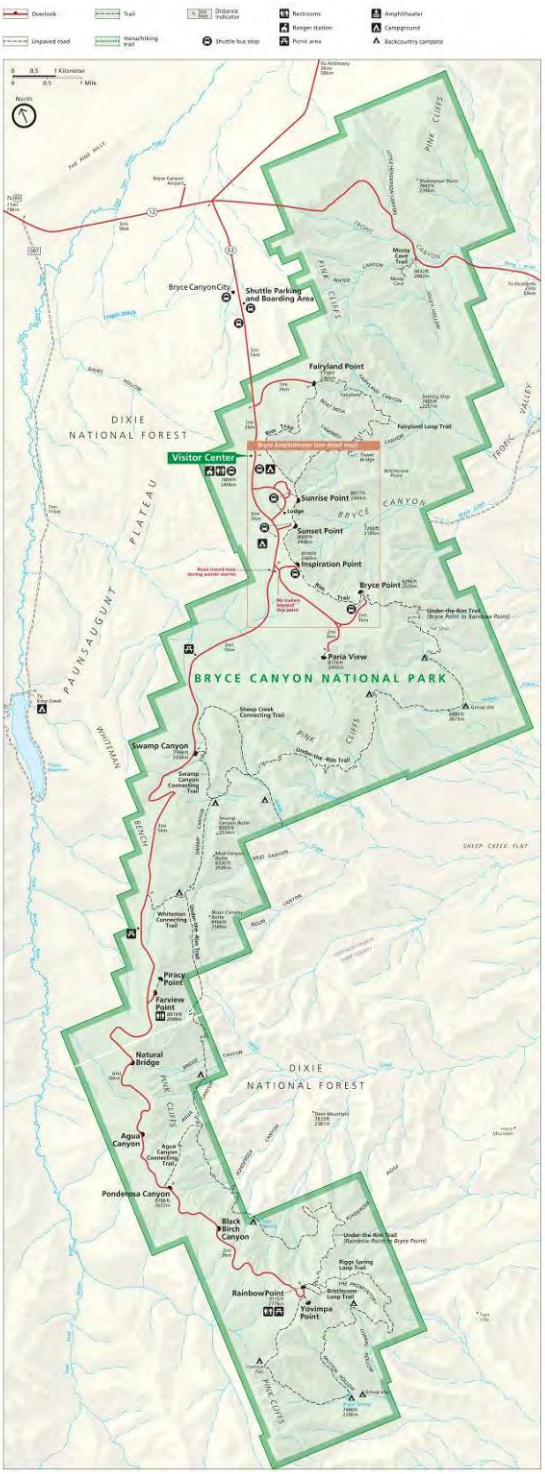
In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent widespread igneous activity during the late Eocene, Oligocene, and Miocene. The Cenozoic igneous activity of south-central Utah is dominated by the Marysvale volcanic field. Named for Marysvale, 86 km (53 miles) north of BRCA, Marysvale volcanics extend from Salina and the Fish Lake Plateau in the northeast to the Markagunt Plateau in the southwest (Rowley et al. 1975). Volcanism began in the Marysvale field by around 36 Ma and tailed off by around 14 Ma, but small-scale eruptions occurred as recently as the latest Pleistocene or possibly early Holocene (Rowley et al. 1994; Biek et al. 2011). The nearest preserved Marysvale volcanics are at the south end of the Sevier Plateau, immediately northwest of BRCA, and it is likely that volcanics covered the park at some point before being eroded. The southern part of the volcanic field, nearest BRCA, collapsed under its own weight around 20 Ma, which led to unusual thrust faulting and folding at the park (Davis and Pollock 2010). Southwest of the Marysvale volcanic field are shallow igneous intrusions, mostly formed between 22 and 20 Ma, known as the Iron Axis (Hacker et al. 2007). More recently, during the Pliocene and Quaternary, small basaltic volcanic eruptions occurred on the Markagunt and Paunsaugunt plateaus, marking the northern edge of the Western Grand Canyon basaltic field (Johnson et al. 2010; Biek et al. 2011, in prep.).

As the Pleistocene began approximately 2.6 Ma and continental glaciers formed over wide areas of the Northern Hemisphere, alpine glaciation occurred in Utah where elevations and precipitation were sufficient (Richmond et al. 1965). Small glaciers were present on both the Sevier Plateau to the north (Osborn and Bevis 2001) and on Brian Head just north of Cedar Breaks National Monument (CEBR; see the CEBR summary in this document for more information), 48 km (30 miles) to the west (Madsen et al. 2002; Biek et al. 2011); glaciation did not directly affect BRCA, though. People were present in the Colorado Plateau by 12,000 years ago (Davis and Pollock 2010). Neighboring GSENM has evidence of the Clovis culture, which existed before 11,000 years ago (Agenbroad 1997).

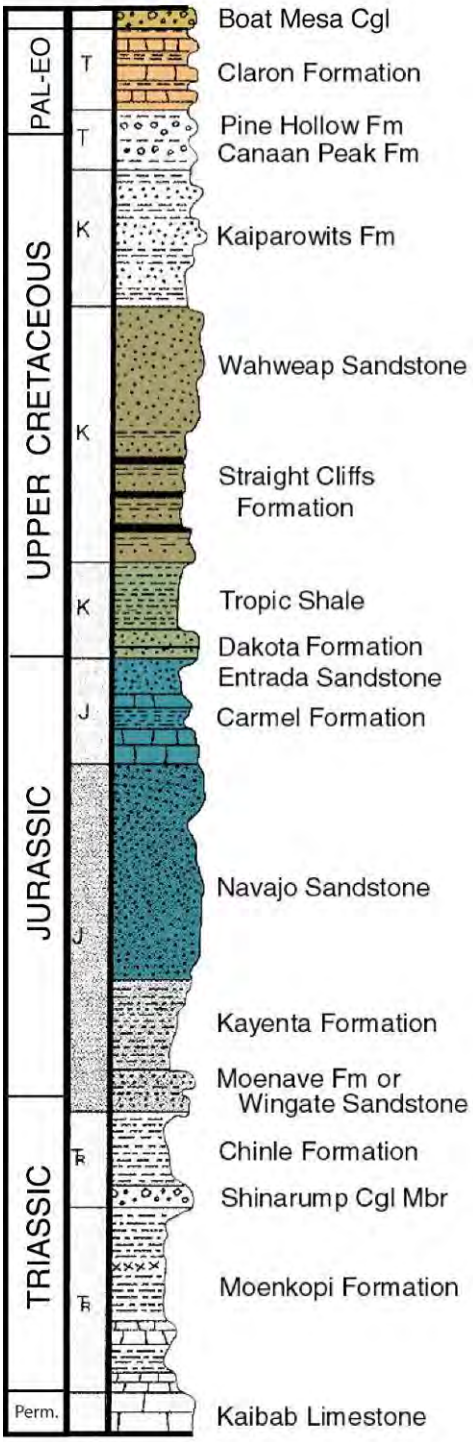
Rock units exposed at BRCA include, from oldest to youngest, the Dakota Formation, Tropic Shale, Straight Cliffs Formation, Wahweap Formation, and Kaiparowits Formation (Upper Cretaceous), the Claron Formation (Paleocene–Middle Eocene), the conglomerate at Boat Mesa (Upper Eocene), and various Miocene?–Quaternary rocks and sediments (see BRCA Figure 3 for a stratigraphic column and BRCA Figure 4 for a geological map) (Bowers 1990; Thornberry-Ehrlich 2005; Davis and Pollock 2010; Biek et al. in prep.). The Dakota Formation, Tropic Shale, Straight Cliffs Formation, Wahweap Formation, Kaiparowits Formation, and rarely the Claron Formation have yielded *in situ* fossils at BRCA, and the conglomerate at Boat Mesa has reworked fossils from Paleozoic formations. The other formations are potentially fossiliferous (BRCA Table 1). Collections of BRCA fossils are held both at the park and outside museums.

Incongruously for a setting with such expansive and impressive rock exposures, the formational nomenclature for the rocks present at BRCA is convoluted. Historically, particular trouble areas are the identity of the rocks between the lower Straight Cliffs Formation and the Claron Formation, and the nomenclature used to describe the rocks currently included in the Claron Formation. This is partly due to east-west facies changes in Upper Cretaceous strata deposited during regression of the Western Interior Sea, and partly due to stratigraphers working in different areas of the well-exposed Kaiparowits Plateau, the moderately well-exposed Paunsaugunt Plateau, and

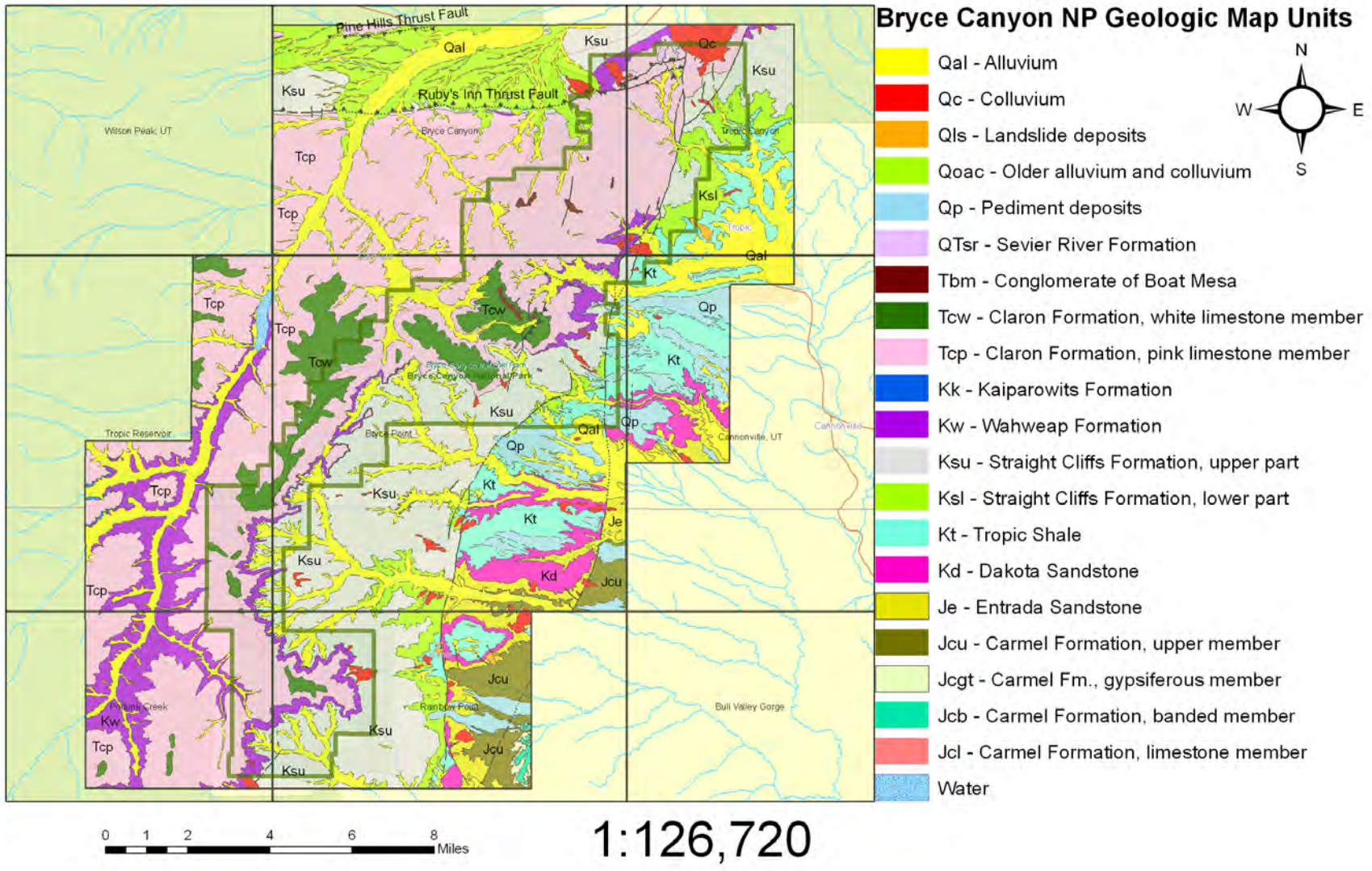
the poorly exposed Markagunt Plateau. Recent geologic mapping has now established correlation of Upper Cretaceous strata across these plateaus. Historical usages are mentioned below for reference.



BRCA Figure 2. Geography of BRCA (NPS).



BRCA Figure 3. Stratigraphic column for BRCA (from Davis and Pollock [2010], used with permission of the Utah Geological Association).



BRCA Figure 4. Schematic geological map of BRCA. The Kaiparowits Formation has since been identified from the park, and the unit here identified as the Sevier River Formation is no longer assigned to that formation (Biek et al. in prep). Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=2164782.

Formation	Age	Fossils Within BRCA	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	None to date; packrat middens and isolated large mammal bones are most likely	Alluvial, colluvial, landslide, and pediment deposits
Upper Cenozoic alluvial deposits (formerly identified as the Sevier River Formation)	Miocene?–Pliocene?	None to date; microbial and invertebrate fossils are most likely	Alluvial fans, valley fill, and minor fluvial and lacustrine deposition
Conglomerate at Boat Mesa	Late Eocene	Reworked Paleozoic invertebrates	Fluvial and overbank settings
Claron Formation	Paleocene–Middle Eocene	Bivalves, gastropods, invertebrate trace fossils, and a bone (probably reworked from an older formation)	Fluvial, alluvial, and lacustrine settings
Kaiparowits Formation	Late Cretaceous	None to date; plants, freshwater invertebrates, and vertebrates are commonly found in the vicinity	Fluvial, floodplain, and lacustrine settings
Wahweap Formation	Late Cretaceous	Bivalves, gastropods, crabs, ostracodes, and diverse vertebrates mostly represented by microvertebrate remains (teeth, small bones, and other small fossils recovered by screenwashing); vertebrates include cartilaginous and bony fish, amphibians, turtles, crocodilians, dinosaurs, and mammals	Fluvial settings
Straight Cliffs Formation	Late Cretaceous	Wood and other plant fossils, bivalves, gastropods, ostracodes, and diverse vertebrates mostly represented by microvertebrate remains; vertebrates include cartilaginous and bony fish, amphibians, turtles, lizards, crocodilians, dinosaurs, and mammals	Heterogeneous coastal deposition, including fluvial, floodplain, deltaic, beach, marsh, and lagoon settings
Tropic Shale	Late Cretaceous	Marine bivalves and gastropods	Open marine
Dakota Formation	Late Cretaceous	Petrified wood; invertebrates and microvertebrate fossils are also common in the area	Terrestrial (especially fluvial) becoming shallow marine over time

BRCA Table 1. Concise tabulation of BRCA stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of BRCA present opportunities for education, interpretation, and continued or future scientific research in the park. Most fossils in the park are microfossils from the Cretaceous rocks. These include bones and teeth of fish, turtles, crocodiles and crocodile relatives, dinosaurs, and marsupials. Vertebrates are relatively common in the strata above the John Henry Member of the Straight Cliffs Formation and below the Claron Formation (Eaton et al. 1998). Rocks along the East Fork of the Sevier River just west of BRCA have also produced many vertebrate fossils (Eaton 1999). Vertebrate localities are common in the BRCA area (Eaton 1999).

Vertebrate fossils from BRCA and the Paunsaugunt Plateau were first studied by Jeffrey Eaton of Weber State University (Ogden, UT) in 1988 (Eaton et al. 2009). Vertebrate fossils were only known from a few sites in BRCA until 2003. Before this time, there were only two microvertebrate locations of any quality in the Wahweap Formation of the park and one in the upper Straight Cliffs Formation. Screenwashing by Eaton during 2003 and 2004 resulted in the discovery of several new productive localities in the John Henry Member of the Straight Cliffs Formation (discussed below) (Eaton 2005). Following this, a six-year paleontological inventory was initiated at BRCA,

with the cooperation of the Bryce Canyon Natural History Association, Bryce Valley High School, the Geological Society of America's GeoCorps America program, the NPS, Southern Utah University, and Weber State University. In the first three years of this inventory, 127 new localities were found, 63 of which are within BRCA. 111 of these localities have mostly vertebrates, 13 have mostly invertebrates, and 3 are paleobotanical localities. Specialists from additional institutions have become involved in research from the fossils of these localities. Among other results, this project has documented the best record of vertebrates from the middle Upper Cretaceous (Coniacian and Santonian stages) in the Western Hemisphere, and the most complete record of Upper Cretaceous ostracodes in the world (Eaton et al. 2009). By October 2010, there were 144 catalogued fossil localities at BRCA, distributed among almost all of the park's formations: Tropic Shale (1), Straight Cliffs Formation (81, with 3 in the Smoky Hollow Member, 62 in the John Henry Member, 5 in the Drip Tank Member, and 3 undifferentiated upper), Straight Cliffs Formation/Wahweap Formation (2), Wahweap Formation (44, 5 in the lower Wahweap Formation), Wahweap Formation/Kaiparowits Formation (11), and Claron Formation (4) (S. Eagan, BRCA chief of resources management, pers. comm., October 2010).

Dakota Formation (Upper Cretaceous)

The Dakota Formation (also Sandstone or Group) is a heterogeneous rock unit with outcrops as far from Utah as Minnesota. Because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in Utah may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). The Dakota Formation is exposed in small areas of east-central BRCA. In the park and its immediate surroundings, it is composed of alternating sandstone, mudstone, carbonaceous mudstone, and coal, with a basal conglomerate, and is 55 to 90 m (180 to 300 ft) thick (Bowers 1990). The Dakota Formation of the Colorado Plateau is often divided into three informal members, with a conglomeratic base, a heterogeneous coaly middle, and an upper marine sandstone. This division reflects environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma. In GSENM, the middle of the formation was deposited around 96.0 Ma (Titus et al. 2005).

The fossils of the Dakota Formation are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. The best-known terrestrial fossils from the Dakota Formation are plant fossils (particularly angiosperm leaves) and vertebrate tracks.

The Dakota Formation is particularly noted for its paleobotanical material. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Formation (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Formation of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Formation for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, the horsetail *Equisetum* (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). The floral assemblage of Utah outcrops includes coal (Molenaar and Cobban 1991), palynomorphs (organic microfossils, in this case including moss spores, fern spores, and gymnosperm and angiosperm pollen; Nichols 1997), the tree fern *Tempskya* (Ash and Read 1976), *Equisetum*, possible gymnosperms, and angiosperms (Tidwell et al. 1976). Plant preservation includes impressions of stems, twigs, and leaves (Craig 1981), and petrified wood (Fisher et al. 1960).

More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks come from Dakota Group rocks that are several million years older than the rocks at BRCA, and include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Formation rocks in Utah (including from near BRCA) reported eggshells, gastropods, ostracodes, fish teeth, lizard jaws, and the teeth of crocodilians, dinosaurs, and mammals (Eaton et al. 1999a; Eaton 2004). Early mammal remains are of particular interest, and include multituberculates, marsupials, and therians (Eaton 1993b). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossils include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist) (van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993b). Dakota Formation marine fossils include algae, foraminifera (another type of "shell"-forming protist) (Tibert et al. 2003), dinoflagellate cysts (a type of plankton) (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites (typically nautilus-like cephalopods) striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Formation may also include reworked fossils

from older rock units, such as Paleozoic fusulinid foraminifera (large foraminifera from the late Paleozoic useful for correlating and dating beds), corals, bryozoans (moss animals), brachiopods (lamp shells), and echinoderms (Peterson 1969b), and reworked petrified wood (Titus et al. 2005).

The Dakota Formation of BRCA is fossiliferous, having yielded silicified wood (Marine 1963). Fossils found in the Dakota Formation of the surrounding Paunsaugunt Plateau and vicinity include organic debris (Sable and Hereford 2004), coal (May and Traverse 1973), palynomorphs (May and Traverse 1973), silicified logs (Sable and Hereford 2004), plant fragments, local concentrations of bivalves and gastropods (Gregory 1951), ammonites (Sable and Hereford 2004), freshwater sharks, rays, lungfish, frogs, turtles, lizards, crocodile relatives, dinosaurs, and multituberculate, marsupial, and other mammals (Eaton 1993b; Santucci and Kirkland 2010). Most of the vertebrate remains come from the upper middle portion (Eaton 1993b). The palynomorphs described by May and Traverse (1973) came from nearby Willis Creek Canyon, and included fern spores, conifer pollen, and angiosperm pollen.

Dakota Formation outcrops of the nearby Kaiparowits Plateau and the rest of GSENM have also yielded diverse fossils. These include algae, foraminifera (Tibert et al. 2003), coal, palynomorphs, petrified wood (Doelling 1977), leaves (Erwin 1997), bivalves, gastropods, ostracodes (seed shrimp) (Tibert et al. 2003), ammonites (Foster et al. 2001), a dragonfly nymph (Titus et al. 2005), invertebrate traces (Tibert et al. 2003), sharks, rays, ray-finned fish, lungfish, amphibians, turtles, lizards, several types of crocodilians and crocodile relatives, small theropods (carnivorous dinosaurs) such as dromaeosaurids and troodontids, tyrannosaurids, armored dinosaurs, hypsilophodonts (small bipedal herbivorous dinosaurs), hadrosaurids (duckbills), mammals (Eaton and Cifelli 1997; Eaton et al. 1999a; Eaton and Kirkland 2001, 2003; Titus et al. 2005), and footprint sites (Hamblin and Foster 2000). Most fossils there were found in floodplain settings of the middle unit, with some from lacustrine rocks (Eaton and Cifelli 1997; Eaton et al. 1999a), and algae, foraminifera, bivalves, gastropods, ostracodes, and burrows from the upper part (Tibert et al. 2003).

Tropic Shale (Upper Cretaceous)

The Tropic Shale crops out in northeastern BRCA. In the park and immediate vicinity, this unit is mostly composed of gray claystone with minor amounts of sandstone and mudstone near the top and bottom. It is 215 to 305 m (700 to 1,000 ft) thick (Bowers 1990). There is no depositional gap between the Tropic Shale and the overlying Straight Cliffs Formation (Davis and Pollock 2010), and the two formations interfinger in some locations (Tilton 2001). The Tropic Shale represents the maximum marine transgression of the Western Interior Seaway. Because the marine transgression reached different areas at different times, deposition occurred over a range of time during the early late Cretaceous (Tibert et al. 2003), roughly between 94 and 90.5 Ma (Titus et al. 2005; Albright et al. 2007b). In the BRCA region, the ammonites present indicate that the Tropic Shale here was deposited after approximately 94 Ma and before 92.5 Ma (Bowers 1990; Cobban et al. 2006). The Tropic Shale is roughly equivalent to the lower Mancos Shale found to the east (Leckie et al. 1997; Tibert et al. 2003), correlating to the Tununk Member of eastern Utah (Elder and Kirkland 1993, 1994; Albright et al. 2007a). The depositional setting was offshore marine, grading to foreshore and shoreface as the sea began to regress (Sable and Hereford 2004).

The Tropic Shale of BRCA is described as abundantly fossiliferous (Marine 1963). One site identified in this unit at the park in 2008 has yielded bivalves and gastropods (S. Eagan, pers. comm., October 2010). In the BRCA area, limestone concretions have yielded marine fossils (Bowers 1990). The Tropic Shale of the Paunsaugunt Plateau has yielded corals, bivalves, ammonites, gastropods, and worm tubes. The prevalent fine-grained rocks have marine forms, while the sandstone beds have marine and brackish-water forms (Gregory 1951).

Most fossils in the Tropic Shale are marine. Coal, wood (Gregory 1951), foraminifera (Tibert et al. 2003; McCormick 2006), corals, gastropods, bivalves, ammonites (Gregory 1951), ostracodes (Hazel 1969), serpulid worm tubes (Gregory 1951), ammonite shell impressions (Landman and Cobban 2007), other invertebrate trace fossils (Peterson 1969b), sharks (Hunt et al. 2006; Santucci and Kirkland 2010), bony fish, turtles, plesiosaurs (marine reptiles) (Albright et al. 2007a, 2007b), mosasaurs (marine reptiles related to monitor lizards) (Peterson 1969b), and a therizinosaurid dinosaur (Zanno et al. 2009) are known to date. The therizinosaurid, a long-necked long-clawed bipedal herbivore, is an unusual find so far from land (Gillette 2007). Fossil pearls are also known (Pollock and Davis 2004; Santucci and Kirkland 2010).

Most vertebrate finds in the Tropic Shale come from an area near Big Water, Utah, 64 km (40 miles) southeast of BRCA (Albright et al. 2007a, 2007b). These sites have yielded fossils from the lower part of the Tropic Shale, deposited between approximately 93.5 to 92 Ma (Albright et al. 2007a). The mollusks, bentonite beds, and limestone beds within the rocks permit high stratigraphic resolution (Eaton et al. 1987; Elder 1991). So far, researchers working in this area have found foraminifera (McCormick 2006), bivalves, ammonites, bony fish, the turtles ?*Desmatochelys* and *Naomichelys*, at least four species of plesiosaurs from two short-necked lineages (pliosaurid *Brachauchenius lucasi* and polycotylids *Dolichorhynchops tropicensis*, *Eopolycotylus rankini*, *Palmulasaurus quadratus*, and *Trinacromerum* ?*bentonianum*), and the therizinosaurid *Nothronychus graffami* (Albright et al. 2007a, 2007b, 2007c; Zanno et al. 2009; McKean 2012).

Straight Cliffs Formation (Upper Cretaceous)

The Straight Cliffs Formation is a heterogeneous unit deposited in multiple paleoenvironmental settings. It outcrops in the

eastern half of BRCA, where it is mapped as a lower and upper section (Bowers 1990). Biek et al. (in prep.) are mapping Upper Cretaceous strata in more detail as part of their Panguitch 30' x 60' quadrangle mapping project, which includes BRCA and its immediate vicinity. Rocks from the Straight Cliffs Formation weather into the Gray Cliffs of southern Utah (Davis and Pollock 2010). The Straight Cliffs Formation of south-central Utah is divisible into four members, based on exposures from the Kaiparowits Plateau. From oldest to youngest, these are the Tibbet Canyon Member, the Smoky Hollow Member, the John Henry Member, and the Drip Tank Member (Peterson 1969a). At BRCA, Bowers (1990) mapped the Tibbet Canyon and Smoky Hollow members as the lower Straight Cliffs Formation, and the John Henry and Drip Tank members as the upper Straight Cliffs Formation; Biek et al. (in prep.) mapped the Drip Tank Member separately. The two lower members combine for a thickness of 100 to 120 m (320 to 400 ft), and the two upper members combine for a thickness of 274 to 395 m (900 to 1,300 ft) (Bowers 1990). Davis and Pollock (2010) report the Drip Tank Member as absent at BRCA, but this is incorrect (J. Eaton, Weber State University Department of Geosciences professor, pers. comm., September 2011; Biek et al. in prep.).

The Tibbet Canyon Member is a cliff-forming sandstone that was deposited in beach and shallow marine settings. It grades out of the Tropic Shale, and grades into the Smoky Hollow Member (Peterson 1969a). The top of the Tibbet Canyon Member might represent fluvio-deltaic deposition (Titus et al. 2005). In the BRCA area, it is orange-gray in color and has abundant bivalves in the upper strata (Tilton 2001). It was deposited during the late Early Cretaceous (Eaton et al. 1999a), approximately 92 Ma. Its fossil assemblage includes bivalves, ammonites, gastropods, invertebrate burrows (Peterson 1969b), sharks, rays, gars, lungfish, crocodilians (Titus et al. 2005), theropod dinosaur teeth (Eaton and Cifelli 1988), and marsupials (Eaton and Cifelli 1997).

The Smoky Hollow Member is a heterogeneous unit of mudstone, limy mudstone, sandstone, and coal. It formed in fluvial, floodplain, lagoonal, and marsh environments, and can be divided into three parts: a lower coal/limy mudstone unit from lagoonal and marsh settings, a middle fluvial sandstone/floodplain mudstone unit, and the upper fluvial Calico sandstone (Peterson 1969a). In the BRCA area, the lower part is mostly gray mudstone with thin carbonaceous shale near the base, the middle section is orange-gray sandstone and gray mudstone, and the Calico sandstone is composed of white to grayish orange pebbly or granular sandstone. There is a disconformity between the middle and Calico sandstone strata (Tilton 2001). The Smoky Hollow Member was deposited during the early Late Cretaceous (Parrish 1999), at a time when the sea had retreated (Eaton and Cifelli 1997). Regionally, its fossil assemblage includes coal (Peterson 1969a), leaves (Erwin 1997), sparse wood fragments and plant impressions (Sable and Hereford 2004), foraminifera, bivalves, gastropods, ostracodes (Eaton et al. 2001), sharks, rays, ray-finned fish such as bowfins and gars, the unusual amphibian *Albanerpeton*, frogs, turtles, lizards, several types of crocodilians and crocodile relatives (Eaton et al. 1999a), pterosaurs (flying reptiles) (Titus et al. 2005), small theropods such as dromaeosaurids and troodontids, tyrannosaurids, armored dinosaurs, hypsilophodonts, hadrosaurids (Eaton et al. 1999a), and multituberculate (Eaton 1995), symmetrodont, marsupial, and possible eutherian mammals (Eaton et al. 1999a). The dinosaurs are represented by teeth from the middle of the member (Parrish 1999). Microvertebrate fossils are abundant in the middle interval (Titus et al. 2005).

The Smoky Hollow Member is fossiliferous at BRCA. Pearce (2009) briefly described fossils from a Utah Museum of Natural History locality (UMNH IP24) within the park. The most fossiliferous level is a carbonaceous siltstone above a lignite that is 15.5 m (51 ft) above the base of the unit. The site is interpreted as a brackish-water lagoon environment, with fossils of bivalves, gastropods, ostracodes, sharks, rays, bowfins, gars, other bony fish, and crocodiles or crocodile relatives. It is part of a regressive sequence that began during deposition of the upper part of the underlying Tibbet Canyon Member. Adjacent depositional environments, both vertically and laterally, include swamps (represented by lignite beds) and barrier beaches (represented by sandstones) (Pearce 2009). Hoffman (2005) described a brackish-water mollusk site at Glory Cove on the northeastern side of the park in this member. Multituberculate mammals are present in the BRCA area (Eaton 1995), and a hadrosaur skeleton has been found in the member east of BRCA (Santucci and Kirkland 2010).

The John Henry Member is a mixed unit similar to the Smoky Hollow Member. It was mostly deposited under nonmarine conditions west of the Kaiparowits Plateau, and under marine conditions east of it (Eaton and Cifelli 1988). It interfingers with the overlying Drip Tank Member. Multiple coal beds are present (Peterson 1969a). The John Henry Member of the BRCA area is composed of alternating grayish orange sandstone and gray to red-purple mudstone in a roughly 2:1 ratio (Tilton 2001). It was deposited during the middle Late Cretaceous (Eaton et al. 1999a), between about 87 and 84 Ma, during another marine transgression (Eaton and Cifelli 1997). Fossils are locally abundant (Titus et al. 2005). Regionally, its fossil assemblage includes coal (Peterson 1969a), palynomorphs (McCord 1997), sparse wood fragments and plant impressions (Sable and Hereford 2004), bivalves, cephalopods (Peterson 1969a), gastropods (Gallin et al. 2010), crabs (Eaton and Cifelli 1988), invertebrate burrows (Peterson 1969a), sharks, rays, ray-finned fish including bowfin and gars, *Albanerpeton*, frogs, turtles, lizards, several types of crocodilians and crocodile relatives, dromaeosaurids, armored dinosaurs, hadrosaurids (Eaton et al. 1999a), mammals (including multituberculates, symmetrodonts, and marsupials) (Eaton 2006a), and dinosaur tracks (Titus et al. 2005).

The John Henry Member is also fossiliferous at BRCA. As noted, there are 62 cataloged fossil sites from this member at the park. Fossils from these sites include wood and other plant fossils, bivalves (BRCA Figure 5), gastropods, ostracodes, and bones and teeth of fish, turtles, crocodilians and crocodile relatives, and dinosaurs (S. Eagan, pers. comm., October 2010). There are several published microvertebrate localities in this unit at the park. They are among the most productive localities in the member, and have produced fossils of frogs, lizards, and mammals including multituberculates, symmetrodonts, and marsupials. Most of the localities are high in the member, with one low in the member (Eaton 2005). Williamson et al. (2011) reported on screenwashed fossils from locality UMNH 1156. Specimens included freshwater ostracodes, mammals, and, unusually, a tooth of the chimaera-like marine cartilaginous petalodontiform fish *Brachyrhizodus*. Another hadrosaur skeleton has been found low in the member east of BRCA (Santucci and Kirkland 2010).



BRCA Figure 5. Corbulid bivalves at Museum of Northern Arizona (MNA) locality 1156 ("Glory Cove 2"), in the John Henry Member (JPK).

The Drip Tank Member is a fluvial cliff-forming sandstone (Peterson 1969a). In the BRCA area, this unit is a white to light gray sandstone with conglomerate beds appearing in the upper two-thirds (Tilton 2001). It interfingers with the overlying Wahweap Formation (McCord 1997). It was deposited during the middle Late Cretaceous (Eaton et al. 1999a), about 84 Ma. Few diagnostic macrofossils are known from this member (Titus et al. 2005), although wood and bones are known from three localities in BRCA (S. Eagan, pers. comm., October 2010). Regionally, its fossils include palynomorphs (Christensen and Lawton 2005), petrified logs (Peterson 1969a), fish scales (Titus et al. 2005), and fragments of turtles, crocodilians (Eaton et al. 1999a), and dinosaurs (Titus et al. 2005). In the BRCA area, it includes locally abundant log casts (Tilton 2001).

Fossils are well-known from the various members at nearby GSENM (McCord 1998). Fossils from the Tibbet Canyon Member include marine invertebrates (McCord 1997), sharks, rays, gars, crocodilians, and marsupials (Eaton and Cifelli 1997; Eaton et al. 1999a). Fossils from the Smoky Hollow Member include dicotyledonous leaf compressions (possibly in this unit) at Alvey Wash (Erwin 1997), sharks, rays, ray-finned fish such as bowfins and gars, the unusual amphibian *Albanerpeton*, frogs, turtles, lizards, several types of crocodilians and crocodile relatives, small theropods such as dromaeosaurids and troodontids, tyrannosaurids, armored dinosaurs, hypsilophodonts, hadrosaurids, and symmetrodont, marsupial, and possible eutherian mammals (Eaton et al. 1999a). Fossils in the John Henry Member include coal (Hamblin and Foster 2000), palynomorphs (McCord 1997), bivalves, ammonites (Eaton 2006a), footprints (Hamblin and Foster 2000), sharks, rays ray-finned fish like bowfin and gars, *Albanerpeton*, frogs, turtles, lizards, several types of crocodilians and crocodile relatives, dromaeosaurids, armored dinosaurs, hadrosaurids (Eaton et al. 1999a), and multituberculate, symmetrodont, and marsupial mammals (Eaton 2006a). Finally, fossils in the Drip Tank Member include turtle and crocodilian fragments (Eaton et al. 1999a).

Wahweap Formation (Upper Cretaceous)

The Wahweap Formation is a heterogeneous formation deposited on alluvial plains by fluvial and floodplain processes (Sable and Hereford 2004). At BRCA, rocks assigned to the Wahweap Formation outcrop in several areas, particularly in a relatively narrow zone following the central axis of the park, between the Cenozoic rocks to the west and the older Cretaceous rocks to the east (Bowers 1990; Davis and Pollock 2010; Biek et al. in prep.). Rocks assigned to the Wahweap Formation at BRCA by Bowers (1990) make up a sequence of buff to brown sandstone and gray to tan mudstone, 60 to 120 m (200 to 400 ft) thick. Bowers (1990) reported that Wahweap strata are 0 to about 200 m (0 to 700 feet) thick at BRCA and that the variation in thickness is due to erosional truncation under the sub-Claron Formation unconformity. The entire formation is about 360 to 400 m (1,200 to 1,300 feet) thick on the west flank of the Paunsaugunt Plateau, and about 230 m (760 feet) thick on the Markagunt Plateau (Moore and Straub 2001; Biek et al. 2011).

On the neighboring Kaiparowits Plateau, the Wahweap Formation is divided into four informal members: the lower, middle, upper, and capping sandstone members. The three members below the capping sandstone are generally composed of sandstone and mudstone, and are interpreted as having been deposited by meandering streams flowing to the north or northeast. The capping sandstone also includes conglomeratic and siltstone beds, and is thought to represent braided stream deposition, with flow to the east and southeast (Lawton et al. 2003; Titus et al. 2005). The capping sandstone includes minor windblown deposits in the Kaiparowits Basin east of BRCA (Simpson et al. 2008). Fossils are abundant in all four members

(Titus et al. 2005). Unlike underlying formations, the Wahweap Formation lacks coal (McCord 1997).

The Wahweap Formation is regarded as entirely nonmarine in deposition (Eaton 1991; McCord 1997). The Western Interior Seaway, which had been retreating out of south-central Utah during deposition of the underlying Tropic Shale and lower Straight Cliffs Formation, was absent from the BRCA area during deposition of the Wahweap Formation (Gregory 1951). The climate appears to have become more arid toward the end of deposition, with possibly semi-arid to arid conditions (Simpson et al. 2008, 2010a). Faulting occurred during deposition in the Kaiparowits Basin (Tindall et al. 2010). Sediment deposition was probably not continuous, with the capping sandstone member in particular differing from the rest of the formation. This member has disconformable contacts at its base and top (Kirkland and DeBlieux 2010). Deposition occurred after 81 Ma and before 76 Ma (Jinnah et al. 2009; Getty et al. 2010; Simpson et al. 2010b).

The Wahweap Formation is fossiliferous within BRCA. There are 44 documented localities from this formation in the park and an additional 12 localities identified as Wahweap/Kaiparowits, with fossils of bivalves, gastropods, crabs, ostracodes, fish (BRCA Figure 6), turtles, crocodilians and crocodile relatives, dinosaurs, and mammals (S. Eagan, pers. comm., October 2010). One published locality is UMNH VP Locality 77, in Campbell Canyon. This site has yielded fossils of sharks, rays, gars, bowfins, salamanders, frogs, turtles, lizards, crocodiles or crocodile relatives, dinosaurs (theropods, hadrosaurs, and possibly armored dinosaurs), multituberculates, and marsupials (Eaton et al. 1998; Munk 1998; Eaton 1999). Assorted vertebrate fossils including vertebrae, ribs, shoulder bones, and a theropod tooth were recovered by Chris Shierup of Northern Illinois University in 1999 from locations in the western Paunsaugunt Plateau, including one within BRCA (Santucci and Kirkland 2010). Nearly 2,000 gastropods have been recovered from a site near Horse Hollow at the south end of BRCA, along with bowfin vertebrae, crocodilian or crocodile relative teeth, hadrosaur teeth, and bone fragments (Williams and Lohrengel 2007). Near BRCA is the Johns Valley borrow pit, a notable paleobotanical locality in the lower Wahweap Formation within GSENM. This locality has yielded fossils of possible cycads, taxodiaceous conifers, and dicotyledonous angiosperms (Erwin 1997). There are also several UMNH and Museum of Northern Arizona (MNA) microvertebrate localities near BRCA described in the literature, originally thought to be from the Kaiparowits Formation or an equivalent. These include MNA localities 1073 (=UMNH 83) and 1074, and UMNH localities 61 and 84 (Eaton et al. 1998; Munk 1998). Fossils from these sites include the remains of gastropods, ostracodes, fish, salamanders, turtles, lizards, crocodilians or crocodile relatives, hadrosaurs, and mammals (Munk 1998).



BRCA Figure 6. Rostral teeth (left) and oral teeth (right) from a new species of freshwater sawfish from the Wahweap Formation. These fossils are examples of the material that can be found through screenwashing (figure courtesy Jim Kirkland, Utah Geological Survey).

Overall, including areas outside of BRCA, the paleontological resources of the Wahweap Formation are diverse. Microvertebrate remains are well-known from this formation. Bacteria, algae, and plants are represented by charophytes (Peterson 1969b), cryptobiotic soils (Simpson et al. 2010a), palynomorphs (Lawton et al. 2003), leaf impressions (Moore and Nealey 1993), compressed plant fossils (Kirkland et al. 2005), petrified wood (Titus et al. 2005), carbonized wood (Moore and Nealey 1993), and root traces (Simpson et al. 2008). Possible cycads, conifers, and angiosperms are among the plants that have been found (Erwin 1997). Petrified wood is common in channel-fill deposits of the various members, with specimens the size of large logs in some cases (Titus et al. 2005). Additionally, reworked palynomorphs from older formations have been found in the capping sandstone (Lawton et al. 2003). Invertebrates are represented by bivalves, gastropods, ostracodes (Peterson 1969b), crab claws (Eaton 1991), and invertebrate wood borings, which have been found in the upper and capping members (Moran et al. 2010). Mollusks sometimes form shell beds, which are most abundant in the middle mudstone-rich member (Kirkland et al. 2005).

Vertebrates of the Wahweap Formation are represented by freshwater sharks and rays, gars, bowfins, and other bony fish, salamanders, frogs, turtles, lizards, crocodilians and crocodile relatives, a variety of dinosaurs, diverse early mammals (Eaton et al. 1999a, 1999b; Titus et al. 2005; Kirkland and DeBlieux 2010), probable lungfish burrows (Simpson et al. 2008), dinosaur eggshell (Eaton 2004), crocodilian or crocodile relative tracks (Simpson et al. 2010c), dinosaur tracks (Simpson et al. 2010d), and theropod digging traces associated with mammal burrows (Simpson et al. 2010b). Many of the trace fossil occurrences are known from the capping sandstone (Simpson et al. 2010b, 2010c, 2010d). The most abundant dinosaurs are

horned dinosaurs and hadrosaurs (Titus et al. 2005), with tyrannosaurids, dromaeosaurids, troodontids, armored dinosaurs, pachycephalosaurs, and hypsilophodonts as well (Eaton et al. 1999a; Weishampel et al. 2004; Titus et al. 2005). Dinosaur bonebeds are present but rare (Kirkland et al. 2005). The first unique dinosaur taxon from the Wahweap Formation was named in 2010: *Diabloceratops eatoni*, an early horned dinosaur (Kirkland and DeBlieux 2010). Mammals have been extensively described, and include multituberculates, symmetrodonts, marsupials, insectivores, and other early forms (Eaton and Cifelli 1988, 2005; Cifelli 1990; Eaton et al. 1999a, 1999b; Titus et al. 2005; Eaton 2006b; Kirkland and DeBlieux 2010). Generally, microvertebrate remains are more common in sandstone channel lags, but there are significant microvertebrate sites in mudstone beds. The lack of articulated remains is attributed to the rarity of overbank deposits (Kirkland et al. 2005).

Claron Formation (Paleocene–Middle Eocene)

The Claron Formation of southwestern Utah has a complicated nomenclatural history involving several informal formation names and differing conceptions of what these formations encompassed. The Claron Formation was originally identified as the Wasatch Formation, better known from northern Utah and Wyoming. This is particularly apparent in the works of geologist Herbert Gregory. Additionally, in some of his publications (Gregory 1945, 1949) the unit now known as the upper white limestone member of the Claron Formation was included in the Brian Head Formation. This formation is discussed in more detail in the CEBR summary. The use of the Wasatch Formation name for these rocks persisted into the 1970s (Bowers 1972). The modern lower (sometimes called pink or red) member of the Claron Formation was informally called the Cedar Breaks Formation (Schneider 1967), which did not attain wide usage. Shortly after, the Canaan Peak Formation and the Pine Hollow Formation were named from rocks formerly assigned to the lower "Wasatch" Formation on the Table Cliffs Plateau east of BRCA (Bowers 1972). Anderson and Rowley (1975) returned the upper white limestone to the Claron Formation and included all of Gregory's original Brian Head Formation rocks in the Claron Formation. The situation increased in complexity during the 1990s. First, a third formation (the Grand Castle Formation, also discussed in more detail in the CEBR summary) was named from former lower Claron Formation rocks on the Markagunt Plateau (Goldstrand 1994; Goldstrand and Mullett 1997). Then, the Brian Head Formation was brought back into usage, limited to rocks above the Claron Formation's upper white limestone member (Anderson 1993; Sable and Maldonado 1997). While it is necessary for the researcher wishing to examine the literature on the Claron Formation to keep in mind the dramatically shifting nomenclature, fortunately the rocks themselves are usually simple to identify in publications from their descriptions: the modern lower Claron Formation is almost always identified as a distinctive pink or red unit, and the modern upper Claron Formation is almost always identified as a white limestone. Biek et al. (2011, in prep.) discussed the Claron Formation on the Markagunt and Paunsaugunt plateaus and noted that the formation is restricted to non-volcanic strata bounded two unconformities: a major Late Cretaceous–early Cenozoic unconformity at the base of the formation, and the unconformity associated with the conglomerate of Boat Mesa at the top.

The Claron Formation of BRCA is divided into a lower and upper member, with a basal red or gray conglomerate (Bowers 1990). The basal conglomerate and can be up to 14 m (45 ft) thick in the vicinity of BRCA (Tilton 2001). The lower member is composed of interbedded sandy limestone, calcareous (lime-rich) mudstone, siltstone, and some sandstone and dolomite, and minor pebbly conglomerate. It is between 120 and 210 m (400 and 700 ft) thick, and ranges in color from pink to red to tan to orange to white. Depositional settings are thought to include low-energy fluvial, floodplain, and minor lacustrine settings. The upper member is composed of white limestone and lesser amounts of calcareous mudstone, siltstone and sandstone. It is up to 90 m (300 ft) thick, and is interpreted as having been deposited in lacustrine and low-energy fluvial settings. Differential erosion of the lower member is responsible for much of the scenery at BRCA (Bowers 1990). Exposures of the Claron Formation at BRCA predominate in the western half to two-thirds of the park. Exposures of the lower member are more widespread, with the upper member mostly limited to the central part of the park (Bowers 1990; Davis and Pollock 2010).

Historically, the Claron Formation was regarded as largely lacustrine, but fluvial processes are now more widely recognized. The lower member is mostly fluvial and alluvial in depositional environment, whereas the upper white member represents both lacustrine and low-energy fluvial environments (Goldstrand 1992; Bown et al. 1997; Ott 1999; Davis et al. [2009] erroneously regarded both members as shallow lacustrine). The lake transitioned from closed to open drainage around 42 Ma (Davis et al. 2009). Laramide-age uplifts to the northeast and northwest contributed sediment to the formation (Goldstrand 1992). The Claron Formation has been heavily affected by pedogenic (soil-forming) processes, which are thought to have destroyed most of its original depositional features and fossils (Mullett et al. 1988; Mullett 1989; Feist et al. 1997). Plant growth is interpreted as moderate, with moist soils that underwent periodic drying (Bown et al. 1997). The reddish coloration of the lower member is from iron staining, while the white of the upper member is thought to have resulted from leaching (Mullett 1989). Dissolution of Claron Formation limestone can produce sinkholes, which are abundant on the Markagunt Plateau to the west (Moore et al. 2004), and shallow caves (Hatfield et al. 2010; Spangler 2010), but these are rare on the Paunsaugunt Plateau. The formation weathers into thousands of rock chimneys at BRCA (Davis and Pollock 2010).

The age of the Claron Formation has been difficult to establish, because of its paucity of fossils, lack of volcanic debris, and the fact that it is bounded by unconformities; it is now widely

considered to be Paleocene to latest middle Eocene in age (Goldstrand 1994; Eaton et al. 2001; Biek et al. 2011, in prep.). The recent discovery of a rodent jaw in the middle of the upper (white) member indicates that it includes rocks of middle Eocene age (Eaton et al. 2011). The upper age limit of the Claron Formation is also constrained by overlying volcaniclastic strata of the Brian Head Formation, which have been dated to 36 Ma (Biek et al. 2011). Older reports of Oligocene-age rocks in the Claron Formation (Bowers 1990) are based on rocks now included in the Brian Head Formation.

Fossils have historically been difficult to find in the Claron Formation, because of pedogenic and diagenetic processes. The most abundant fossils are microbial structures, mollusks, and arthropod traces. Microbial body fossils and traces from the Claron Formation include oncolites, stromatolites (Taylor 1993), and charophytes (Moore and Nealey 1993; Moore et al. 2004). These fossils are mostly found in the upper member (Moore and Nealey 1993; Moore et al. 2004). Plant fossils include palynomorphs (Goldstrand and Mullett 1997; Foster et al. 2001), leaf impressions (Gregory 1950), seeds (Schneider 1967), and root traces (Mullett et al. 1988; Mullett 1989; Taylor 1993). Invertebrate body fossils include small bivalved animals that could be bivalves, conchostracans (clam shrimp), or ostracodes (Moore and Nealey 1993), bivalves, gastropods (Gregory 1950), and ostracodes (Brouwers et al. 2000; Eaton et al. 2011). Shelled invertebrates may be found as molds or casts (Sable and Hereford 2004). Invertebrate trace fossils include ant nests, wasp and bee cocoons and brood cells with possible parasite traces, dung beetle nests (Bown et al. 1997), and crayfish burrows (Hasiotis and Bown 1997). The insect traces have been found in the clastic rocks of the lower member (Bown et al. 1997). Vertebrates are represented by fish bones and a fragmentary rodent jaw (Eaton et al. 2011).

Fossils are known but have not been described from the Claron Formation of BRCA, although a popular science article by Vokes (1952) reports freshwater bivalves and gastropods in limestone and lime-rich sandstone beds of the park's upper cliffs and slopes, and specimens of the gastropod *Physa* have been found in the park, often located by park staff or visitors. Typically these snail shells are found as float (loose specimens) along trails and near overlooks (Kenworthy 2006). Four localities in the Claron Formation have been cataloged from BRCA, including bivalves, trace fossils, and a large bone (S. Eagan, pers. comm., October 2010; the bone is possibly dinosaurian and likely reworked from an older unit, given the apparent absence of bones elsewhere in the formation). Gastropods have been recovered from the lower member through screenwashing (J. Eaton., pers. comm., September 2011). Gastropods are known from the upper member on Whiteman Bench just west of BRCA (Bowers 1990). At CEBR, the lower Claron Formation has microbial structures, seeds (possibly of hackberry or the genus *Prunus*, which includes plants that produce almonds, cherries, peaches, and plums), and possible invertebrate burrows (Schneider 1967). The upper Claron Formation of CEBR has gastropods and insect trace fossils (Gregory 1950).

Conglomerate at Boat Mesa (Upper Eocene)

There are several exposures of a conglomeratic unit present in north-central BRCA, known informally as the conglomerate at Boat Mesa or the Boat Mesa Conglomerate. It is a light-colored capping conglomerate between 12 and 30 m (40 and 100 ft) thick. The lower contact with the Claron Formation is a disconformity. The conglomerate lacks fossils that were contemporaneous with deposition, so its age can only be estimated indirectly. Lacking age control, Bowers (1990) and Davis and Pollock (2010) considered it to be Oligocene in age. However, Biek et al. (2011, in prep.) reported a U-Pb age on zircon of 36 Ma from the overlying Brian Head Formation on the southwest Sevier Plateau, immediately northwest of BRCA, and Eaton et al. (2011) reported middle Eocene mammal fossils in the upper member of the Claron Formation southwest of BRCA; therefore, the conglomerate at Boat Mesa must be late Eocene in age. The unit is interpreted as representing fluvial and overbank deposits (Davis and Pollock 2010). It has yielded reworked Permian invertebrates from chert clasts within BRCA (Santucci and Kirkland 2010).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within BRCA. However, they are known to preserve fossils elsewhere, and future field investigations within the park may recover fossils from one or more of them.

Kaiparowits Formation (Upper Cretaceous)

The Kaiparowits Formation is best known from GSENM (Hutchinson et al. 1997). Although earlier maps commonly included the Kaiparowits Formation as part of a lumped Wahweap–Kaiparowits unit on the Markagunt and Paunsaugunt plateaus (for example, Gregory [1950] and Sable and Hereford [2004]), many researchers failed to find Kaiparowits Formation strata on the Paunsaugunt Plateau (for example, Bowers [1990] and Titus et al. [2005]). However, the Kaiparowits Formation is now known to be present in the northeast part of BRCA (Biek et al. in prep.) and is also newly identified on the west flank of the Paunsaugunt Plateau (Biek et al. 2011).

The Kaiparowits Formation is roughly divisible into sandstone-dominated lower and upper intervals below and above a mudstone-rich middle interval (Getty et al. 2010). In the western Paunsaugunt Plateau, only the lower part of the formation is exposed. A generalized section from the western part of the plateau includes a basal mudstone up to a few meters thick; then

an interval of poorly cemented fine-grained bluish-gray or greenish-gray to gray sandstone, which takes up about two-thirds of the section; and then an upper interval of grayish mudstone and siltstone. The formation typically weathers into badlands, and is up to 120 m (200 ft) thick on the plateau, but can reach a thickness of 860 m (2,820 ft) in the Kaiparowits Basin (Biek et al. 2011).

The Kaiparowits Formation is interpreted as representing a broad alluvial floodplain (Hutchinson et al. 1997), tens to hundreds of km or miles from the shoreline (Roberts et al. 2008). Depositional environments include a variety of channel settings, floodplains, lakes, and swamps. A period of sea-level rise during deposition resulted in some tidal and estuarine influence (Roberts 2007). Plant fossils indicate a relatively warm and wet climate (Roberts 2007). The formation was deposited in response to the Sevier orogeny and related basin-forming (Getty et al. 2010). Streams flowed to the east and northeast from sediment sources in Sevier uplifts to the west and southwest (modern-day California and Nevada) (Goldstrand 1994). Sedimentation rates were high (41 cm [16 in] per 1,000 years) (Roberts 2007). Deposition occurred between approximately 76.2 and 74.1 Ma (Roberts et al. 2008).

The Kaiparowits Formation has not yet been surveyed for fossils at BRCA, but is abundantly fossiliferous in the vicinity, and so has a high probability of being fossiliferous at the park. Gastropods are locally abundant in the mudstone interval in the western Paunsaugunt Plateau (Biek et al. 2011). Regionally, it is known for its rich terrestrial assemblage (Titus et al. 2005). It has yielded diverse floral remains (Getty et al. 2010), such as algal cysts (Nichols 1997), spores and pollen (from mosses, ferns, clubmosses, cycads, conifers, and angiosperms; Lohrengel 1969), leaves, wood, root traces, and amber (Roberts 2007). Invertebrates are represented by freshwater or brackish water bryozoans, bivalves, gastropods (Roberts 2007; Roberts et al. 2008), ostracodes (Roberts and Tapanila 2006), and several types of trace fossils, including burrows, social insect nests (Roberts and Tapanila 2006), wood boring traces (Moran et al. 2009), and insect borings into dinosaur bones (Roberts et al. 2007). There is an example of a shell bed including at least 1.4 million unionid bivalves in an estuarine system (Roberts et al. 2008). Finally, the vertebrate assemblage of the Kaiparowits Formation includes sharks and rays, bony fish (including gars, bowfins, sturgeons, and others), salamanders, frogs, turtles, lizards, snakes, crocodilians and crocodile relatives, dinosaurs, the bird *Avisaurus*, early mammals (including multituberculates, symmetrodonts, marsupials, insectivores, and others) (Hutchinson et al. 1997; McCord 1997; Eaton et al. 1999a; Titus et al. 2005), and eggshells (Bray 2002; Roberts 2007). Dinosaurs are common, and include tyrannosaurids, ornithomimids (ostrich-mimics), dromaeosaurids, troodontids, oviraptorosaurs ("egg thieves"), armored dinosaurs, pachycephalosaurs, horned dinosaurs, hypsilophodonts, and hadrosaurs (Eaton et al. 1999a; Titus et al. 2005). Several new dinosaur taxa have been recovered (Zanno and Sampson 2005; Gates and Sampson 2007; Boyd et al. 2009; Getty et al. 2010; Sampson et al. 2010; Zanno et al. 2011). Crocodilians and crocodile relatives are also diverse (Wiersma et al. 2004; Titus et al. 2008). Roberts (2007) provides a detailed discussion of the formation's depositional environments and associated fossils.

Upper Cenozoic alluvial deposits (Miocene?–Pliocene?)

Late Tertiary alluvial-fan and volcaniclastic deposits, previously known as the Sevier River Formation, are exposed in several areas of southwestern Utah (Anderson and Rowley 1975; Rowley et al. 1975; Biek et al. 2011, in prep.). They are the same as Gregory's (1945, 1950) Parunuweap Formation (Anderson and Rowley 1975). Within BRCA, these deposits are found only in the extreme northeastern part of the park (Bowers 1990), although they are widespread on the Paunsaugunt and Markagunt plateaus (Biek et al. 2011, in prep.). The geologic map included with Thornberry-Ehrlich (2005) depicts the "Sevier River Formation," but the text omits it. In the BRCA area, these deposits are composed of light-brown to brownish-gray conglomeratic sandstone and conglomerate (Bowers 1990), interpreted as remnants of valley fill deposits left by the ancestral Sevier River system (Davis and Pollock 2010). The deposits also incorporate volcanic fragments and limestone pebbles (Bowers 1990). The sediments were deposited mostly as alluvial fans and fine-grained basin-fill (Gregory 1944, 1945), with minor fluvial and lacustrine deposits (Anderson and Rowley 1975; Rowley et al. 1975).

The age of the deposits at BRCA has not been established, but they are likely equivalent to latest Miocene to early Pliocene alluvial-fan deposits mapped on the east flank of the Markagunt Plateau (Biek et al. 2011). Fossils are not yet known from the BRCA-area alluvial deposits. Sparse fossils, mostly of algae and invertebrates, are known from similar deposits in southwestern Utah, but the stratigraphic relationships of these deposits with those at BRCA are not yet clear. The closest examples among these are Pliocene-age diatoms (a type of alga that secretes silica "shells") from lake deposits near Hillsdale, 22 km (14 miles) west of BRCA (Crawford 1951; Setty 1963).

Quaternary sediments (Pleistocene-Holocene)

The Quaternary deposits of BRCA include a variety of alluvial, colluvial, landslide, and pediment deposits, with grain sizes ranging from clay to boulders. The thicknesses of these units can be as great as 30 m (100 ft) or more. Deposits are found scattered throughout the park (Bowers 1990). Fossils are not currently known from these deposits at BRCA, but may eventually be found. Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Summaries of Quaternary fossil sites in southern Utah and the vicinity include Mulvey et al. (1984), Agenbroad and Mead (1989), Mead and

Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (1999, 2000), and Coats et al. (2008).

Quaternary fossils are not as well-known from the BRCA region as their older counterparts. Finds have included a handful of pond or former pond localities with pollen and similar material, and wood and freshwater bivalves and gastropods from lake and other settings. Pleistocene or Holocene pond sediments are not present in any great quantities at BRCA. More probable paleoecological fossils for BRCA include middens and phytoliths (microscopic bits of minerals found in the cells of some plants): ancient packrat middens are potential finds in alcoves and other protected recessed areas, while phytoliths could be present in the soil, as they are at Capitol Reef National Park (CARE; see the CARE summary in this document for more details) (Fisher et al. 1995). Should pond or bog fossils be found at BRCA, they would probably be comparable to lake finds from the vicinity, which include organic debris, peat, charcoal, spores, pollen, various macroscopic plant fossils (such as needles, cones, and wood), and aquatic insects (Mulvey et al. 1984; Anderson et al. 1999; Weng and Jackson 1999; Madsen et al. 2002). One such site is known from CEBR (Alpine Bog), with another (Lowder Creek Bog) just outside of the monument (Anderson et al. 1999). Fossiliferous Quaternary lake deposits are known from several places in Zion National Park (ZION; see the ZION summary in this document for more details), about 55 km (34 miles) west of BRCA (Hamilton 1979; Hevly 1979; Biek et al. 2009). Fossils of large mammals are probably more likely to be found just outside of BRCA in deposits in the vicinity of the Sevier River than they are to be found within the park; Agenbroad (1997) noted the presence of a number of large mammal finds on the Colorado Plateau, but most of these are isolated bones from terrace deposits near large rivers. The most commonly-found large mammals are sloths, mammoths, horses, tapirs, camels, bison, oxen, and mountain goats (Agenbroad 1997). Another possibility for river settings are mollusks; Pleistocene beds in the Red Creek Valley on the Paunsaugunt Plateau have freshwater bivalves and gastropods (Gregory 1951). Reports of a mammoth tooth or jaw fragment from BRCA are in error, as the specimen in question is actually a hadrosaur jaw fragment with exposed columns of teeth (BRCA Figure 7) (Kenworthy 2006).



BRCA Figure 7. Partial hadrosaur jaw previously mistaken for a mammoth tooth (JPK).

Park Collections

As of October 2006, there were 541 catalogued BRCA paleontological specimens, 225 of which were curated at BRCA (6 deaccessioned) (Kenworthy 2006). Since that time, BRCA collections have been moved to ZION (S. Eagan, pers. comm., October 2010). Of the others, 298 specimens were in the collections of the Utah Museum of Natural History (Salt Lake City, UT), and 18 were in the collections of the Museum of Northern Arizona (Flagstaff, AZ) (Kenworthy 2006). For example, the fossils from UMNH VP 77 are curated at the Utah Museum of Natural History (Eaton 1999), and were studied at Weber State University (Ogden, UT) (Munk 1998). As of 2006, most paleontological specimens were going to be transferred from BRCA to the UMNH (Kenworthy 2006).

Cultural Resource Contexts

Humans have been present in south-central Utah for at least 11,000 years (Agenbroad 1997), so it would not be surprising to find paleontological resources at BRCA associated with or as cultural resources, such as fossils taken for cultural purposes or used as tools. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for BRCA during July 1999 (NPS 1999), and produced a geologic resource evaluation report for the park in 2005 (Thornberry-Ehrlich 2005). The participants of the scoping meetings developed a series of recommendations for future work, including: studying the Cretaceous rocks to determine the presence or absence of dinosaurs in the backcountry; the completion of a comprehensive inventory and study of park paleontological resources; determination of the locations of paleontological specimens removed from the park to private collections; and drawing visitor attention to park fossil resources with graphics, brochures, and exhibits. With the impending completion of the six-year collaborative study, an assessment of the state of these goals would be useful. Kenworthy (2006) provides some additional recommendations.

- BRCA reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 78 localities, 75 of which were reported to be in good condition in 2009. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- Fossil site vandalism and fossil theft are potential issues for BRCA. It is known to occur in the area: petrified logs are sometimes stolen from the Wahweap Formation of GSEMN, for example (Kirkland et al. 2005).

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Data Sets

Bryce Canyon National Park Paleontological Archives
5/1985–present. (hard copy data; reports; electronic data; photographs; maps; publications). Originated by Santucci, Vincent; status: Active.

Bryce Canyon National Park Collections
1923–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by BRCA staff; status: Active.

Bryce Canyon National Park Digital Geologic Map
2010–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=2164782.

Museum of Northern Arizona
1990s–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by MNA staff; status: Active.

Paleontological Survey of Late Cretaceous Sections, including the Kaiparowits Formation
1990s–present. (field notes; hard copy data; photographs). Originated by Eaton, Jeff, Weber State University, Ogden, UT; status: Active.

Utah Museum of Natural History
1990s–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by UMNM staff; status: Active.

Weber State University
1988–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by WSU staff; status: Active.

Zion National Park Collections
2000s–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by ZION staff; status: Active.

Canyonlands National Park

Canyonlands National Park (CANY), Utah, was established September 12, 1964. It protects a canyon-cut landscape of rock, with spires and mesas between the gorges and valleys (CANY Figure 1). Archeological remnants of prehistoric American Indian occupation are also present. The boundaries of CANY changed November 12, 1971. It currently encompasses 136,620.99 ha (337,597.83 acres), all but 11.09 ha (27.40 acres) of which are under federal administration. CANY consists of two units, a main unit and the small detached Horseshoe Canyon unit, roughly 11 km (7 miles) west of the main unit (CANY Figure 2). CANY is part of the Southeast Utah Group of the National Park Service (NPS), headquartered in Moab, Utah. The Southeast Utah Group also includes Arches National Park (ARCH), Hovenweep National Monument (HOVE), and Natural Bridges National Monument (NABR); summaries for these parks and monuments are also included in this report.

Geologic Background

CANY is located mostly in northwest San Juan County, Utah, with smaller parts in Wayne, Garfield, and Grand counties. The park shares its western border with Glen Canyon National Recreation Area (GLCA) and is adjacent to Canyon Rims BLM Recreation Area on the east and Dead Horse Point State Park on the northeast. ARCH is about 18 km (11 miles) to the northeast, and Moab, the nearest large community, is in the same area 19 km (12 miles) to the northeast. Western CANY is dominated by the drainage of the Green River, and eastern CANY is dominated by the Colorado River. The two rivers converge in south central CANY. These rivers can be used to divide the main unit of CANY into three districts: south and west of the Green River and Colorado River is The Maze District; south and east of the Colorado River is The Needles District; and between the two rivers, in central and northern CANY, is the Island in the Sky District. Geographically, CANY is east of the San Rafael Desert, northeast of the Henry Mountains, west of the La Sal Mountains, and northwest of the Abajo Mountains.



CANY Figure 1. A view into one of CANY's canyons (VLS).

The geology of CANY and the vicinity has been described in hundreds of documents (note that "Canyonlands" is also used for the CANY area in general, so a reference to Canyonlands may describe the park itself or the surrounding region). References that are useful for understanding the overall geology of CANY include Baars (1971, 1983, 1993, 2004, 2010), Rigby et al. (1971), Sumsion and Bolke (1972), Lohman (1974), Huntoon et al. (1982, 2002), Dubiel et al. (1985), Huntoon and Dubiel (2000), Billingsley et al. (2002), and KellerLynn (2005). References with particular bearing on the paleontology of CANY include Rigby and Stokes (1971), Steele-Mallory (1982), Biggar and Adams (1987), Elias et al. (1992), Hasiotis (1991, 1992, 1993, 1995), Lucas (1995), Lucas et al. (1995), Elias (1997), Santucci et al. (1998), Heckert et al. (1999), Sumida et al. (1999a, 1999b), Anderson et al. (2000), Koch and Santucci (2002), Reheis et al. (2005), Coats et al. (2008), and Santucci and Kirkland (2010). The Geologic Resources Division of the NPS coordinated a geologic resource inventory scoping session for CANY, ARCH, HOVE, and NABR during May 1999 (NPS 1999a), and produced a geologic resource evaluation report for CANY in 2005 (KellerLynn 2005). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The geologic record exposed at CANY covers the Middle Pennsylvanian to the Middle Jurassic, and the Quaternary (see the appendix for a geologic time scale). The oldest exposed rocks are more than 300 million years old and date to the middle and late Pennsylvanian. At this time, CANY and much of southeastern Utah and southwestern Colorado were within the Paradox Basin, a large structural basin, and the Pennsylvanian rocks are part of the basin fill. These rocks, belonging to the Hermosa Group, were deposited during a Paleozoic ice age (Rueger 1996). No glaciers were present near the Paradox Basin, but the rocks do preserve sea level fluctuations caused by glacial/interglacial cycles, which were in turn influenced by Milankovitch orbital cycles (Blakey 2008), as are documented for the recent glacial and present interglacial periods. Sea level fluctuations are well-documented in the Paradox Basin, which at the time was a deep basin rimmed by reefs during sea level highstands, with deposition of vast amounts of salt near the center of the basin during lowstands (Blakey 2008). The salt deposits would later have important implications for the structural geology and modern topography of CANY and the vicinity (Baars 1983; Walsh and Schultz-Ela 2003; Gutierrez 2004).

At the beginning of the Permian, around 300 Ma (million years ago), a period of mixed terrestrial and marine deposition

represented by the Cutler Group followed the shallow marine conditions. The Pennsylvanian-age Hermosa and largely Permian-age Cutler groups were deposited during a period of tectonic activity, with the Paradox Basin sinking while the Uncompahgre Uplift formed to the east and provided sediment to fill the basin (Melton 1972; Blakey 2008; Baars 2010). The area including CANY was equatorial at this time (Huntoon et al. 2002). Deposition occurred in coastal marine and terrestrial settings as a shallow sea transgressed (advanced over the land) during the Early Permian (Condon 1997; Baars 2000). Large coastal ergs (giant dune fields) developed on the down-wind sides of the sea (McKee 1979; Blakey 2008; Blakey and Ranney 2008) as it expanded and contracted several times (Huntoon and Kamola 1988). A hiatus of at least 25 million years follows the Early Permian rocks at CANY.

Triassic rocks at CANY are principally represented by the Moenkopi Formation of the Early–Middle Triassic and the Chinle Formation of the Late Triassic. Southern Utah during the Early and Middle Triassic was a broad coastal plain that was occasionally flooded by a sea to the west (Stewart et al. 1972a; Blakey et al. 1993; Hintze and Kowallis 2009). The top of the Moenkopi Formation is separated from the Chinle Formation by millions of years, with intervening erosion (Pipiringos and O'Sullivan 1978). The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time (Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003).

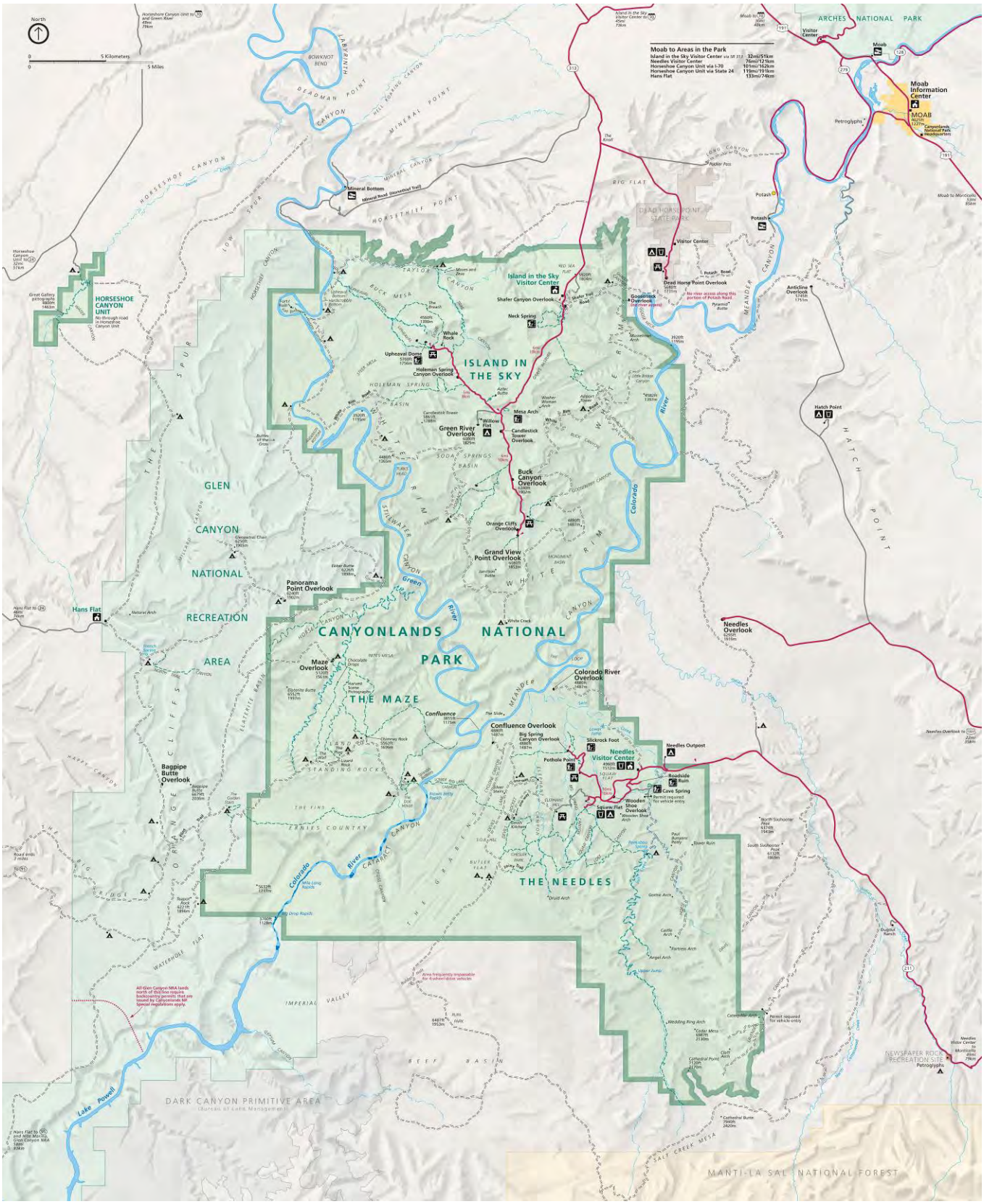
Chinle Formation rocks show that the area gradually became drier over time as the supercontinent Pangaea moved north. During its deposition, the Four Corners area of the Southwest moved from near the equator to arid-zone latitudes. What had been a setting of rivers, lakes, and deltas turned arid (Tanner 2003). This is represented at CANY by the first of several giant ergs, the Wingate Sandstone of the Glen Canyon Group (Blakey and Ranney 2008). The Wingate Sandstone erg of the Late Triassic and Early Jurassic was ended by a more humid period recorded in the Kayenta Formation. The Kayenta Formation represents a broad, gentle flood plain separated from the sea to the west (Blakey 1994; Peterson 1994). About 190 Ma, this setting was gradually replaced by the vast erg that became the Navajo Sandstone (Harshbarger et al. 1957; Dickinson and Gehrels 2003, 2009; Rahl et al. 2003; Lucas et al. 2005), the largest erg known from North America (Blakey 2008; Blakey and Ranney 2008). This Sahara-like desert was found throughout the Four Corners area into Nevada and Wyoming (Hintze and Kowallis 2009). At this time, the latitude of the area was only about 10° N, near the western edge of the supercontinent Pangaea (Loope et al. 2004a). Strong cross-equatorial winds blew from the northwest during the winter, driving dune movement, and monsoonal rains occurred during the summer (Loope et al. 2004a, 2004b). Eventually another shallow sea began to encroach on the CANY area from the north and west, inundating the Navajo erg before being replaced by another erg (Blakey et al. 1983). However, except for a short part of the post-Navajo Sandstone events, the next approximately 170 million years are unavailable for direct sampling at CANY as rocks from this span of time are not present in the park.

Although CANY and the Colorado Plateau as a whole show the effects of the Late Cretaceous–Eocene Laramide Orogeny (mountain-building event) and later events, a substantial part of CANY's structural geology derives from the influence of salt at depth from the Paradox Basin (Baars 2010). Salt dissolution and flow, for example, are responsible for The Grabens in The Needles District (Baars 2010), and salt processes may be responsible for Upheaval Dome (Jackson et al. 1998; Baars 2010), although there is also much evidence that it represents an impact crater instead (Kreins et al. 1999; Huntoon 2000; Buchner and Kenkmann 2008; Baars 2010). The Circle Cliffs and Monument uplifts and the Kaiparowits and Henry Mountains structural basins are examples of local Laramide structures (Anderson et al. 2010). Extensive downcutting followed recent uplift, leading to the namesake canyons of CANY. In addition to periods of uplift, local igneous activity occurred on the Colorado Plateau during the late Eocene, Oligocene, and Miocene. Igneous bodies emplaced at depth formed the nearby Abajo, Henry, and La Sal mountain ranges between about 31 and 23 Ma (Nelson et al. 1992; Nelson 1998; Nelson and Davidson 1998).

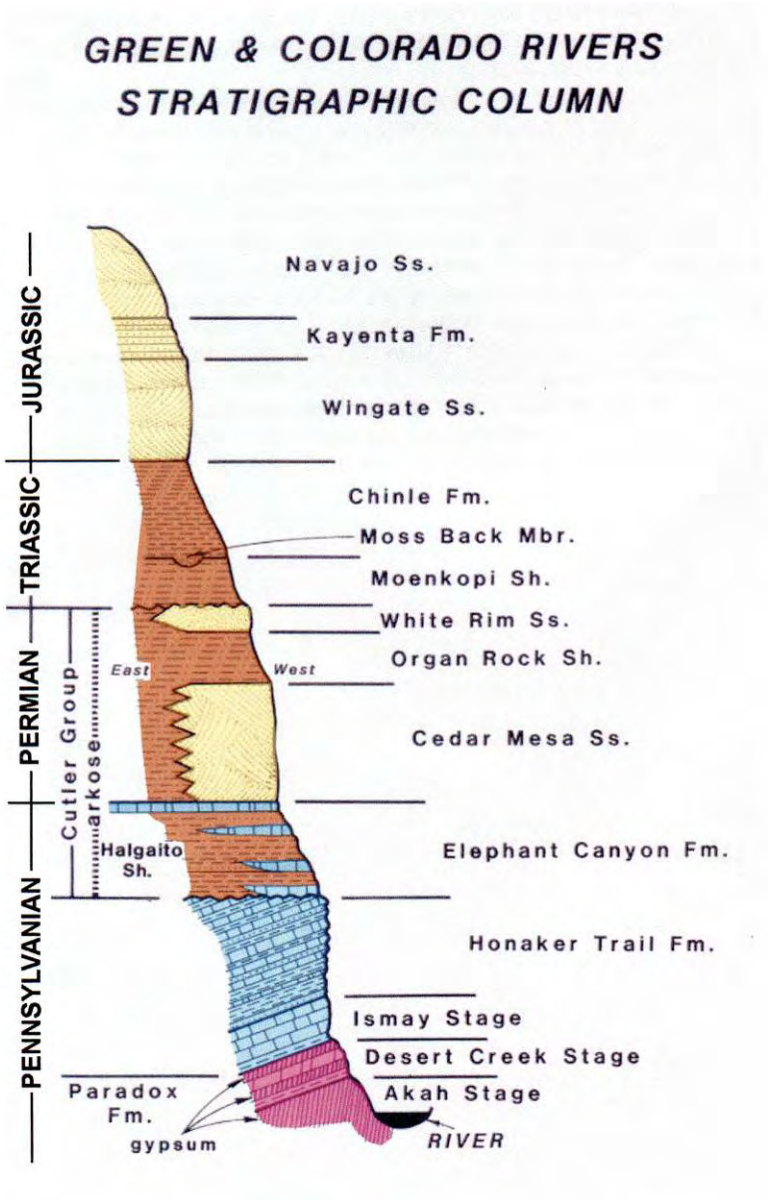
The long post-Jurassic depositional hiatus at CANY ended sometime during the Pleistocene, preserving evidence of the biota that lived there during the Ice Age. The megafaunal extinctions on the Colorado Plateau claimed such animals as the mammoth, the ground sloth *Nothrotheriops shastensis*, the shrub ox *Euceratherium collinum*, and the mountain goat *Oreamnos harringtoni*. Dated dung deposits from the Colorado Plateau indicate that they went extinct between 11,800 and 11,000 radiocarbon years ago (Mead and Agenbroad 1992). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate. The Colorado Plateau was cooler and more moist during the last glacial period, becoming more arid afterward. It has been most arid during the last 1,500 years (Elias et al. 1992). It had been thought that both megafauna and human cultural sites more than 6,000 years old were rare on the Colorado Plateau, but this is now known to be inaccurate (Agenbroad 1997). The human record at CANY includes at least the last 10,000 years, but there is little physical evidence for early human activities in the park except for rock art (NPS no date). There is evidence for two main periods of occupation: approximately 850 or 900 AD, by the Fremont people; and from 1075 to the late 12th century, by the Ancestral Puebloans (also known as the Anasazi). The Fremont people left

little in the way of artifacts, but the Ancestral Puebloans produced numerous structures in the area (Lohman 1974).

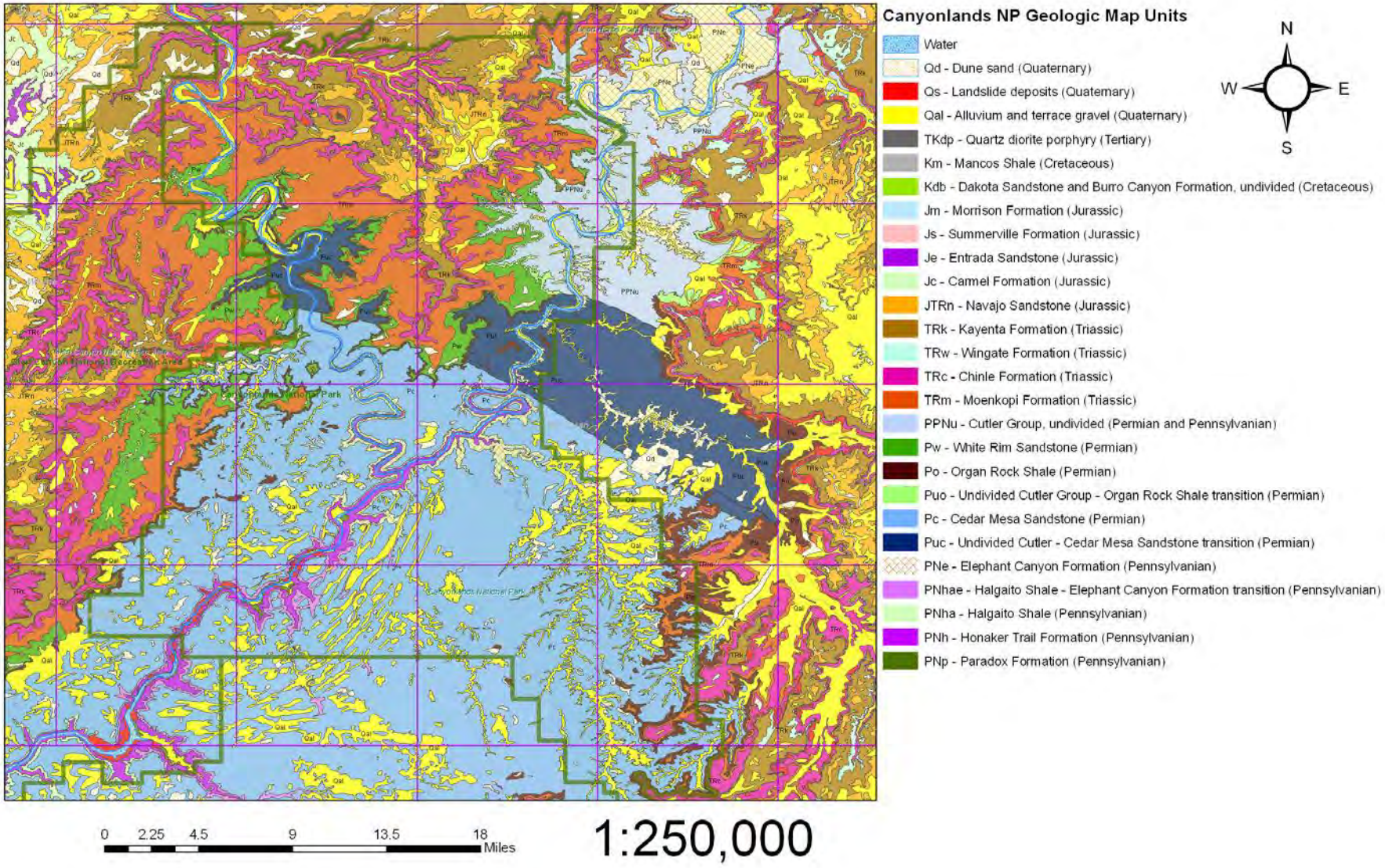
Rock units exposed at the main unit of CANY include, from oldest to youngest: the Paradox Formation and Honaker Trail Formation of the Hermosa Group (Middle–Upper Pennsylvanian); the Elephant Canyon/Halgaito Formation/lower Cutler beds, Cedar Mesa Sandstone, Organ Rock Formation, and White Rim Sandstone of the Cutler Group, as well as undivided rocks of the Group (Upper Pennsylvanian–Lower Permian); the Moenkopi Formation (Lower–Middle Triassic); the Chinle Formation (Upper Triassic); the Wingate Sandstone, Kayenta Formation, and Navajo Sandstone of the Glen Canyon Group (Upper Triassic–Lower Jurassic); and a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene). The Horseshoe Canyon Unit has outcrops of the Wingate Sandstone, the Kayenta Formation, the Navajo Sandstone, the Carmel Formation of the San Rafael Group (Middle Jurassic), the Entrada Sandstone of the San Rafael Group (Middle Jurassic; not currently recognized on a map, but noted as present by G. Cox, CANY archeological technician [pers. comm, March 2010]), and Quaternary deposits (see CANY Figure 3 for a stratigraphic column and CANY Figure 4 for a geological map) (Huntoon et al. 1982). The Paleozoic units make up most of the park's outcrops, especially in The Needles and The Maze districts, with vast outcrops of the Cedar Mesa Sandstone. Triassic and Jurassic rocks dominate the Island in the Sky District (Huntoon et al. 1982). Paleontological resources at CANY are known from the Honaker Trail Formation (marine fossils), Elephant Canyon Formation (marine fossils), Cedar Mesa Sandstone (marine fossils and tetrapod tracks), Organ Rock Formation (trace fossils), White Rim Sandstone (trace fossils and a crinoid fragment), undivided Cutler Formation (tracks), Moenkopi Formation (marine fossils, trace fossils, and plant debris), Chinle Formation (petrified wood, plant debris, shells, bones, and tracks), Kayenta Formation (tracks), Navajo Sandstone (trace fossils), and Quaternary deposits (plant matter, pollen, mollusks, insects, bones, and packrat middens) (CANY Table 1).



CANY Figure 2. Geography of CANY (NPS).



CANY Figure 3. Stratigraphic column for CANY (from Baars [2010], used with permission of the Utah Geological Association).



CANY Figure 4. Schematic geological map of CANY. Map graphic adapted from digital geologic map data available at the following URL: http://nrinfo.nps.gov/Reference.mvc/Profile?Code=1038974.

Formation	Age	Fossils Within CANY	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Pollen and packrat middens	Alluvial, eolian, fluvial, and landslide deposits
Entrada Formation	Middle Jurassic	None to date; fossils are very rare in this formation in general	Desert with large eolian sand dunes
Carmel Formation	Middle Jurassic	None to date; marine invertebrates are most likely	Marginal marine, including tidal flat and salt flat settings
Navajo Sandstone	Early Jurassic	Stromatolites and other microbial structures, petrified wood, root casts, ostracodes, and vertebrate burrows	Desert with large eolian sand dunes
Kayenta Formation	Early Jurassic	Dinosaur tracks	Fluvial, floodplain, and eolian settings
Wingate Sandstone	Late Triassic–Early Jurassic	None to date; vertebrate tracks are most likely	Desert with large eolian sand dunes
Chinle Formation	Late Triassic	Petrified wood, *Neocalamites*, *Pagiophyllum*, other plants, gastropods, bivalves, conchostracans, crayfish, invertebrate burrows (from crayfish, worms, and insects), a horseshoe crab resting trace, bones of semionotid and turseodid fish, metoposaurs, and phytosaurs, a partial skeleton of the aetosaur *Stagonolepis*, indeterminate reptile bones, and tetrapod tracks	Fluvial, floodplain, lacustrine, and eolian settings, becoming drier over time
Moenkopi Formation	Early–Middle Triassic	Natural casts of brachiopods, bivalves, ammonites, and gastropods	Coastal plain with a shallow marine incursion
Cutler Group undivided	Late Pennsylvanian–Early Permian	Possible amphibian tracks	Fluvial, alluvial fan, floodplain, and marine settings
White Rim Sandstone	Early Permian	Microbial laminations, root traces, a crinoid fragment, invertebrate traces, and general bioturbation	Coastal sand dunes with both eolian and marine influences
Organ Rock Formation	Early Permian	Root traces and burrows	Eolian sand dunes and fluvial channels
Cedar Mesa Sandstone	Early Permian	Foraminifera, bryozoans, brachiopods, crinoids, and vertebrate tracks	Coastal eolian sand dunes with occasional marine incursions
Elephant Canyon Formation / Halgaito Formation / lower Cutler beds	Late Pennsylvanian–Early Permian	Foraminifera, petrified wood, bryozoans, brachiopods, bivalves, cephalopods, gastropods, scaphopods, ostracodes, trilobites, and fish vertebrae	Coastal delta settings
Honaker Trail Formation	Middle–early Late Pennsylvanian	Algae, foraminifera, corals, bryozoans, brachiopods, bivalves, gastropods, trilobites, and crinoids	Open marine

Formation	Age	Fossils Within CANY	Depositional Environment
Paradox Formation	Middle Pennsylvanian	None to date; pollen, spores, and marine invertebrates are most likely	Restricted marine, cycling from normal marine water to evaporative conditions (a "dead sea")

CANY Table 1. Concise tabulation of CANY stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of CANY present opportunities for education, interpretation, and continued or future scientific research in the park.

Hermosa Group: Honaker Trail Formation (Middle–lower Upper Pennsylvanian)

The Honaker Trail Formation and the overlying stratigraphic interval (Elephant Canyon Formation/Halgaito Formation/lower Cutler beds) have gone through several interpretations (see discussions in Condon [1997] and Anderson et al. [2010]). Briefly, the Hermosa Formation was raised to the Hermosa Group and divided into formations in 1958, including the Paradox Formation and overlying Honaker Trail Formation (Wengerd and Matheny 1958). The definition and extent of the main part of the Honaker Trail Formation is broadly accepted within the CANY area, whereas the upper part of the formation and overlying strata have been controversial and the subject of several name changes. This interval had been known as the Halgaito Formation or the Rico Formation (an obsolete term of historical importance; Anderson et al. [2010]), but was redefined and named the Elephant Canyon Formation with the type locality at CANY's Elephant Canyon in 1962, with some of the lower portion assigned to the Honaker Trail Formation (Baars 1962). These new names were not universally adopted (Williams 1964; Rigby et al. 1971; Lohman 1974). The clearest definitions of the nomenclature were made in Loope et al. (1990), Baars (1991), Condon (1997), and Anderson et al. (2010). Depending on how one interprets the type section of the Elephant Canyon Formation, the debated strata have been called the upper Hermosa Formation and conformably overlying lower Cutler beds (Loope et al. 1990), the upper Honaker Trail Formation and conformably overlying lower Cutler beds (Condon 1997), or the Honaker Trail Formation and overlying Elephant Canyon Formation, separated by an angular unconformity (a contact where a gap in time is combined with an event that tilted the underlying beds) (Baars 1991). Additionally, because names and defined contacts differ significantly among workers, the relationships of the units are not one-to-one, so references using one system of nomenclature cannot necessarily be translated by substituting another. Because the NPS GRE publication (KellerLynn 2005) and digital geologic map use the Honaker Trail/Elephant Canyon nomenclature, they are preferred here for most of CANY in the interest of continuity. By whichever names are used, the strata between the underlying Paradox Formation and the overlying Cedar Mesa Sandstone at CANY are extensively fossiliferous and distinct.

The Honaker Trail Formation of CANY is 320 to 460 m (1,050 to 1,500 ft) thick, and is composed of dark limestones, cherty limestones, and blue, gray, and red sandstones and shales, forming ledges and slopes (Huntoon et al. 1982). It forms rugged cliff walls on Cataract Canyon. Some workers maintain that the upper surface is a disconformity with the Elephant Canyon Formation (Baars 2010), while others workers dispute the disconformity (Loope, 1990; Condon, 1997). It was deposited during the Middle to early Late Pennsylvanian (Baars 2010), approximately 306 Ma. It is mostly marine in origin (Dubiel et al. 1989; Baars 2010). Sediment from the Uncompahgre Uplift to the east was deposited as the Paradox Basin subsided (Melton 1972; Doelling 2010), during this, the most extensive marine period in the Paradox Basin (Rigby et al. 1971).

While the Honaker Trail Formation is known to be fossiliferous, discerning what is known from it is obscured by the nomenclatural issues discussed above. In particular, it is not always possible to establish what unit is being discussed in references using "Hermosa Formation," such as those prior to 1958. For example, Williams (1949) reported corals, bryozoans (moss animals), brachiopods (lamp shells), bivalves, gastropods, trilobites, and crinoids (sea lilies) from the Hermosa Formation of southwestern Colorado; and Wengerd (1951) described reef limestones and bioherms in Hermosa Formation rocks from the Goosenecks of the San Juan River and upstream, with corals, bryozoans, brachiopods, gastropods, and crinoids. Of interest for CANY, McKnight (1940) mentioned the presence of foraminifera (amoeba-like protists that form "shells"), sponges, corals, bryozoans, brachiopods, bivalves, gastropods, trilobites, and crinoids in an area of Utah including the Island in the Sky District and farther north, and Rigby and Stokes (1971) described sponges from limestone ledges of the Hermosa Formation near Spanish Bottom. Geologic data for this publication is minimal, and the geographical information could place the find in the Honaker Trail Formation or the Elephant Canyon Formation. The six specimens pertain to the sponge *Haplistion sphaericum*. Two are in the collections of the National Museum of Natural History (USNM 166524 and 166525), two are in the collections of Brigham Young

University (BYU 1513 and 1514), one is at the University of Utah (number not given), and the other is in CANY collections (Rigby and Stokes 1971).

Definite Honaker Trail Formation fossils at CANY include algae (Rigby et al. 1971), fusulinid foraminifera (large foraminifera from the late Paleozoic, useful for correlating and dating beds), corals, bryozoans, brachiopods, bivalves, gastropods, trilobites, and crinoids (Baars 1993, 2010; Santucci and Kirkland 2010). Rigby et al. (1971) include several fossiliferous locations of their upper limestone member of the Hermosa Formation in their detailed description of the Colorado River canyon from Moab to Hite. Fossils are sometimes replaced by chert (Rigby et al. 1971). Palynomorphs (organic microfossils like spores and pollen) are known from cores just outside of the park, although at least some are probably reworked from the underlying Paradox Formation (Kosanke 1995). Plants have been recovered from a sandstone unit in the middle of the formation south of the Texas Gulf Sulfur Potash plant, about 10 km (6 miles) east of CANY. Fossils include coal, petrified logs, and compressed leaves from at least 20 taxa of lepidodendrales (clubmoss-like trees), horsetails, ferns, and conifers. The depositional environment is interpreted as a swampy delta (Tidwell 1988).

Other Honaker Trail Formation fossils include root traces from eolian (wind-blown) rocks (Loope 1994), other foraminifera, conodonts (extinct eel-like chordates), pellets (Ritter et al. 2002), borings (Loope 1994), and burrows (Loope and Kuntz 1987, as the upper member of the Hermosa Formation). Fossils of the upper Honaker Trail Formation include algae, fusulinid foraminifera, sponges, corals, bryozoans, brachiopods, bivalves, cephalopods, gastropods, trilobites and other arthropods, crinoids, echinoids, and trace fossils (Melton 1972; this could possibly pertain to the Elephant Canyon/Halgaito/lower Cutler interval instead).

Cutler Group: Elephant Canyon Formation/Halgaito Formation/lower Cutler beds (Upper Pennsylvanian–Lower Permian)

The relationships and usages of the Elephant Canyon Formation, Halgaito Formation (formerly Shale), and "Rico Formation" are complicated and differ by author, as discussed above. Condon (1997) recommended the informal term "lower Cutler beds" for all of the strata in the Elephant Canyon/Halgaito interval to side-step the controversy associated with the formal names. Per Huntoon et al. (1982), one of the maps used for the GRE program's digital map of the park, the Halgaito Formation and Elephant Canyon Formation are in use, with the Halgaito Formation entering a transition to the Elephant Canyon Formation beginning south of CANY and continuing through part of the southwestern corner, then being replaced by the Elephant Canyon Formation throughout the rest of the park. Anderson et al. (2010) proposed something similar, but substituted the lower Cutler beds for the Halgaito Formation in the area of northern GLCA that borders CANY. In simplified form, these strata can be considered a continuum going from north to south across CANY and GLCA: in the north are carbonate-dominated rocks attributed to the Elephant Canyon Formation, in the south are clastic-dominated rocks attributed to the Halgaito Formation, and between them is a transitional zone that has been called the lower Cutler beds (Anderson et al. 2010; G. Wills, pers. comm., February 2012). CANY may lack strata that would fit the definition of the Halgaito Formation; strata formerly identified as such (for example in Huntoon et al. [1982]) would instead be identified as lower Cutler beds (G. Willis, pers. comm., February 2012). The Rico Formation is considered obsolete, as noted above, and is included here for reference.

The Halgaito Formation is a red bed unit of sandstone, siltstone, claystone, and conglomerate, with thin marine beds (Huntoon et al. 1982). The Elephant Canyon Formation is similar to the Honaker Trail Formation, with a combination of gray fossiliferous limestone, limy sandstone, and red and gray limy siltstone and mudstone. The lower contact with the Honaker Trail Formation may be a disconformity (Billingsley et al. 2002; Baars 2010), or may be conformable (Loope et al. 1990; Condon 1997). The Elephant Canyon Formation intertongues with or grades into the overlying Cedar Mesa Sandstone (Baars 1991; Sumida et al. 1999b). Laterally, it also interfingers with red beds of the undivided Cutler Group (Baars 1983). Near the top of the Elephant Canyon Formation of CANY is a limestone layer known as the Shafer Limestone or Lime (McKnight 1940; Baars 2010). The Elephant Canyon/Halgaito/lower Cutler interval is 120 to 455 m (400 to 1,500 ft) thick at CANY (Huntoon et al. 1982).

The Elephant Canyon/Halgaito/lower Cutler interval was deposited during the Late Pennsylvanian and Early Permian (Sanderson and Verville 1990), sometime between 305 and 280 Ma. The setting for this interval is thought to have been a coastal plain to shallow marine complex in an arid area (Terrell 1972; Condon 1997), similar in some ways to the modern Colorado River delta and the Gulf of California. There was a combination of dunes, channels, mudflats, tidal flats, supratidal areas, deltas and prodeltas, bays, bars, and open marine areas that formed in repetitive cycles as a given area shifted from marine to nonmarine and back (Terrell 1972). Limestones and some of the shales and sandstones are marine, with common marine fossils; the fossils suggest relatively clear, warm shallow water. Depositional environments shifted rapidly with little transition (Campbell 1987).

The Elephant Canyon/Halgaito/lower Cutler interval is fossiliferous at CANY and adjacent areas. Because of the uncertainties of the geologic nomenclature, the following fossil reports are listed according to the unit from which they were described. Aside from fossiliferous limestone in general, fusulinids (Billingsley et al. 2002; Baars 2010), large bivalves from Meander Canyon (Baars 1993), palaeoniscid ray-finned fish vertebrae from Elephant Canyon near the Green River-

Colorado River confluence (UCLA collections) (Vaughn 1967; Sumida et al. 1999a), and petrified wood at mile 36.8 on Meander Canyon (Webb et al. 2004) have been reported from the Elephant Canyon Formation of CANY. Baars (1993) reported the presence of fusulinids, corals, bryozoans, brachiopods, bivalves, gastropods, trilobites, and crinoids in Elephant Canyon Formation rocks of the general Canyonlands area, and brachiopods, bivalves, gastropods, and rare trilobites in outcrops of the Shafer dome area just northeast of CANY. Condon (1997) included a photograph of a petrified stump within Upheaval Dome. Sumida et al. (1999a) reported xenacanth shark and palaeoniscid fossils from the Elephant Canyon Formation/Halgaito Formation transition of the Cane Creek Anticline, northeast of CANY. Several publications using the old Rico Formation terminology have reported fossils from CANY. Baker (1933) reported brachiopods, crinoids, and other fossils and shells from near The Slide. McKnight (1940) provided details for fourteen fossiliferous "Rico Formation" sites within CANY. Five are below the Shafer Limestone, and had bryozoans, brachiopods, bivalves, gastropods, and cephalopods. Four are from the Shafer Limestone, and had bryozoans, brachiopods, bivalves, gastropods, scaphopods, and trilobites. The remaining five sites are above the Shafer Limestone and had bryozoans, brachiopods, bivalves, gastropods, and ostracodes (seed shrimp). Rigby et al. (1971) reported bryozoans, brachiopods, bivalves, gastropods, and crinoids from outcrops of the "Rico Formation" as part of their guide to the Colorado River canyon between Moab and Hite, and noted several fossil-bearing locations within CANY.

From the literature, there appears to be three sites with Pennsylvanian or Permian vertebrate fossils just beyond the eastern border of CANY, but they are easily confused: two are equivalent stratigraphic units, but have different locations; two have equivalent locations, but are different stratigraphic units; and two have similar plant fossils, but differ in both stratigraphy and location. Lohman (1974) reported a red bed site in the middle of the "Rico Formation" in the eastern part of the Shafer dome, adjacent to the northeastern corner of CANY. This site is notable for the presence of a petrified wood "log jam" of 20 to 30 silicified logs up to 46 cm (18 in) in diameter, with an associated stump more than 90 cm (36 in) across, and shark teeth (including *Deltodus* and/or *Petalodus*). Sumida et al. (1999a) mention teeth from xenacanth sharks and palaeoniscids from a site in Halgaito Formation/Elephant Canyon Formation transitional rocks. The stratigraphy is reasonably equivalent to the Shafer dome site, but this site is on Indian Creek, and per its coordinates is across from the Loop, so is unlikely to be the same as Lohman's site. Finally, Sumida et al. (1999b) refer to another site on the Indian Creek drainage, but in the Cedar Mesa Sandstone or Cedar Mesa Sandstone/undivided Cutler Group transition (see below). This site also has a petrified wood "log jam," confusingly enough.

Fossils reported from the Elephant Canyon Formation in general include wood fragments, microbial mats, foraminifera, corals, bryozoans, brachiopods, bivalves, gastropods, cephalopods, trilobites, echinoderms, invertebrate burrows (Terrell 1972), and palaeoniscid fish (Sumida et al. 1999a). They are most abundant in the limestones (Campbell 1987). Fossils reported from the Halgaito Formation in general include seed ferns, lycopsids (clubmosses), the tree-like horsetail *Calamites*, foraminifera, bivalves, gastropods, crinoids, echinoids (sea urchins), sharks, palaeoniscid and phyllodont fish, lungfish, crossopterygians (lobe-fins), temnospondyls (extinct early amphibians), aïstopods (limbless amphibians), anthracosaurs (reptile-like tetrapods), early diapsid reptiles (the group including lizards, snakes, and crocodilians), and several types of pelycosaur synapsids ("mammal-like reptiles" of popular use), such as *Edaphosaurus* (Vaughn 1973; Sumida et al. 1999a, 1999b, 1999c). The flora indicates moist conditions. Bones are largely limited to stream channels (Vaughn 1973). Fossils reported from the "Rico Formation" include root traces, leaf impressions, logs (Loope 1984), foraminifera, corals, bryozoans, brachiopods, bivalves, scaphopods (tusk shells), gastropods, cephalopods, ostracodes, trilobites, crinoids (McKnight 1940), and fish (Lohman 1974), with corals and crinoids limited to the Shafer Limestone, and only bryozoans, brachiopods, bivalves, gastropods, and ostracodes above the Shafer Limestone (McKnight 1940).

Cutler Group: Cedar Mesa Sandstone (Lower Permian)

The Cedar Mesa Sandstone is a coastal sandstone unit with both marine and eolian influences, near a fluctuating shoreline (Baars 2010), although some older references interpreted it as more subaqueous (Kirkland 1963; Chamberlain and Baer 1973; Vaughn 1973). At CANY, it is up to 455 m (1,500 ft) thick, and is mostly composed of light red to white sandstone and siltstone, with interbedded dark red heterogeneous clastic (composed of rock fragment) rocks. It transitions into undivided Cutler Formation rocks across a broad area of CANY, which can be seen by the dark red interbedded units. Fossiliferous limestones are also included in this transition (Billingsley et al. 2002). Banded exposures of light-colored Cedar Mesa Sandstone beds and red Cutler Group sandstone beds are particularly prominent in The Needles District, representing overlapping stratigraphic intervals of coastal and fluvial origin (Baars 2010). Other contacts of the Cedar Mesa Sandstone, with the overlying Organ Rock Formation and underlying Elephant Canyon Formation, are also intertonguing and gradational at CANY (Billingsley et al. 2002). Outcrops of the Cedar Mesa Sandstone cover much of southern CANY (Huntoon et al. 1982).

The Cedar Mesa Sandstone was deposited during the Early Permian (Wolfcampian stage) (Blakey and Knepp 1989; Condon 1997), sometime between approximately 299 and 280 Ma. Multiple episodes of ergs are separated by erosional surfaces thought to be from marine transgressions (Baars 2010). Sand was blown by strong north-northwest winds (Parrish and Peterson 1988; Peterson 1988; Condon 1997).

Fossils are not common in the Cedar Mesa Sandstone, but a few reports of fossils have been made from CANY. There are foraminifera and (rare) crinoids in the sandstone (Baars 2010), and fossiliferous limestone is present in the transition with the undivided Cutler Group (Billingsley et al. 2002). Langford et al. (2007) found bryozoans, brachiopods, and crinoids from marine transgressive surfaces in sandstones of The Needles District. Loope (1984) plotted the location of two Cedar Mesa Sandstone track sites (both rockfall specimens from dune sandstones in the lower part of the unit) within CANY, and there may be track sites in the park which are accessible by river (Santucci and Kirkland 2010). Additional tracksites were reported in this formation by one of the authors, including a trackway with didactyl (two-toed) tracks reminiscent of the lower Permian *Dromopus* (VLS, March 2010), and possibly one of the longest vertebrate trackways known from the Paleozoic of North America (VLS, June 2010).

Just east of CANY, at Indian Creek, a few vertebrate fossils are known from the Cedar Mesa Sandstone or Cedar Mesa Sandstone/undivided Cutler Group transitional zone in one or two sites. Sumida et al. (1999b) described a site in the Cedar Mesa Sandstone that when first located included a petrified wood "log jam" with several large conifers, the largest 8 m (26 ft) long and 1 m (3 ft) in diameter, as well as bones of the temnospondyl amphibian *Eryops* and pelycosaur *Sphenacodon*. Unfortunately, this location was vandalized in 1995, with all fossils removed. The logs were recovered in 1996, but the vertebrate fossils are still missing (Sumida et al. 1999b). Stanesco et al. (2002) later published an abstract with a brief description of fossils from Indian Creek, but in the Cedar Mesa Sandstone/undivided Cutler Group transitional zone. These fossils include root casts and conifer logs in an arroyo complex cut into eolian rocks, and disseminated plants, vertebrate bone fragments, and teeth from an overlying shallow lake limestone (Stanesco et al. 2002). While the stratigraphic difference does not preclude these being the same site given transitions are a matter of interpretation, the vandalism of the first site would suggest that this second publication at least refers to a different area of the site.

Three published track sites have been found in the Cedar Mesa Sandstone at adjoining GLCA, with two more sites just outside of its boundaries. A site at Dirty Devil, now submerged, appears to preserve a predation event, with a larger animal catching a smaller animal. The other track sites are not as potentially dramatic. Synapsids (the group including pelycosaurs, other mammal relatives, and true mammals) were the primary track makers. Lizard-like tracks are also known from one site (Lockley et al. 1998). Several articulated tetrapod vertebrae have been found from the middle of the Cedar Mesa Sandstone in the Canyonlands area (Loope 1984). Loope (1984) also reported root traces, cylindrical burrows, and occasional sand-sized fragments of marine invertebrate fossils from area outcrops. Fossils found in the Cedar Mesa Sandstone farther afield from CANY include ghost shrimp burrows and other invertebrate trace fossils from near the top of the unit (Chamberlain and Baer 1973), rare casts of marine invertebrates (Kirkland 1963), and vertebrate tracks (Sumida et al. 1999b).

Cutler Group: Organ Rock Formation (Lower Permian)

The Organ Rock Formation (or Shale or Tongue) is a red bed sandstone and shale unit. At CANY, it is 75 to 120 m (250 to 400 ft) thick, and is composed of dark red sandstone and siltstone, with a few conglomeratic ledges in the upper section (Huntoon et al. 1982; Billingsley et al. 2002). It is generally a slope-former at CANY, and grades from more coarse rocks in eastern CANY as it transitions out of undivided Cutler Group sandstone, to finer-grained rocks in the western part of the park (Baars 2010). It intertongues with the underlying Cedar Mesa Sandstone at CANY, where the contact is gradational and arbitrary (Billingsley et al. 2002); the contact with the overlying White Rim Sandstone is sharp, but with some interfingering (Tubbs 1989). It is thought to be equivalent to the Hermit Formation (or Shale) of the Grand Canyon area (Blakey and Knepp 1989). Deposition of the Organ Rock sediments occurred during the Early Permian (Leonardian stage), sometime between approximately 280 and 271 Ma (Blakey and Knepp 1989; Stanesco et al. 2000). Plant fossils indicate that the climate was semi-arid, with long growing seasons (Vaughn 1964). Its depositional environments ranged from marginal marine to continental lowlands (Dubiel et al. 1989). Eolian rocks are common at CANY, with interbedded channel rocks. Eolian rocks are particularly common west of the Colorado River; dune movement was to the southeast. Fluvial and floodplain rocks dominate to the southeast (Stanesco et al. 2000).

Organ Rock Formation fossils from CANY are limited. There are root traces on several horizons, representing periods of stability (Stanesco et al. 2000). Burrows are also present (Chan 1989). Elsewhere, vertebrates and plants are known. The small vertebrate assemblage is best known from the Monument Valley area. Fossils are largely restricted to stream channel deposits found about 11 m (35 ft) above the base of the unit. Fossiliferous deposits are typically in conglomerate or associated sandstone, or sandstone lenses. These fossiliferous locations stand out from surrounding red bed mudstones, shales, and sandstones (Vaughn 1964). The vertebrates include sharks, palaeoniscid fish, at least three types of labyrinthodont amphibians including *Eryops*, the reptiliomorphs (early reptile-like tetrapods) *Seymouria sanjuanensis*, *Diadectes*, and *Tseajaia campi*, and the pelycosaurs *Ophiacodon*, *Ctenospondylus, Dimetrodon*, and *Sphenacodon* (Vaughn 1973; Sumida et al. 1999a, 1999b, 1999c). Fish are rare; scales are only known from coprolites (fossil feces). Additionally, tetrapod tracks are also known, from a greater stratigraphic range than body fossils (Vaughn 1964). Plant fossils include specimens of *Callipteris*, *Protoblechnum*, *Supaia*, *Taeniopteris*, *Walchia*, and *Yakia*, representing pteridosperms (seed ferns) and possible ferns and conifers (Vaughn 1964; Mamay and Breed 1970). These are

similar to the floral assemblage of the Hermit Formation (Mamay and Breed 1970).

Cutler Group: White Rim Sandstone (Lower Permian)

The White Rim Sandstone is a shallow water marine and nearshore eolian unit, deposited during a major marine transgression (Dubiel et al. 1989). It grades into marine rocks of the Toroweap Formation and overlying Kaibab Formation to the west (Steele 1987). At CANY, it is 0 to 75 m (0 to 250 ft) thick and composed of gray sandstone (Huntoon et al. 1982), eventually pinching out against the Organ Rock Formation (Huntoon and Kamola 1988). Because it lacks time-diagnostic fossils or other material that could be used to determine its age, the White Rim Sandstone can only be dated indirectly. It must be of Early Permian age, at least in part, because it is known to interfinger with the underlying Organ Rock Formation (Anderson et al. 2010). The contact with the overlying Moenkopi Formation is a disconformity (Huntoon et al. 1982). Interpretation of White Rim Sandstone paleoenvironments has proven contentious. It was usually assumed to be eolian, but the first detailed examination proposed it was a shallow marine sandstone (Baars and Seager 1970). Steele-Mallory (1982) reinterpreted the unit as largely coastal eolian deposits and interdunes (the spaces between dunes). A third interpretation combines the two previous attempts, considering the lower part eolian and the upper part marine, with more marine influence farther west. Sea level rose discontinuously with an accelerating rate, and marine waters only reworked a thin part of the upper White Rim Sandstone in CANY (the "marine veneer"). The White Rim Sandstone of the CANY area includes dune, interdune, sabkha (salt flat), and sand sheet deposits, as part of an erg centered on the Dirty Devil area to the west between a more distant western sea and alluvial fans from the Uncompahgre Uplift to the east. The climate was arid to semi-arid (Hansley 1995). Evidence of marine influence is confined to the top of the formation at CANY (Kamola and Huntoon 1994). Interestingly, there are thin red beds at the top of the White Rim Sandstone in the Elaterite Basin and Stillwater Canyon, which are not part of the overlying Moenkopi Formation (Baars 2010). They may be a tongue of the undivided Cutler Group (Baker 1946), fluvial rocks of the Kaibab Formation, or a new and previously unrecognized unit, and have some bearing on issues related to the Cutler Group–Moenkopi Formation contact in southern Utah (Baars 2010).

Fossils are rare in this unit, but there are several reports from CANY and the surrounding area. Steele-Mallory (1982, 1987) reported a crinoid fragment from a site at White Crack in CANY, between the White Rim Trail and the Green River–Colorado River confluence, and noted the presence of fecal pellets and bioturbation (erasing of sedimentary structures by extensive organism activity) in interdune deposits. Park outcrops include root traces and bioturbation in sheet sand, tree trunk traces and root traces in dunes, and possible burrows and root traces in the upper reworked rocks; microbial laminations are present in a sabkha deposit north of Anderson Bottom in the park. Dune traces can be found between Queen Anne's Bottom and Anderson Bottom (Chan 1989). Baars and Seager (1970) reported possible impressions of something similar to a kelp-like brown algae from the Elaterite Basin, which they interpreted as evidence for shelf marine deposition. Burrows are also known from the lower part of the unit in the basin (Huntoon and Chan 1987) and on the Shafer Trail (Huntoon and Dubiel 2000). Elsewhere, the upper marine reworked portion has yielded echinoids (Baars 1987), abraded shells, and marine trace fossils (Orgill 1971; Steele 1987; Huntoon and Kamola 1988; Kamola and Chan 1988).

Undivided Cutler Group and transitional zones (Upper Pennsylvanian–Lower Permian)

The Cutler Group at CANY includes transitional zones where various named formations are replaced by undivided rocks of the Cutler Group (undivided Cutler Group rocks are often designated as Cutler Formation). Broadly, this occurs going from southwest to northeast and from low strata to high strata in the outcrops at CANY (skewed by the absence of rocks younger than the Cedar Mesa Sandstone in most of the southern half of the park). Per Huntoon et al. (1982) and Doelling (2004), the southern half of CANY is dominated by the Cedar Mesa Sandstone, which then changes to a Cedar Mesa Sandstone/undivided Cutler Group zone in the park's midsection, and finally northeastern outcrops of this interval are solely undivided Cutler Group rocks. Above it, the Organ Rock Formation is present in central and northwestern CANY, changing to undivided Cutler Group rocks via a transitional zone in northeastern CANY. Finally, the White Rim Sandstone is present throughout northern CANY, but disappears to the northeast. Undivided Cutler Group rocks at CANY are red fluvial sandstones and white marine sandstones with some red shales. The unit is at most between 550 and 670 m (1,800 and 2,200 ft) thick, and forms a series of slopes and ledges in outcrop. The transitional zones are marked by intertonguing of the formations involved (Huntoon et al. 1982; Doelling 2004). The depositional setting is interpreted as large alluvial fans and floodplains on the southwest flank of the uplift (Rigby et al. 1971).

Stokes (1986) published a photograph of possible amphibian tracks from unspecified Cutler Formation rocks on the lower Labyrinth Canyon, in northwestern CANY. A few fossils have been found in undivided Cutler Group rocks near CANY: material has been found in the Cedar Mesa Sandstone transitional zone at Indian Creek just east of the park (see the Cedar Mesa Sandstone section above); wood is also known in the general CANY area (Rigby et al. 1971); and fossil logs and leaves have been found just east of CANY in undivided rocks (Baker 1933). Otherwise, the nearest significant finds from undivided Cutler Group rocks come from the Placerville area of western Colorado, about 145 km (90 miles) to the east of CANY (Baird 1965; Lewis and Vaughn 1965; Laurin 1994). Finds here include labyrinthodont amphibians, early amniotes and amniote

relatives, pelycosaurs (Lewis and Vaughn 1965), and footprints of labyrinthodonts and amniote relatives (Baird 1965). There is bioturbation in an unnamed reddish sandstone locally present above the White Rim Sandstone in CANY, but this unit could belong to the Cutler Group or the Moenkopi Formation (Steele 1987).

Moenkopi Formation (Lower–Middle Triassic)

The Moenkopi Formation is a heterogeneous unit present in several western states. At CANY, it averages 185 m (600 ft) thick, mostly composed of dark brown mudstone and siltstone (Baars 2010). The lower contact with the Cutler Group and the upper contact with the Chinle Formation are disconformities (Huntoon et al. 1982). The Moenkopi Formation dates to the Early and early Middle Triassic (Heckert et al. 2005a), between approximately 250 and 240 Ma. It has been divided into several members, depending on location. In southeastern Utah, names were assigned to the members only relatively recently, aside from the Sinbad Limestone (Blakey 1974). At CANY and the surrounding area, it may be described as an undivided formation, as a basal member (Hoskinnini Member) and an undivided remainder (Baars 2010), or as multiple members (Huntoon et al. 1982). Huntoon et al. (1982) used, from oldest to youngest, the Hoskinnini Member, lower slope-forming member, Sinbad Limestone, ledge-forming member, and upper slope-forming member, following the stratigraphy of Stewart et al. (1972a). These unnamed members were later reevaluated by Blakey (1974), who named the lower slope-forming member the Black Dragon Member, the ledge-forming member the Torrey Member, and the upper slope-forming member as part of the Moody Canyon Member. At least one report (Lucas 1995) uses completely different nomenclature for the Moenkopi Formation of CANY, based on names from western Colorado and adjoining parts of Utah. The Moenkopi Formation of CANY and adjoining areas is essentially the remains of a large coastal plain mudflat, influenced by and sometimes covered by the neighboring sea. The Black Dragon Member and Sinbad Limestone are interpreted as coastal and marine units, the Torrey Member as a large delta, and the Moody Canyon Member is interpreted as a restricted shallow water to sabkha setting (Blakey 1974).

The Hoskinnini Member is a cliff-forming red bed unit (Finnell et al. 1963), composed of silty sandstone and sandy siltstone. It was transferred from the underlying Cutler Group (Stewart 1959), and is occasionally classified as a formation on its own, with the age mooted as possibly at least partially Permian (Blakey 1974; Baars 2010). This unit was probably deposited under subaqueous but nonmarine conditions in an enclosed or restricted setting (Blakey 1973). The Hoskinnini Member lacks diagnostic fossils (Finnell et al. 1963) but does have stromatolites (layered sedimentary structures formed by microorganisms) (J. Kirkland, pers. comm., July 2009). Heavy bioturbation is present in the Hoskinnini Member on the Shafer Trail (Huntoon and Dubiel 2000).

The Black Dragon Member is a heterogeneous unit of mudstone to sandstone, limestone, dolomite, and gypsum (Blakey 1974). It is slightly younger than the Hoskinnini Member but occupies a similar stratigraphic position, and grades into the Hoskinnini Member in the White Canyon area. The zone of interfingering begins west of CANY and continues south through Indian Head Pass and Happy Jack Mine (Huntoon et al. 1994). The upper contact with the Sinbad Limestone is transitional (Stewart et al. 1972a) to intertonguing (Blakey 1974). Because of its similarity to the Torrey Member, where the Sinbad Limestone has pinched out the Torrey and Black Dragon members are difficult to differentiate. The Sinbad Limestone pinches out just west of the Colorado River generally, and the Black Dragon Member can only be distinguished a little farther east. It can be over 60 m (200 ft) thick in the park, though (Blakey 1974). It was deposited in coastal plain and tidal flat settings (Morris et al. 2010). Fossils are not currently well-known in the Black Dragon Member. Some body fossils are known, such as bivalves, gastropods, a possible ammonite, ostracodes, and a conodont fragment. These body fossils are best known from the San Rafael Swell (Blakey 1974). Trace fossils are known as well (Blakey 1974).

The Sinbad Limestone Member was recognized before the other members, because it is distinctive compared to the surrounding red beds of sandstone, siltstone, mudstone, and claystone (McKee 1954). It is mostly composed of limestone and dolomite, with some limy siltstone. Generally, it is very thin at CANY, less than 3 m (10 ft), and typically pinches out west of the Colorado River (Blakey 1974). It represents the easternmost marine carbonates of the Early Triassic of the Colorado Plateau. Depositional environments included offshore bars, shallow shelves, and lagoons (Blakey 1974). Because it is a marine unit, Lucas et al. (2007a) proposed removing it from the Moenkopi Formation as part of a reorganization to raise the formation to a group and limit it to terrestrial rocks. It grades into the overlying Torrey Member (Stewart et al. 1972a). The Sinbad Limestone at CANY, on the Green River near the confluence, has yielded a few poorly preserved casts and molds of brachiopods (*Lingula*), bivalves (*Monotis*), gastropods, and ammonites (*Meekoceras*) (Baker 1933; McKnight 1940; Santucci and Kirkland 2010). The database of the New Mexico Museum of Natural History and Science's collections includes gastropods collected from Sinbad Limestone outcrops of the Upheaval Dome area of CANY (P-27083) (NMMNH online database no date). Fossils of the Sinbad Limestone in general include stromatolites, algae, foraminifera, sponges, brachiopods, bivalves, ammonite cephalopods, gastropods, scaphopods, ostracodes, crinoids, echinoids, conodonts, possible fish material, and marine trace fossils (McKee 1954; Stewart et al. 1972a; Blakey 1974; Dean 1981; Goodspeed and Lucas 2007). Newly-recognized in the San Rafael Swell area are brittle stars and serpulid worm tubes (Hautmann and Nützel 2005).

The Torrey Member is mostly composed of sandstone and siltstone, but many other rock types are present in minor

quantities. Often the base is siltstone, followed by ledgey sandstone in the middle, and then an upper sandstone/siltstone section. It is 45 to 75 m (150 to 250 ft) thick at CANY (Blakey 1974). Its rocks generally represent deltaic and fluvial settings (Goodspeed and Lucas 2007). It grades into the overlying Moody Canyon Member (Stewart et al. 1972a). Fossils from the Torrey Member include horsetails (*Equisetum*), brachiopods, bivalves, ostracodes, fish scales (Blakey 1974), fish bones (Mickelson et al. 2006), a skull of the labyrinthodont amphibian *Parotosuchus* from Hite Crossing (Lucas and Schoch 2002), invertebrate traces, fish swim traces, and tetrapod tracks (Mickelson et al. 2006). Extensive finds, particularly tracks and traces from invertebrates and vertebrates, have now been reported from the Torrey Member and equivalents at GLCA, CARE, and the San Rafael Swell (Schultz et al. 1995; Mickelson et al. 2006). Tracks have been found as sandstone casts in mudstone. To date, reptile tracks are dominant, with horsetail molds, invertebrate trace fossils (such as those of worms and horseshoe crabs), fish fin marks, and fish bones also found (Mickelson et al. 2006). Blakey (1974) noted the presence of fish and amphibian bones in this member at Bears Ears and Indian Canyon "in southeastern Canyonlands National Park," but this would seem to be in error, as Indian Creek and its canyon are almost entirely outside of CANY on its east-central border, and Bears Ears is in Manti–La Sal National Forest to the southeast; perhaps general Canyonlands was intended instead of specific CANY.

The Moody Canyon Member is a heterogeneous unit with red to brown siltstones and mudstones, sandstones, and an upper section of dolomite and gypsum. It is notable for its continuous regular beds, and is less than 30 m (100 ft) thick at CANY (Blakey 1974). Fossils are rare in this unit, with only *Chirotherium* tracks from an outcrop in the Orange Cliffs (Blakey 1974) reported to date.

There are a few records of Moenkopi Formation fossils at or near CANY where the member is not specified. Trace fossils such as burrows are known to be present at CANY and the vicinity (Baars 2010), such as burrows and trails on the low broad tidal flats of the Moenkopi Formation between Moab and Hite (Rigby et al. 1971). Carbonaceous debris was reported in the Moenkopi Formation at Taylor Canyon (Stewart et al. 1972b). Track sites have been found in Moenkopi Formation rocks at GLCA (Mickelson et al. 2006). Two have been published: one at Farley Canyon, and one at Trachyte Point. The Farley Canyon site has three types of tracks, including horseshoe crab tracks, swim traces, and lizard-like tracks; the horseshoe crab tracks are most abundant. The Trachyte Point tracks include horseshoe crab tracks and swim marks (Lockley et al. 1998). Fish scales and a possible coelacanth fragment were reported from the upper part of the formation in the Deer Flat area of White Canyon (Finnell et al. 1963).

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location (Stewart et al. 1972b). In CANY, these include from oldest to youngest the Monitor Butte, Moss Back, Petrified Forest, Owl Rock, and Church Rock members (Huntoon et al. 1982), but these are often lumped together, especially above the Moss Back Member (Lohman 1974). They can be divided into bentonitic (clay formed from volcanic ash) and red bed parts of the Chinle Formation, with the Petrified Forest Member at the top of the bentonitic portion and the Owl Rock Member as the base of the red bed portion (Stewart et al. 1972b). In the Canyonlands area, the members are difficult to distinguish (Baars 1993), but they will be briefly described for reference. At CANY, the Chinle Formation is 100 to 300 m (330 to 660 ft) thick, and is composed of red, purple, green, and yellow clay-rich sandstones and siltstones forming ledges and slopes (Huntoon et al. 1982). The lower contact with the Moenkopi Formation is a disconformity (Huntoon et al. 1982), and the upper contact with the Wingate Sandstone is gradational (Tanner 2003). The Chinle Formation is particularly known for petrified wood and other plant fossils, and such fossils have been reported from CANY and the vicinity (Baker 1946; Williams 1964; Rigby et al. 1971; Dubiel 1982, 1983; Patterson et al. 1989; Baars 1993, 2010). The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) because this division is known to be in the overlying Wingate Sandstone (Lucas et al. 1997a, 2006a).

The Monitor Butte Member is interpreted as valley fill deposited under a monsoonal climate (Demko et al. 2005). It is a mudstone, siltstone, and sandstone unit (Parrish 1989), with deposition in a system of braided and meandering streams, lakes, and swamps, with volcanic input (Blakey and Gubitosa 1983). It is not present extensively at CANY, and is mainly known from the southwestern part of the park (Stewart et al. 1972b).

The Moss Back (or Mossback) Member was deposited in fluvial, crevasse splay, and floodplain settings (Dubiel and Brown 1993). At CANY, it records two alternating fluvial-lacustrine sequences with high and low fluvial outputs, and inputs of volcanic debris from the southwest (Gubitosa 1981). The Moss Back Member may be the equivalent of the Sonsela Sandstone unit of the Chinle Formation found in eastern Arizona (Blakey and Ranney 2008).

The Petrified Forest Member was deposited by streams in a similar way to the Monitor Butte Member, as a sandstone/mudstone fluvial unit with volcanic input (Blakey and Gubitosa 1983, 1984). It is divisible into smaller units at locations like Petrified Forest National Park (PEFO), but this is not generally done at CANY (but see Heckert et al. 1999).

Chinle Formation fossil vertebrates are most diverse in the Petrified Forest Member's various units (Parrish 1989).

The Owl Rock Member is a carbonate and mudstone unit (Heckert and Lucas 2002). It was traditionally interpreted as largely lacustrine because of its carbonate beds (Blakey and Gubitosa 1983; Dubiel 1989; Parrish 1989). This was revised in the 1990s, with the carbonate units instead thought to represent ancient soil horizons from a semi-arid setting (Lucas et al. 1997a; Tanner 2000). Further study concluded that the Owl Rock Member was a combination of both soil horizons and small ponds and lakes, on a fluvial floodplain (Lucas and Heckert 2003; Fraser et al. 2005). The Owl Rock Member represents the beginning of aridification in the Chinle Formation as Pangaea drifted north (Heckert and Lucas 2002; Tanner 2003). It grades into the overlying Church Rock Member in the CANY area (Stewart et al. 1972b).

There is a distinctive sandstone-siltstone bed at CANY informally called the "Black Ledge," found between the Owl Rock and Church Rock members. It is sometimes put in the Owl Rock Member (KellerLynn 2005), the Church Rock Member (Stewart et al. 1972b), or not committed to either (Huntoon and Dubiel 2000; Baars 2010).

The Church Rock Member is a siltstone-sandstone unit with stream (Blakey and Gubitosa 1983), lake margin, and eolian deposition, with eolian more prominent toward the top as the unit transitions in the Wingate Sandstone (Dubiel 1989). Deposition during the Church Rock Member shows the continuation of aridification begun in the underlying Owl Rock Member (Tanner 2003), as the dunes of the Wingate Sandstone approached from the northwest (Blakey and Gubitosa 1983). At the top of the Church Rock Member in the CANY area is a sandstone and conglomerate bed called the Hite Bed (Stewart et al. 1972b; Dubiel et al. 1989).

Fossils are known from several locations in the Chinle Formation at CANY. Santucci and Kirkland (2010) reported that the formation's assemblage within CANY includes *Neocalamites*, *Pagiophyllum*, other plants, gastropods, bivalves, conchostracans, crayfish (CANY Figure 5), invertebrate burrows (from crayfish, worms, and insects), a horseshoe crab resting trace, semionotid and turseodid fish, metoposaurs, phytosaurs, aetosaurs, and tetrapod tracks. A partial skeleton of the herbivorous armored aetosaur *Stagonolepis* has been recovered from fluvial lower Petrified Forest Member rocks in the park by personnel of the New Mexico Museum of Natural History and Science (NMMNH). The specimen, NMMNH P-26938, includes scutes, vertebrae, ribs, and other bones, mostly still in matrix, and is currently at the NMMNH. The locality is NMMNH Locality 3279 (Heckert et al. 1999). A track identified as *Grallator* (representing a small theropod dinosaur) was found at Upheaval Dome in the Church Rock Member (Lucas et al. 1995; Santucci et al. 1998); it was left *in situ* (Lucas et al. 1995). Tracks of *Brachytherium* (aetosaur-like?) and *Pentasauropus* (dicynodont synapsid [a type of mammal relative]?) are known from unspecified Chinle Formation rocks at the park as well (Santucci et al. 1998). Unionid bivalves have been collected from the Chinle Formation at Upheaval Dome (McKnight 1940). Possible worm borings are present in the Church Rock Member at Taylor Canyon, and carbonaceous debris is found in the Petrified Forest and Owl Rock members at the same location (Stewart et al. 1972b). Bioturbated siltstone is also present in the Owl Rock Member at the park (Lucas et al. 1995). Finally, fossil wood and otherwise unidentified reptile bones are known from the Chinle Formation above the Moss Back Member in general at CANY (Lohman 1974).



CANY Figure 5. Fossil crayfish from the Owl Rock Member of the Chinle Formation of CANY, compared to a modern crayfish (photo courtesy Jim Kirkland, Utah Geological Survey).

Diverse fossils are known from the Chinle Formation of the CANY area, including microbial mats and other structures (Blakey and Gubitosa 1982), leaves, stems, needles, cones, wood, bivalves, gastropods, crayfish, conchostracans (clam shrimp), traces from worms, crayfish, horseshoe crabs, and insects, and bones of fish, metoposaur amphibians, aetosaurs, and crocodile-like phytosaurs (Hasiotis 1992, 1993). Petrified

wood is common and usually black, while amphibian and reptile bone fragments are rare (Baars 1993). Teeth have been found on the north side of Lockhart Canyon, possibly within CANY (Baker 1933). The Monitor Butte Member has yielded plant fragments, petrified wood, the plant genera *Phlebopteris* and *Zamites* (Dubiel 1982, 1983), bivalves, conchostracans, ostracodes, fish bones and scales (Dubiel 1987), bones of amphibians, phytosaurs, crocodile relatives, and other tetrapods, coprolites (Parrish and Good 1987), burrows, and bioturbation (Dubiel 1987), with productive localities on the Red and White canyons (Dubiel 1982, 1983, 1987). The Moss Back Member has wood fragments (Dubiel 1982), bivalves, a partial limb bone of a crocodile relative (Parrish and Good 1987), and trace fossils in the same area (Dubiel 1987), and the Petrified Forest Member has produced bivalves, gastropods, lungfish toothplates, bones of metoposaurs, aetosaurs, phytosaurs, other archosaurs, and a cynodont (another type of mammal relative) (Parrish and Good 1987), other vertebrate bones, coprolites, and burrows on the two canyons as well (Dubiel 1987). Phytosaur fossils ranging from fragments to skulls have been found in numerous locations in southeastern Utah, from the Monitor Butte Member to the Petrified Forest Member (Parrish and Good 1987). Crayfish and crayfish burrows are known from the Petrified Forest, Owl Rock, and Church Rock members, in locations including Hart's Draw, Hart's Point, and Stevens Canyon (Hasiotis and Mitchell 1989), with thousands present in a multi-layer mass mortality bed in the Owl Rock Member (Hasiotis 1995). More possible crayfish burrows are present in the Owl Rock Member on the Shafer Trail; a loose rock with conifer frond impressions has also been found on the Trail, from either the Black Ledge or Owl Rock Member (Huntoon and Dubiel 2000). Ostracodes are present but rare in the Owl Rock Member of the Red and White canyons area (Dubiel 1987). Fossils from the Chinle Formation at GLCA include petrified wood and carbonaceous debris, gastropods, crayfish burrows, bones, coprolites, and dinosaur tracks (McDonald and Santucci 2000; Anderson et al. 2010; Santucci and Kirkland 2010). Track sites have been found in Chinle Formation rocks at the park, and two have been described (Lockley et al. 1998). Widely distributed petrified logs are present in the Moss Back Member of the GLCA area (Finnell et al. 1963). Finds from the uppermost members of the Chinle Formation in San Juan County include ferns and other plants, bivalves, gastropods, ostracodes, conchostracans, fish, tetrapod bone fragments, and invertebrate and reptile tracks (Milner 2006; Spears et al. 2006). Among these is a significant fish site in Lisbon Valley (Spears et al. 2006).

The various members of the Chinle Formation have similar types of fossils, but differ in the genera and species present and which are most commonly found. Plants are represented by charophyte algae (Lucas and Keitzke 1993), diverse palynomorphs (Gottesfeld 1972; Fisher and Dunay 1984; Parker 2004), and leaves, wood, and other remains of lycopods, horsetails, ferns and fern-like plants, cycads, bennettitales, ginkgoes, conifers, and taxa of uncertain classifications (Ash 1972, 1989, 2003, 2005a; Stewart et al. 1972b; Parker and Ash 2004). Fungal remains are also present (Ash 1972; Stewart et al. 1972b). Invertebrates are represented by tentaculites (enigmatic possible mollusks, misidentified as worm fossils [Ash 2005b]), bivalves, gastropods, ostracodes, conchostracans (Breed 1972; Stewart et al. 1972b), crabs (Rinehart et al. 2003), crayfish (Miller and Ash 1988), insects (Breed 1972; Stewart et al. 1972b), and various trace fossils (Caster 1944; Breed 1972; Gillette et al. 2003; Demko et al. 2005), including borings in wood (Walker 1938; Tapanila and Roberts 2012) and a variety of feeding traces on plants (Ash 1996). Reworked Paleozoic fossils, primarily of invertebrates, are occasionally found in the lowest part of the formation, and include foraminifera, algae, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, ostracodes, and crinoids (McKee 1937; Stewart et al. 1972b).

The Chinle Formation's vertebrate assemblage has been summarized in a number of documents, including Stewart et al. (1972b), Colbert (1972), Parrish (1989), Long and Murry (1995), Heckert et al. (2005a, 2005b), Irmis (2005a), Parker (2005, 2009), Lucas et al. (2007b), Loughney et al. (2011), and Parker and Martz (2011). Vertebrates include: freshwater sharks; ray-finned fish; lungfish; coelacanths; several types of enigmatic tetrapods known from isolated bones and teeth; metoposaur amphibians; procolophonids (stocky lizard-like herbivorous and insectivorous reptiles); sphenodontians (relatives of the modern tuatara); trilophosaurids (reptiles that resembled large herbivorous lizards); tanystropheids (long-necked piscivorous reptiles); the unusual aquatic reptile *Vancleavea*; drepanosaurids (beaked, possibly aquatic chameleon-like reptiles); several forms of crocodile-like phytosaurs, known under a variety of names; several forms of herbivorous armored aetosaurs, also known under a variety of names; the small unusual armored *Revueltosaurus*; "rauisuchian" crocodile relatives including poposaurids and rauisuchids, which resembled large carnivorous dinosaurs, and shuvosaurids (toothless forms that resembled ostrich-mimic dinosaurs); "sphenosuchians" (small terrestrial crocodile relatives); lagerpetonids (small bipedal dinosaur relatives); silesaurids (quadrupedal dinosaur relatives); several early dinosaur taxa; and dicynodonts (a type of non-mammalian therapsid, or "mammal-like reptile"). Lungfish, metoposaurs, aetosaurs, and phytosaurs are the most common forms. Vertebrate trace fossils include coprolites (fossil dung) (Hunt et al. 1998), swim traces, and footprints possibly attributable to sphenodontians, tanystropheids, phytosaurs, aetosaurs, carnivorous and herbivorous dinosaurs, and mammal relatives (Hunt and Lucas 2007). Bones accumulated in floodplains, bogs, small ponds, and in or near channels (Parrish 1989).

Glen Canyon Group: Kayenta Formation (Lower Jurassic)

The Kayenta Formation is mostly made up of fine- to coarse-grained sandstone with some small amounts of interbedded shale and siltstone, and rare limestone and conglomerate (Vandiver 1936). At CANY, it is 60 to 80 m (200 to 260 ft) thick, and is mostly composed of reddish brown sandstone.

Siltstone, shale, and limestone interbeds are also present. The lower part of the unit forms cliffs, and the upper part forms slopes (Huntoon et al. 1982). The resistant lower beds cap the cliffs of the Wingate Sandstone. Beds are interpreted as fluvial (Baars 2010). The lower contact of the Kayenta Formation is gradational with the Wingate Sandstone at nearby GLCA (Anderson et al. 2010). The upper Kayenta Formation intertongues with the lower Navajo Sandstone, representing a time when the Navajo Sandstone dunes advanced over the rivers of the Kayenta Formation (Harshbarger et al. 1957). The age of the Kayenta Formation is now thought to be middle Early Jurassic, between approximately 197 and 190 Ma (Lucas et al. 2005). The Kayenta Formation was formed by shifting freshwater braided and meandering streams (Colbert 1974); floodplain deposits are also known (Lucas et al. 2005).

The Kayenta Formation is known for its dinosaur tracks, and a few are known at CANY (CANY Figure 5) (Santucci et al. 1998; Santucci and Kirkland 2010). Tracks are known from the upper part of the formation, and from the Kayenta Formation–Navajo Sandstone transition in the Horseshoe Canyon unit (Santucci et al. 1998). Dinosaur tracks and freshwater bivalves have been found from the transitional zone to the northwest of CANY in the San Rafael Swell, and bivalves are known from the Moab area (Baker 1946). Additionally, reed impressions are known from the CANY area (Rigby et al. 1971).



CANY Figure 5. An example of a tridactyl dinosaur track (VLS).

Tracks are well-known to the south in GLCA. Six sites have been described from this park and its immediate vicinity, from Explorer's Canyon, Long Canyon, Mike's Mesa, Slick Rock Canyon (two sites), and at neighboring Rainbow Bridge National Monument (RABR). Most tracks were made by small (*Grallator*) or large (*Eubrontes*) theropods (Lockley et al. 1998). The recent low water levels in Lake Powell have revealed more sites, especially in the Kayenta Formation–Navajo Sandstone transition zone. At least 29 track sites have been found in the Kayenta–Navajo transition of GLCA (Santucci and Kirkland 2010).

Vertebrate body fossils of the Kayenta Formation have been described in many publications, but most are from a silty facies found to the south in Arizona (see Lucas et al. [2005] or the Navajo National Monument summary of the Southern Colorado Plateau Network report for a partial list). Vertebrate body fossils of the sandy facies of the Kayenta Formation, which would be typical of CANY, include fish scales, crocodylomorphs (crocodiles and their relatives), and tritylodontid synapsids (early mammal relatives) such as *Kayentatherium* (Clark and Fastovsky 1986). Kayenta Formation vertebrate trace fossils include coprolites (Luttrell and Morales 1993), therapsid tracks, lizard-like tracks (Hamblin and Foster 2000), and dinosaur tracks such as *Grallator* and *Eubrontes* (Lucas and Tanner 2007). The relevant ichnofossil taxonomy can be complicated. A recent summary for the Chinle Formation and Glen Canyon Group can be found in Lucas et al. (2006b). Kayenta Formation fossils that are not vertebrates or vertebrate traces include microbial limestone (Luttrell and Morales 1993), petrified wood (Jenkins and Shubin 1998), plant impressions (Rigby et al. 1971), freshwater bivalves and gastropods (Harshbarger et al. 1957), ostracodes (Lucas and Tanner 2007), and invertebrate trails and burrows (Luttrell and Morales 1993).

Glen Canyon Group: Navajo Sandstone (Lower Jurassic)

The Navajo Sandstone is a fine- to medium-grained quartz sandstone noted for its prominent cross-bedding formed by dune progression. There are also lenticular beds of cherty limestone (Harshbarger et al. 1957). At CANY, it is 75 to 135 m (250 to 450 ft) thick and is composed of buff to pale orange sandstone, sandy limestone, and chert stringers. It intertongues with the underlying Kayenta Formation (Huntoon et al. 1982). While dating of Glen Canyon Group units has proven difficult to establish, the bulk of the Navajo Sandstone probably dates to the middle and late Early Jurassic (Lucas et al. 2005), between approximately 190 and 175 Ma.

The Navajo Sandstone represents the largest erg known from North America (Blakey 2008). As discussed in the "Kayenta Formation" section, the upper Kayenta Formation intertongues with the lower Navajo Sandstone as the dunes of the latter overrode the streams of the former. Recent research on geochronology of detrital zircon grains in the Navajo Sandstone suggests that most of the sand was eroded from the ancestral Appalachian Mountains, transported westward by a continental-scale river system to the western shore of Jurassic North America, and was then blown southward and incorporated into the Navajo–Nugget–Aztec dune field (Dickinson and Gehrels 2003, 2009; Rahl et al. 2003). Preservation of this tremendous thickness of wind-blown sand was made possible because of basin subsidence associated with Early Jurassic compressional deformation near the west margin of North America. This deformation caused the continental interior to flex downward, creating accommodation space for sediment accumulation (Allen et al. 2000).The Navajo Sandstone formed in a tropical arid setting with eolian, playa, and sabkha deposits (Mahaney et

al. 2004). Wetter restricted interdunes are also known, and are characterized by mudstone with planar bedding, dolomitic limestone, and sandstone (Irmis 2005b). Interdunes are best known from the vicinity of Moab. At locations such as this, interdunes ranged from fluvial sandstones to lake limestones. During periods of increased precipitation, vegetation spread (Wilkens 2008). Fossils are mostly associated with interdunes (Loope et al. 2004b). As time went on during the deposition of the Navajo Sandstone, the climate became drier (Irmis 2005b), although there were several periods of increased precipitation with durations on the order of tens of thousands of years (Milan et al. 2008).

The Navajo Sandstone is not extensively fossiliferous. To date, fossil organisms and traces represented include stromatolites, root traces (Irmis 2005b), pollen, spores, horsetails (*Equisetum*), ferns (Loope et al. 2004b), cycadeoids, conifers (Wilkens 2008), petrified wood (Irmis 2005b), possible fungal filaments (Mahaney et al. 2004), worm burrows (Stokes 1978), insect burrows, scorpion tracks (Loope et al. 2004b), unionid bivalves (Wilkens 2008), ostracodes, conchostracans (Harshbarger et al. 1957), a crocodylomorph, the small bipedal carnivorous theropod dinosaur *Segisaurus* (Irmis 2005b), at least two kinds of sauropodomorph dinosaurs ("prosauropods") including *Seitaad* (Sertich and Loewen 2010), a tritylodontid synapsid, and vertebrate trace fossils (Irmis 2005b). The assemblage may be a persistent remnant of the Kayenta Formation assemblage (Winkler et al. 1991). Sandstone dune deposits have mostly yielded vertebrates, while limestone interdune beds have mostly yielded plants and invertebrates (Stokes 1991), although arthropod tracks (*Octopodichnus*) are known from the dunes (Hunt and Lucas 2006). Interdune fossils include stromatolites, root casts, horsetails, ostracodes, and insect traces, with locally common petrified wood (Irmis 2005b).

Vertebrate trace fossils are much more common than body fossils in the Navajo Sandstone, and have been extensively described. More than 60 dinosaur track sites are known in the Navajo Sandstone across its depositional area (Loope 2006; Milan et al. 2008). The various vertebrate trace fossils have gone under different names and interpretations, depending on the author. Recent works (Irmis 2005b; Hunt and Lucas 2006; Loope 2006) seem to have settled on seven or eight primary ichnogenera: *Brasilichnium* (synapsid, but see Milan et al. 2008 for an interpretation as a sauropodomorph), *Batrachopus* (crocodylomorph), *Grallator* (small theropod), *Anchisauripus* (medium theropod), *Eubrontes* (large theropod), *Navahopus* (small ?sauropodomorph), *Otozoum* (larger sauropodomorph), and *Anomoepus* (ornithischian or "bird-hipped" dinosaur) (Irmis 2005b). A few oddities are also known, including lizard-like tracks (*Lacertipus*), a track once thought to be pterosaurian in origin ("*Pteraichnus*"), bird-like *Trisauropodiscus* tracks (Irmis 2005b), and undescribed large ornithopod (beaked herbivorous dinosaur) tracks (Santucci and Kirkland 2010) , and large vertebrate burrows thought to have been made by mammals and/or mammal relatives (Lucas et al. 2006c; Riese et al. 2011).

So far, large traces (*Anchisauripus*, *Eubrontes*, and *Otozoum*) are only known from interdunes, with smaller traces (*Batrachopus*, *Brasilichnium*, and *Grallator*) the only tracks found in dune sands (Loope 2006), although this differentiation may be due to preservational bias (Irmis 2005b). In general, the vertebrate trace fossils are consistent with the known body fossils.

A few fossil occurrences are known from CANY and adjoining areas. Interdune deposits at Horseshoe Canyon have stromatolites and other microbial features (Waiss 2005), and wood (Wilkens 2008). Complex vertebrate burrows are known from wet interdunes in and around CANY (Hasiotis et al. 2007). Carbonate mounds associated with interdune lake carbonate deposits have microbial structures, ostracodes, and fecal pellets (Dorney and Parrish 2009). Silicified wood has been reported from the Maze west of the Colorado River (Stokes 1991). There is evidence for microbial ("algal") marshes in the CANY–ARCH area (Baars 1983). Root casts are known from Navajo Sandstone outcrops on the Shafer Trail, which includes part of CANY (Huntoon and Dubiel 2000). Much larger fossil records are known from GLCA and the Moab area.

A variety of fossil resources are known from the Navajo Sandstone of GLCA and the immediate vicinity. Ten tracksites were discussed by Lockley et al. (1998). An undescribed reptile skeleton was found in a lacustrine limestone bed in the Navajo Nation section of GLCA (Santucci and Kirkland 2010). Traces of termite mounds were reported by McDonald and Santucci (2000), and wood has also been reported (NPS 1999b). Finally, two crocodylomorph skeletal specimens have been collected by the New Mexico Museum of Natural History and Science from north-central Arizona near the Utah border at West Canyon (Rinehart et al. 2001; Irmis 2005b), which runs into and is partially within GLCA.

Moab-area finds are unusual when viewed against Navajo Sandstone fossils from elsewhere, as they are dominated by plant fossils from lake settings. Fossil trees are particularly well known from Moab. Some are stumps in place, while others were knocked down and have been moved (Loope 1979). Those trees in place may have been killed by rising water tables, while those that were moved may have been knocked down by dunes that collapsed from heavy rains. Some lived for decades and attained diameters of nearly 1 m (3 ft) (Parrish and Falcon-Lang 2007). The Moab interdune localities also preserve invertebrate burrows (Riese et al. 2011), unusual complex burrows possibly made by tritylodontids (Lucas et al. 2006c), and a variety of plant fossils besides trees, including root traces (Riese et al. 2011), plant hash, and impressions and natural casts and molds of plant material (Wilkens 2008). When precipitation increased, dry uplands had cycadeoids, while araucarian conifers grew along lake margins (Wilkens 2008).

Quaternary sediments (Pleistocene–Holocene)

Quaternary sedimentary deposits at CANY are made up of rock fragments of clay to gravel sizes, as well as larger slumped landslide blocks. They include alluvial and terrace deposits associated with flowing water, dunes formed by wind action, and landslide deposits. Quaternary deposits in CANY are on the order of 3 to 12 m (10 to 40 ft) thick (Huntoon et al. 1982; Billingsley et al. 2002). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Paleoecological material is common at CANY although mostly unpublished, and there are numerous sites with mammoth, bison, shrub ox, musk ox, sloth, mountain goat, camel, or tapir bones, or rock art of mammoth or bison within a 100 km/60 mile radius of the park (Agenbroad 1997; Elias 1997). Summaries of Quaternary fossil sites in southern Utah and the vicinity include Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008). CANY has several published packrat middens, and thousands of Quaternary specimens have been collected from it. The great majority of these are plants and mollusks, although a few insects and bones of amphibians, reptiles, birds, and mammals (mostly rodents, but including a few mammoth specimens) have been collected (see "Park Collections" below).

Published Quaternary fossils from CANY are currently limited to twenty-six non-sediment samples used for dating Quaternary sediments at the park (Biggar and Adams 1987), packrat middens and associated fossils (Elias et al. 1992; Anderson et al. 2000; Coats et al. 2008), a mention of pollen from the Virginia Park–Needles area (Reheis et al. 2005), and passing mentions of sedimentary charcoal used for dating sediments (Goldstein et al. 2005; Reheis et al. 2005). The twenty-six non-sediment samples dated by Biggar and Adams (1987) included charcoal, wood, chenopod seeds, other plant matter, and some charred bone. They came from nine localities at (or in one case immediately south of) CANY, in the Needles fault zone and along Salt Creek. Five of the samples, including the seeds and the sample with bone, were essentially modern, and the rest were younger than 6,420 ± 730 radiocarbon years before present (8,765–5,610 calendar years before present).

Three fossil packrat middens have been described in detail from the park (Elias et al. 1992), and middens from four other midden localities have been mentioned briefly in the literature (Anderson et al. 2000; Coats et al. 2008). Packrat middens, collections of plant material and food waste constructed by packrats (*Neotoma* spp.) and cemented by their viscous urine, can be well-preserved in dry caves, and illustrate the environment of the builder's foraging range. Middens are important tools for reconstructing the ecology and climate of the late Pleistocene and Holocene of the southwestern United States (Strickland et al. 2001). The following brief descriptions of the CANY middens are based on information from the online U.S. Geological Survey/National Oceanic and Atmospheric Administration Packrat Midden Database (http://esp.cr.usgs.gov/data/midden/), supplemented by other publications where appropriate.

Three middens from the same area of eastern CANY, Dead Owl 1, Hoodoo 1 and Woodenshoe 1, were studied for arthropod taxa by Elias et al. (1992). Midden Dead Owl 1a is dated to 3,830 ± 70 radiocarbon years before present (4,420–4,075 calendar years before present), and had three beetle taxa and an ant. Midden Hoodoo 1 dates to 27,660 ± 340 radiocarbon years before present (32,790–31,270 calendar years before present), and had two beetle taxa and a lepidopteran. Midden Woodenshoe 1 is dated to 6,980 ± 120 radiocarbon years before present (8,010–7,605 calendar years before present), and had six beetle taxa. Anderson et al. (2000) noted the presence of three other midden localities at CANY: Island in the Sky, with middens dating to between 35,400 and 400 calendar years before present; The Maze, with middens between 800 and 0 calendar years before present; and Salt Creek, with middens between 39,900 and 0 calendar years before present. Finally, Coats et al. (2008) revised the age of the Salt Creek middens to between 38,000 and 10,000 calendar years before present, and reported 27 middens there, and added White Rim as a locality, with 12 middens between >48,000 and 380 radiocarbon years before present. Midden fossils from CANY are at Northern Arizona University (Flagstaff), per the CANY collections database (V. Webster, Southeast Utah Group museum curator, pers. comm., September 2009). The great bulk of this material remains unpublished.

Packrat middens are usually found in alcoves, caves, and rock shelters. Large alcoves and caves are also present at CANY (for example False Kiva), so there is the potential for finds similar to those of nearby Cowboy Cave, Bison Alcove at ARCH (a large rock shelter in the Entrada Sandstone, with numerous packrat middens and the skeletal remains of at least two bighorn sheep and a bison [Santucci et al. 2001]), and GLCA's alcove and cave sites. Santucci et al. (2001) presented an overview and cited selected examples of National Park Service fossils found in caves, including at ARCH and GLCA.

Cowboy Cave is approximately 11 km (7 miles) south of the Horseshoe Canyon Unit, on the same drainage, in a Navajo Sandstone cliff (Jennings 1980). The contents of the cave were extensively described by Jennings (1980). It is notable for having both extensive Pleistocene material and four distinct Holocene cultural layers above, spanning at least the last 13,000 years (humans present only since approximately 9,000 years ago). In the Pleistocene layer, various types of mammal dung and hair were found with packrat middens and a handful of bones, mostly bison, but also two juvenile mammoth tusks and a possible elk bone (Barnett and Coulam 1980; Hansen 1980; Jennings 1980; Spaulding and Peterson 1980). The cultural levels show human use of stone, bones, hair, hides, shells, plant

parts, and other local organic materials to make tools, ornaments, bags, and other artifacts (Hewitt 1980; Hull 1980a, 1980b; Janetski 1980; Lucius 1980; Weder 1980). Prehistoric dung at the cave mostly came from bison, but also includes feces from Shasta ground sloths (*Nothrotheriops shastensis*), elephants, horses, elk, elk or camel (Mead and Agenbroad 1992), rabbits, rodents (Hansen 1980), and humans (Hogan 1980).

Alcove and rock shelter sites are of great importance at GLCA. These sites are commonly found at the contact of the Kayenta Formation and Navajo Sandstone (McDonald and Santucci 2000). The best-known Quaternary site from GLCA is Bechan Cave (Northern Arizona University Quaternary Studies Program [NAU QSP] site 872; NPS GLCA Accession 81), found near the confluence of the Colorado River and Escalante River. Bechan Cave, containing 300 cubic meters (10,600 cubic feet) of fossil dung among other fossil resources, has been described in a number of publications. Its fossils include: fungal spores in droppings (Davis 1987, 2003; Davis and Shafer 2006); pollen (Davis 1990); charcoal, wood (Agenbroad and Mead 1989), seeds and plant fragments (Davis et al. 1984); insects (Elias et al. 1992); dung assigned to packrats, cottontail rabbits, Shasta ground sloths, mammoths, possibly horses, the extinct shrub ox *Euceratherium collinum*, bighorn sheep, and extinct mountain goats (*Oreamnos harringtoni*); hair from shrews, packrats and other rodents, bats, coyotes, bear, ground and mylodont sloths (*Glossotherium* and *Nothrotheriops*), mammoths, horses, bison, bighorn sheep, and deer; bones from toads, snakes, grouse-sized birds, a variety of rodents, rabbits, and *Euceratherium*; and packrat middens covering the last 12,000 years (Agenbroad and Mead 1987, 1990a; Mead and Agenbroad 1992; Santucci et al. 2001). Human artifacts are also known from younger levels of the cave (Agenbroad et al. 1989). Fourteen plant samples and seven dung samples from mammoths and shrub oxen have been dated from Bechan Cave, yielding a range that spans from 13,505 ± 580 radiocarbon years before present (17,835–14,460 calendar years before present) to 11,630 ± 150 radiocarbon years before present (13,800–13,215 calendar years before present) (Mead and Agenbroad 1992). Nine other alcove sites have been described from GLCA: BF Alcove, Cottonwood Alcove, Grobot Grotto, Hooper's Hollow, Mammoth Alcove, Oak Haven, Oakleaf Alcove, Shrubox Alcove, and Withers Wallow (Mead and Agenbroad 1992). Most of these are near Bechan Cave in the Forty-Mile Canyon (Agenbroad and Mead 1990b), except for Cottonwood Alcove and Oakleaf Alcove on the San Juan Arm (Mead and Agenbroad 1992). These locations are also mostly known for Pleistocene mammal dung.

Another potential source for Quaternary fossils are caliche (calcium carbonate hardpan) beds capping the Entrada Sandstone in the western Horseshoe Canyon unit. These beds are interpreted as fluvial deposits. There are mollusk fossils eroding out of conglomeratic deposits in these beds in at least one location near the western boundary of the Horseshoe Canyon unit, probably in adjoining BLM land (G. Cox, pers. comm., March 2010).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within CANY. However, they are known to preserve fossils elsewhere, and future field investigations within the park may recover fossils from one or more of them.

Hermosa Group: Paradox Formation (Middle Pennsylvanian)

The Paradox Formation is a thick heterogeneous unit. At CANY, the full thickness does not outcrop, with the upper part of the unit being the oldest rocks exposed in the park (Sumsion and Bolke 1972); about 120 m (400 ft), mostly gypsum with some dolomite and black shale, can be seen in and around the mouth of Gypsum Canyon in southern CANY (Baars 2010). The CANY rocks include, from oldest to youngest, the Akah, Desert Creek, and Ismay stages. The Akah Stage is composed of gypsum, dolomite, and black shale, followed by evaporites of the Desert Creek Stage, and carbonates of the Ismay Stage. Deposition of the park's exposed segment occurred during the Middle Pennsylvanian, with the Akah Stage attributed to the middle Middle Pennsylvanian (mid Desmoinesian stage, about 307 Ma) (Baars 2010). The Paradox Formation's depositional area was a restricted basin with cyclic aridity, represented by cycles of rocks laid down under normal marine to more saline conditions (Wengerd 1962). It has been described as a "dead sea" (Baars 1983). While marine in origin, evaporites are common, and invertebrate fossils are rare (Rueger 1996). Because it was based on evaporitic beds, authors may choose to not map the Paradox Formation where such beds are not present, even if correlative layers are (Franczyk et al. 1995). Although well-known for its salt deposits (29 cycles of salt deposition are known), the Paradox Formation lacks salt exposures in the Moab region. However, salt at depth has been a major factor in the surface geology of the CANY–ARCH region (Baars 1983).

While fossils have not yet been reported from the Paradox Formation of CANY, there are several reports from the area. Palynomorphs have been found from several cores north (Cane Creek), northeast (Shafer No. 1), and east (Gibson Dome) of CANY, representing arborescent and herbaceous coal swamp ferns and gymnosperms on the western shore of the seaway (Kosanke 1995; Rueger 1996). Brachiopods are known from this unit on the Colorado River between Moab and Hite (GLCA) (Rigby et al. 1971). Rugose corals, bryozoans, brachiopods, crinoids, and trilobites have been reported from an exposure at ARCH (Swanson et al. 2005). Other fossils known from the Paradox Formation include stromatolites (Chidsey and Eby

2000), algae, fusulinid foraminifera, sponges, echinoids, conodonts, and pellets (Ritter et al. 2002). Additionally, the Paradox Formation is known as an oil-producing rock unit (Doelling 2010). Microbial mounds are regarded as the oil sources, and formed on shallow carbonate shelf settings surrounding the evaporites (Baars 1983). Microbial mound limestones are known in the CANY area (Fallin 1985).

Glen Canyon Group: Wingate Sandstone (Upper Triassic–Lower Jurassic)

The Wingate Sandstone is an eolian sandstone unit that spans the Triassic–Jurassic boundary (Lucas et al. 2006a). Limestone lenses are also present (Harshbarger et al. 1957). At CANY, it is 90 to 120 m (300 to 400 ft) thick, and is composed of cliff-forming brown sandstone, mostly exposed in the northern part of the park. It is known for its cliffs, capped by the Kayenta Formation. Although most outcrops are interpreted as wind-blown deposits, some may be fluvial at CANY (Baars 2010). Sand sheet and sabkha deposits are present in the area as well (Huntoon and Dubiel 2000). Wind blew from the northwest (Patterson et al. 1989). The Wingate Sandstone grades out of the underlying Church Rock/Rock Point Member of the Chinle Formation (Tanner 2003), grades into the overlying Kayenta Formation (Anderson et al. 2010), and interfingers at its western contact with the Dinosaur Canyon Member of the Moenave Formation (Donohoo-Hurley et al. 2007). After much debate concerning the age of the Wingate Sandstone and other Glen Canyon Group units, it was thought that the Wingate Sandstone was entirely Jurassic in age, but the discovery of a skull of the phytosaur *Redondasaurus* at the base of the formation in Lisbon Valley shows that at least some of the unit is Late Triassic in age (Lucas et al. 1997a). Trace fossils suggest that the Triassic–Jurassic boundary is in the middle to upper 75 m (246 ft) of the formation (Lucas et al. 2006a).

The Wingate Sandstone has had a complicated history (Colbert and Mook 1951). It was originally defined to contain much the same rocks as what was later named the Entrada Sandstone, but the Entrada Sandstone eventually took over most of these rocks, including the original type section for the Wingate Sandstone (at Fort Wingate, New Mexico). Furthermore, the Wingate Sandstone was later divided into the lower Rock Point Member and the upper Lukachukai Member, but where the Rock Point Member has been assigned to the Chinle Formation (as discussed above), the Lukachukai Member is abandoned as redundant (Dubiel 1989).

In general, fossils are uncommon in the Wingate Sandstone, with the phytosaur skull mentioned above being the only body fossil of note, but trace fossils are known from multiple levels in Utah, Arizona, and Colorado. Trace producers include invertebrates (Odier et al. 2004), mammal-like animals, nondinosaurian reptiles, and theropod, prosauropod, and possible sauropod ("brontosaur") dinosaurs (Odier et al. 2004; Lucas et al. 2006a, 2006b). Mammal-like tracks are often found nearly alone and are confined to dunes (Odier et al. 2004). There appears to be distinct Late Triassic and Early Jurassic assemblages, with mammal-like tracks limited to the Late Triassic and prosauropod tracks appearing in the Early Jurassic (Lucas et al. 2006a, 2006b). The nearest fossil occurrences are at GLCA and the vicinity, where track sites have been found. Three have been described, one associated with a stromatolitic playa (Lockley et al. 1998).

San Rafael Group: Carmel Formation (Middle Jurassic)

The Carmel Formation is a marine and terrestrial Middle Jurassic unit. West of CANY, it is a marine unit, but it becomes terrestrial in the east (Blakey et al. 1983). At CANY, it is only exposed in the Horseshoe Canyon Unit (Huntoon et al. 1982; Dubiel et al. 1985). Although it can be divided into several members (Doelling and Willis 2006; Sprinkel et al. 2011), only one unit is present in the CANY area: the Dewey Bridge Member (Baars 2010). In the past, this interval was regarded as the lowest part of the overlying Entrada Sandstone (Williams 1964), and many older references (Lohman 1975; Baars 1983; Doelling 1988) call the rocks the Dewey Bridge Member of the Entrada Sandstone. The interval is lithologically similar to the overlying Entrada Sandstone, but is stratigraphically continuous with the marine Carmel Formation (Doelling 2010). It has a disconformable contact with the Navajo Sandstone in the CANY area (Huntoon et al. 1982), and is composed of yellowish-orange to reddish-brown sandstone, reddish-brown mudstone, and local limestone and gypsum, between 30 and 190 m (100 and 625 ft) thick (Dubiel et al. 1985). It forms slopes or ledgey slopes (Huntoon et al. 1982). In the area of the Horseshoe Canyon unit, it represents a tidal flat and sabkha setting (Dubiel et al. 1985), with the sea to the west (Baars 1993). The Carmel Formation dates to the Middle Jurassic, with an age of approximately 169 to 162 Ma based on radiometric dates from central and southwestern Utah (Sprinkel et al. 2011).

The Dewey Bridge Member of the Carmel Formation is not known to be fossiliferous at CANY, although fossiliferous Carmel Formation limestone is known northwest of CANY, in the San Rafael Swell (Baker 1933). Marine fossils are most common in the lower part of the formation, while terrestrial trace fossils are known from the upper part of the formation, particularly from northeastern Utah. Marine fossils include dinoflagellates (a type of plankton) (Sprinkel et al. 2011), algae, foraminifera, radiolarians, sponges, corals, bryozoans, brachiopods, bivalves, ammonite cephalopods, gastropods, ostracodes, crinoids, echinoids, sea stars, and trace fossils including pellets and worm tubes (Imlay 1964; Blakey et al. 1983; Nielson 1990; Wilson 1998). The Carmel Formation's marine fossils are sometimes found in association with what had been cavities (Baker and Wilson 1999) or ostreoliths (oyster balls) (Wilson et al. 1998). Terrestrial fossils include spores, pollen (Sprinkel et al. 2011), and crocodilian and dinosaur tracks (Lockley and Hunt 1995), from northeastern Utah (Santucci and Kirkland 2010).

San Rafael Group: Entrada Sandstone (Middle Jurassic)

The Entrada Sandstone has not yet been depicted on a geological map within CANY, but is present in the western part of the Horseshoe Canyon unit. Some outcrops have been mapped as Quaternary sand dunes (G. Cox, pers. comm., March 2010). As mentioned above, the CANY outcrops are capped by Pleistocene-age caliche deposits. The Entrada Sandstone is dominated by red sandstone, with some silt (Harshbarger et al. 1957). It is divided into multiple members depending on location. In the CANY region, it was described by Huntoon et al. (1982) as a pale-orange fine-grained sandstone, 70 to 150 m (225 to 250 ft) thick, including the Dewey Bridge Member, Slick Rock Member, and Moab Tongue; the Dewey Bridge and Moab units are now assigned to the Carmel Formation and Curtis Formation, respectively (Doelling 2010). The formation lacks age-diagnostic fossils and is regarded as late Middle Jurassic in age (Lucas et al. 1985). The climate was semi-arid to arid (Ekdale and Picard 1985) and hot, with sparse vegetation. The Entrada dunes made up the last major erg of the area, and formed when Utah was about 22° N (Loope 2008).

The Entrada Sandstone is not known to be fossiliferous at CANY, and fossils are generally rare in the formation. Many reports of fossils from the Entrada Sandstone are from the Moab Member, which is now regarded as part of the Curtis Formation (Doelling 2010). These include the well-known megatracksite of the Moab area (Lockley 1991; Lockley et al. 2007), and the crocodylomorph *Entradasuchus* (Hunt and Lockley 1995). Trace fossils have been reported from the Escalante Member in south-central Utah, including rare root traces, insect burrows (Loope 2008), vertebrate burrows, traces of sand-swimming vertebrates (Graverson et al. 2007), and theropod (large *Megalosauripus* and smaller *Therangospodus*) and sauropod (cf. *Brontopodus*) dinosaur tracks (Foster et al. 2000). Hoggan (1970) reported bivalves and crinoids in marine facies of the Entrada Sandstone in the Vernal area.

Park Collections

CANY fossils are held at the park and elsewhere. The CANY collections database includes 1094 catalog numbers for tens of thousands of specimens; at least 25,647 objects are included, but many lots do not have item counts. All but 7 of the catalog numbers pertain to Quaternary specimens. The Quaternary specimens include gnetophytes (13 catalog numbers), conifers (99), dicots (166), monocots (43, typically grasses), unidentified plants (143), bivalves (20), gastropods (479), insects (17), amphibians (1), lizards (2), rodents (31), mammoths (4), unidentified mammals (17), unidentified animals (11), composite specimens (34), unknown (3), and bulk sediments (4). Aside from the bivalves and gastropods, which derive from two Northern Arizona University (NAU) sites known as Needles Project and Snail Knob, and the mammoth bones, which were found in the Horseshoe Canyon Unit, these specimens were collected by NAU researchers from packrat middens. Some of these mollusks and midden specimens may be essentially modern. The remaining seven catalog numbers include: CANY 105, five crinoid stems from an unspecified unit; CANY 112, three brachiopods from a limestone bed of an unspecified unit; CANY 164, three pieces of petrified wood from an unspecified unit; CANY 34479, bones of the aetosaur *Stagonolepis* from the Chinle Formation (this is the CANY record for the aetosaur collected by NMMNH as described above, NMMNH P-27083); CANY 34480, gastropod impressions from the Sinbad Limestone (this is the CANY record for the gastropods collected by NMMNH as described above, NMMNH P-26938); and CANY 36367 and 36368, two rocks with fern impressions probably from the Chinle Formation. Most specimens from CANY are in the collections of NAU (middens) or the Museum of Northern Arizona (MNA; Flagstaff) (mollusks). The four mammoth specimens and the five pre-Quaternary specimens not in NMMNH collections are kept in Southeast Utah Group collections (V. Webster, pers. comm., September 2009).

Several other institutions have CANY specimens in their collections. As mentioned in the Honaker Trail Formation section, six specimens of the sponge *Haplistion sphaericum* from CANY are in four institutions. Two are in the collections of the National Museum of Natural History (Washington, D.C.) (USNM 166524 and 166525), two are in the collections of Brigham Young University (Provo, Utah) (BYU 1513 and 1514), one is at the University of Utah (Salt Lake City, Utah) (number not given), and the other is in CANY collections (Rigby and Stokes 1971) (although note that this specimen was not included in the CANY database). As noted, the MNA has all of the Quaternary mollusk fossils in the CANY database, and the packrat midden material is at NAU (V. Webster, pers. comm., September 2009). The collections of the New Mexico Museum of Natural History and Science (Albuquerque) include gastropods (NMMNH P-27083) and aetosaur fossils (NMMNH P-27083) (Heckert et al. 1999; NMMNH online database no date). Finally, the University of California, Los Angeles has palaeoniscid fish fossils from CANY (UCLA collections) (Sumida et al. 1999a).

Cultural Resource Contexts

CANY has numerous archeological sites and a record of human presence covering at least the last 10,000 years (NPS no date), including Clovis points (G. Cox, pers. comm., March 2010; the Clovis culture is one of the oldest of the continent), so it would not be surprising to find paleontological resources with or as cultural resources. For example, a bag made of antelope hide, dating to between 770 and 970 AD, was recovered in 2005 from a dune at the Horseshoe Canyon Unit. Among its contents were marsh-elder and other seeds, three leather pouches, a pressure flaker made from an antler, a chert biface, and chert flakes (Geib and Robins no date). As noted earlier, nearby Cowboy Cave was occupied by humans over multiple intervals spanning approximately 9,000 years, with numerous artifacts made from the contemporaneous biota, providing a model for potential finds at CANY. Stokes (1972) interpreted three petroglyphs in southeastern Utah, including one near Moab, as representing proboscideans (possibly mastodons), but such identifications are extremely controversial. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for CANY, ARCH, HOVE, and NABR during May 1999 (NPS 1999a), and produced a geologic resource evaluation report for CANY in 2005 (KellerLynn 2005). The participants of the scoping meetings developed a series of recommendations for future work including continuation of paleontological resource inventories, increased protection, staff training, and the completion of a paleontological resource management plan.

- The park should consider future field inventories for paleontological resources to more fully document *in situ* occurrences of fossils. The park may consider a formal site documentation and condition assessment for significant fossil localities. Monitoring of significant sites should be undertaken at least once a year in the future. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- Within the Northern Colorado Plateau Network (NCPN), the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and increase understanding of this important fossil-bearing unit. A similar USGS-NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Capitol Reef National Park

Capitol Reef National Park (CARE), Utah, was proclaimed August 2, 1937 as a national monument, and was redesignated as a national park December 18, 1971. It protects a 160-km (100-mile) long geologic structure called Waterpocket Fold, a giant monocline (step-like fold) in Mesozoic sedimentary rocks that steps down from west to east. CARE encompasses a variety of geological features, including arches, domes, and pillars (CARE Figure 1). Rock art and other archeological resources left by the Fremont Culture are also found within CARE, as well as Mormon settlements from the late 19th century, such as the settlement at Fruita. The park's name refers to dome-shaped white caprock on the fold ("Capitol" like the Capitol Dome in Washington, D.C.), and to the travel barrier ("Reef") presented by Waterpocket Fold. CARE's boundaries changed July 2, 1958; January 20, 1969; and December 18, 1971. It today encompasses 97,895.18 ha (241,904.26 acres), all but 271.13 ha (669.97 acres) of which are under federal administration (CARE Figure 2).

Geologic Background

CARE is located in south-central Utah, with the bulk of its acreage in western Wayne and central and eastern Garfield counties. Small sections are also present in extreme southeastern Sevier County and extreme southwestern Emery County. Geographically, the park is a long and relatively narrow strip that tapers as it stretches southeast from the junction of Sevier, Emery, and Wayne counties to an area 16 km (10 miles) west-northwest of the Glen Canyon National Recreation Area (GLCA) Visitor Center. CARE is not near any major population centers, but it is surrounded by numerous large geographic features. To the northeast are the San Rafael Swell and Desert, to the east are the Henry Mountains, to the southeast is Glen Canyon, running along the southwest are the Circle Cliffs and more distantly the Kaiparowits Plateau, and on the west is the Aquarius Plateau and the associated Boulder Mountain and Thousand Lake Mountain. The Fremont River cuts across northern CARE as it flows east. Several federal properties border CARE: Fishlake National Forest is on the north and northwest borders, GLCA is on the south and southwest, Grand Staircase–Escalante National Monument (GSENM) borders on the southwest and west, and Dixie National Forest borders on the west. The park is situated on the western margin of the Colorado Plateau, with the High Plateaus on the west (Graham 2006). A geologic scoping session, coordinated by the Geologic Resources Division of the National Park Service (NPS), was held for CARE during September 1999 (NPS 1999a), leading to a geologic report for the park (Graham 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The geology of CARE and the vicinity has been described in numerous publications. References that are useful for understanding the geology of CARE include Gregory and Anderson (1939), Marine (1962), Smith et al. (1963), Davidson (1967), Olson and Muench (1972), Petersen and Pack (1982), Collier (1987), Billingsley et al. (1987), Nelson (1989), Harris and Tuttle (1990), Doelling and Willis (1999), NPS (1999a), Doelling (2004), Mathis (2004), Marchetti and Cerling (2005), Graham (2006), Doelling and Kuehne (2007), Martin et al. (2007), McLelland et al. (2007), Sorber et al. (2007), Kjemperud et al. (2008), and Morris et al. (2010). References with particular bearing on the paleontology of CARE include Peabody (1956), Imlay (1964), Lammers (1964), Sohl (1965), Ash (1975, 1982, 1993), Dean (1981), Petersen and Roylance (1982), Lupe (1984), Huntoon and Kamola (1988), Kamola and Chan (1988), Cole (1992), Cole and Henderson (1993), Burres (1994), Cole et al. (1997), Kirby and McAllister (1998), Santucci et al. (1998), Cole and Murray (1999), Cobban et al. (2000), Koch and Santucci (2002), Eisenberg (2003), Mickelson et al. (2006a), and Santucci and Kirkland (2010).



CARE Figure 1. The Castle, one of CARE's many striking landforms; it is composed of an eroded remnant of Wingate Sandstone (VLS).

The geologic record exposed at CARE covers the lower Permian to the Upper Cretaceous, and the Quaternary (see the appendix for a geologic time scale). The oldest exposed rocks, represented by the Cutler Group, were deposited shortly after 300 Ma (million years ago), under terrestrial conditions in the vicinity of a fluctuating sea. The Cutler Group was deposited during a period of tectonic activity. To the east, the Paradox Basin was

sinking while farther east the Uncompahgre Uplift began to rise and provided sediment to fill the basin (Melton 1972; Blakey 2008; Baars 2010). The area including CARE was near the equator at this time (Morris et al. 2010). Under dry equatorial conditions, ergs (giant dune fields) formed across southeastern Utah, with CARE on the northwestern margin of the ergs (McKee 1979; Blakey 2008; Blakey and Ranney 2008). Beyond the sand was a shallow sea, which periodically transgressed (advanced over the land; the opposite condition is a regression) into what is now CARE during the Early Permian. The Kaibab Limestone, the youngest Permian rocks preserved at CARE, record an extensive marine transgression (Morris et al. 2010). A gap or hiatus of some 20 million years follows the Permian rocks at the park.

Triassic rocks at CARE are principally represented by the Moenkopi Formation of the Early–Middle Triassic and the Chinle Formation of the Late Triassic. Southern Utah during the Early and Middle Triassic was a broad coastal plain that was occasionally flooded by a sea to the west (Stewart et al. 1972a; Blakey et al. 1993; Hintze and Kowallis 2009). The top of the Moenkopi Formation is separated from the Chinle Formation by millions of years, with intervening erosion (Pipiringos and O'Sullivan 1978). The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time (Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003).

Chinle Formation rocks show that the area gradually became drier over time as the supercontinent Pangaea moved north. During its deposition, the Four Corners area of the Southwest moved from near the equator to arid-zone latitudes. What had been a setting of rivers, lakes, and deltas turned arid (Tanner 2003). This is represented at CARE by the first of several giant ergs, the Wingate Sandstone of the Glen Canyon Group (Blakey and Ranney 2008). The Wingate Sandstone erg of the Late Triassic and Early Jurassic was ended by a more humid period recorded in the Kayenta Formation. The Kayenta Formation represents a broad, gentle flood plain separated from the sea to the west (Blakey 1994; Peterson 1994). About 190 Ma (million years ago), this setting was gradually replaced by the vast erg that became the Navajo Sandstone (Harshbarger et al. 1957; Dickinson and Gehrels 2003, 2009; Rahl et al. 2003; Lucas et al. 2005), the largest erg known from North America (Blakey 2008; Blakey and Ranney 2008). This Sahara-like desert was found throughout the Four Corners area into Nevada and Wyoming (Hintze and Kowallis 2009). At this time, the latitude of the area was only about 10° N, near the western edge of the supercontinent Pangaea (Loope et al. 2004a). Strong cross-equatorial winds blew from the northwest during the winter, driving dune movement, and monsoonal rains occurred during the summer (Loope et al. 2004a, 2004b).

Eventually another shallow sea began to encroach on the CARE area, inundating the Navajo erg and setting the stage for another series of transgression-regression cycles and their resulting shallow marine and coastal terrestrial deposition (Morris et al. 2010). The shallow sea, known as the Sundance or Carmel–Twin Creek Seaway, moved into the area from the north (Imlay 1980), reaching as far south as northern Arizona. The Carmel–Twin Creek Seaway was active at about the same time as an outbreak of volcanic activity on the southern margin of the Colorado Plateau dated to approximately 170 Ma (Blakey and Parnell 1995); minor crustal extension also occurred in central Utah at this time (Blakey 1989; Marzolf 1993). In the CARE area, the seaway was largely hypersaline (Blakey 2008; Blakey and Ranney 2008). The Entrada Sandstone erg that replaced it would be the last large erg preserved at CARE. It was in turn covered by the last marine transgression of the Jurassic, represented by the Curtis and Summerville formations (Morris et al. 2010). Aridity decreased through the Late Jurassic (Lockley and Wright 2003) as Pangaea continued moving north; southern Utah reached 22° N during this time (Loope 2008). The more humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the last geologic formation of the Jurassic in CARE, the well-known Morrison Formation. By the time it was deposited, the area had moved out of the low subtropical latitudes (Lockley and Mickelson 1997). The Morrison Formation is thought to have been a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004).

The Sevier Orogeny (mountain-building event) began during the Jurassic in response to the collision of the North American and Farallon plates along the western margin of North America (Morris et al. 2010), and persisted into the Paleogene (Hintze and Kowallis 2009). During this orogeny, thrust faulting west of CARE affected sedimentary rocks to the east, eventually forming large mountains west of CARE in central Utah (Morris et al. 2010). These uplifts provided sediment for the formations deposited during the Cretaceous (Dickinson and Gehrels 2008), and the weight of the mountains depressed the adjacent areas, forming the foreland basin (including CARE) that would become a long-lived depositional basin (Morris et al. 2010). The Early Cretaceous is represented sparsely at CARE in comparison to the periods of time before and after, with only part of the Cedar Mountain Formation present. The Cedar Mountain Formation is another terrestrial fluvial unit (Peterson et al. 1980).

Terrestrial processes continued until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Sandstone signals the onset of several cycles when the Western Interior Seaway transgressed and regressed over the land (Gustason 1989; Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of

numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Marine deposition at CARE from the Seaway is represented by the fine-grained members of the Mancos Shale, with periodic regressions represented by sandy members deposited in shoreline settings. At the same time the Seaway was present, volcanoes to the west introduced large amount of ash, which over time have altered to bentonitic clays. The Seaway began to retreat altogether from CARE during deposition of the upper part of the formation, and the depositional setting transitioned to coastal plains (Morris et al. 2010). While formations from between approximately 80 Ma and the Pleistocene are not preserved in CARE, Late Cretaceous and Cenozoic formations of this age are present in the surrounding area (Doelling and Willis 1999; Doelling 2004; Doelling and Kuehne 2007; Morris et al. 2010).

Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of the Laramide Orogeny (Cather 2003). This mountain building event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (Tindall 2000; Davis and Bump 2009). One such monocline is Waterpocket Fold (Morris et al. 2010). Other examples of nearby Laramide structures include the Circle Cliffs and Monument uplifts and the Kaiparowits and Henry Mountains structural basins (Anderson et al. 2010). Extensive downcutting followed recent uplift; the incision rates on the Fremont River in CARE are some of the highest on the Colorado Plateau (Marchetti and Cerling 2005).

In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent local igneous activity during the late Eocene, Oligocene, and Miocene. Igneous bodies were emplaced at depth, forming the nearby Henry Mountains between 31 and 23 Ma (Nelson 1998). Boulder and Thousand Lake mountains are capped by two sets of volcanic rocks: latite flows from approximately the end of the Oligocene, between 26 and 23 Ma, and Miocene-age basalt flows (Doelling and Kuehne 2007). Several relatively small intrusive dikes and sills dated to between 4.6 and 3.7 Ma cut strata in the northern part of the park (Doelling 2004; Doelling and Kuehne 2007; McLelland et al. 2007; Sorber et al. 2007). These may represent the roots of a former volcanic field (Morris et al. 2010).

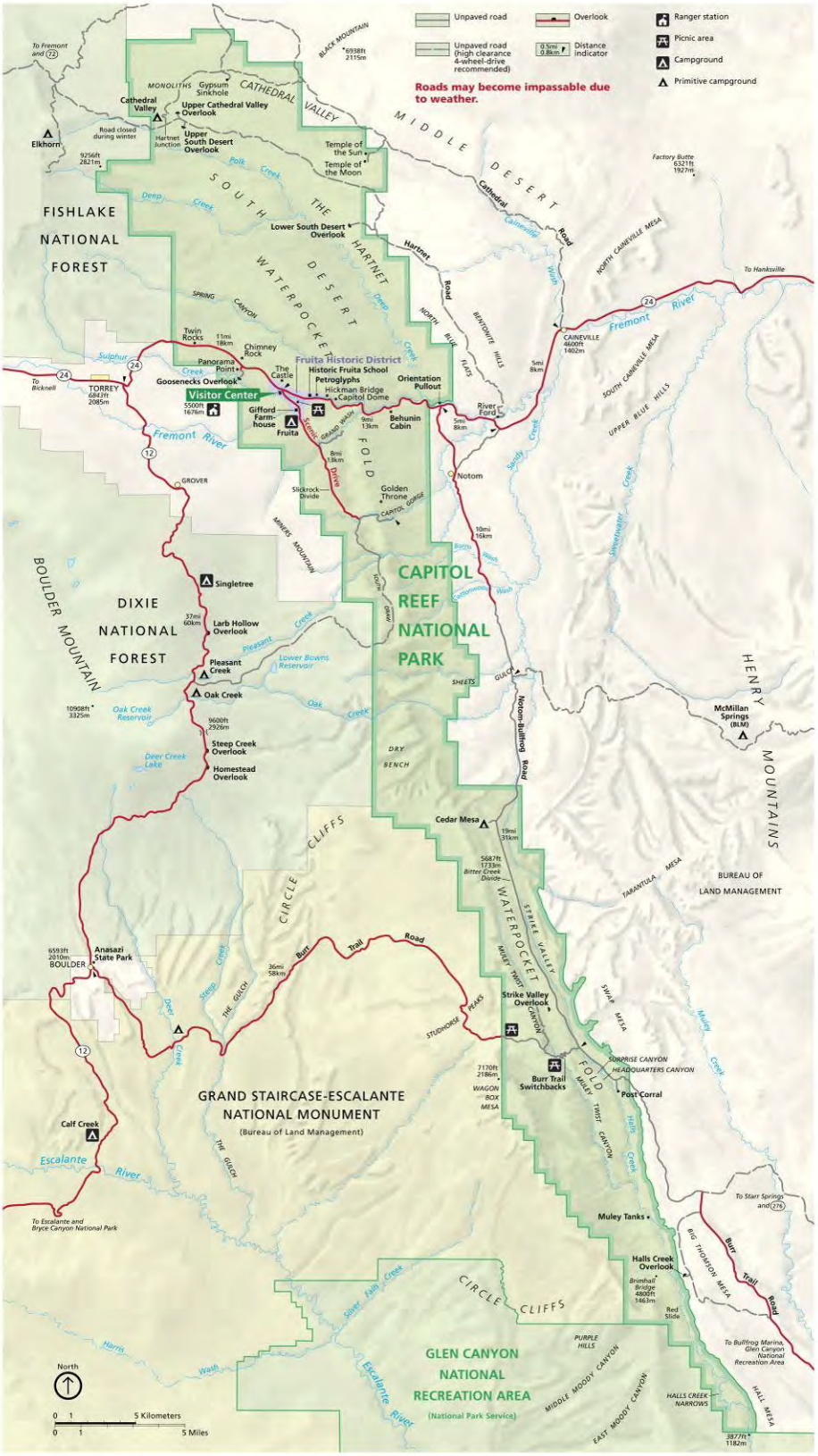
As the Pleistocene began approximately 2.6 Ma and glaciers formed over wide areas, alpine glaciation occurred in central Utah where elevations and precipitation were sufficient (Richmond et al. 1965). The nearest area that had substantial glaciation is Boulder Mountain, glaciated from its flat top at approximately 3,350 m (11,000 ft) above sea level down to 2,730 m (8,950 ft) (Gould 1939). Two primary periods of glaciation are differentiated in central Utah (Osborn and Bevis 2001), the older Bull Lake glaciation (approximately 150,000 years ago) (Marchetti et al. 2005), and the more recent Pinedale glaciation associated with the last glacial maximum at approximately 21,000 ± 2,000 years ago (Marchetti et al. 2007). Only Pinedale glacial deposits have been identified for certain on Boulder Mountain; putative Bull Lake deposits (the Carcass Creek Drift) are actually debris flow deposits that postdate the Bull Lake phase (Marchetti et al. 2005).

The long post-Cretaceous deposition gap at CARE ended sometime during the Pleistocene, preserving evidence of the biota that lived there during the Ice Age. The megafaunal extinctions on the Colorado Plateau claimed such animals as the mammoth, the ground sloth *Nothrotheriops shastensis*, the shrub ox *Euceratherium collinum*, and the mountain goat *Oreamnos harringtoni*. Dated dung deposits from the Colorado Plateau indicate that they went extinct between 11,800 and 11,000 radiocarbon years ago (Mead and Agenbroad 1992). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate. The Colorado Plateau was cooler and more moist during the last glacial period, becoming more arid afterward. It has been most arid during the last 1,500 years (Elias et al. 1992). It had been thought that both megafauna and human cultural sites more than 6,000 years old were rare on the Colorado Plateau, but this is now known to be inaccurate (Agenbroad 1997). Numerous archeological sites are known from CARE (Smith et al. 1963), and rock art produced by the Fremont Culture is well-known within the park (Mathis 2004). The Fremont people lived in the region between 1,500 and 700 years ago, farming on the Fremont River floodplain (Graham 2006).

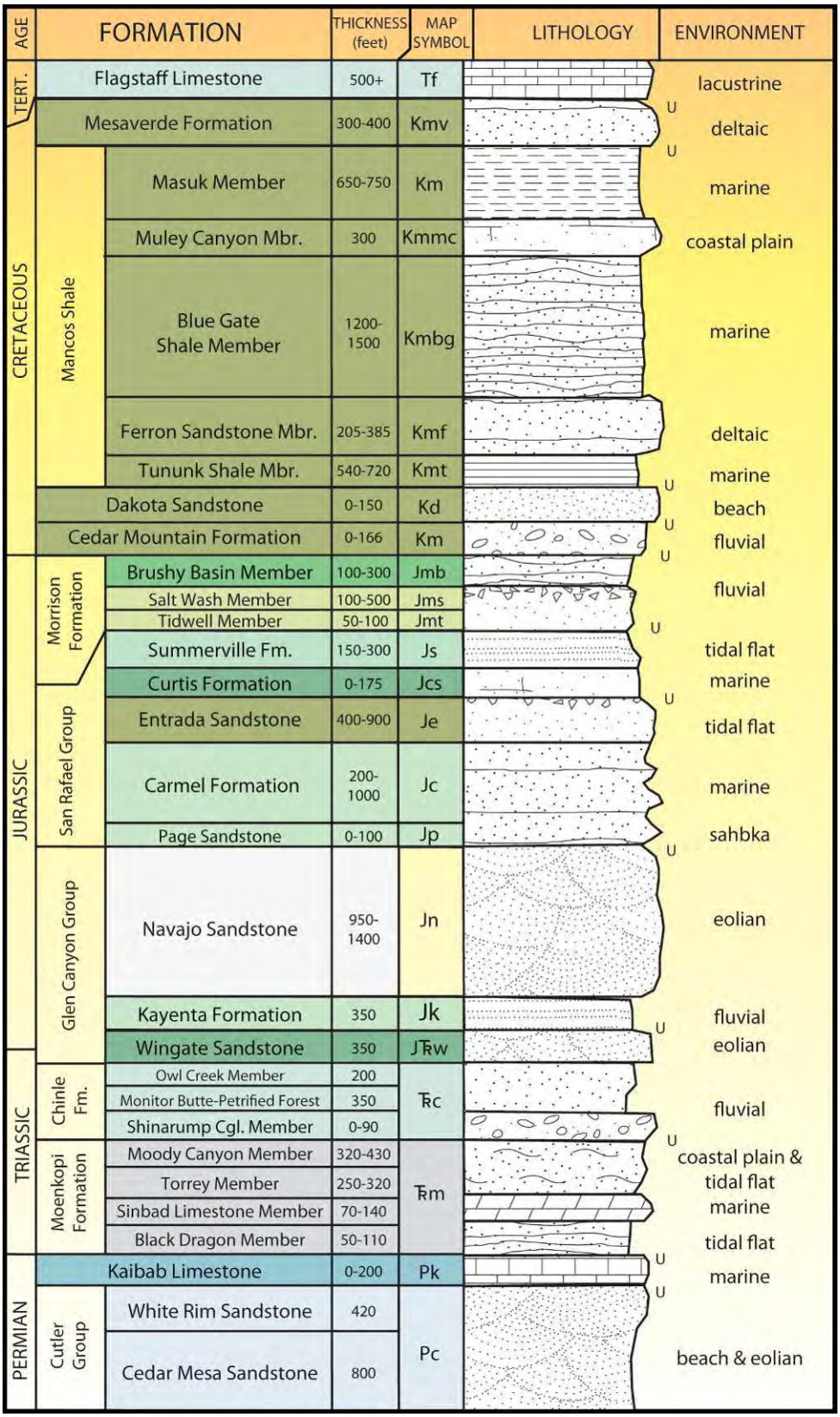
Bedrock units exposed at CARE include, from oldest to youngest: the Cutler Group (Lower Permian, regarded as the Cedar Mesa Sandstone and White Rim Sandstone); the Kaibab Limestone (Lower Permian); the Moenkopi Formation (Lower–Middle Triassic); the Chinle Formation (Upper Triassic); the Wingate Sandstone of the Glen Canyon Group (Upper Triassic–Lower Jurassic); the Kayenta Formation and Navajo Sandstone of the Glen Canyon Group (Lower Jurassic); the Page Sandstone, Carmel Formation, and Entrada Sandstone of the San Rafael Group (Middle Jurassic); the Curtis and Summerville formations of the San Rafael Group and the Morrison Formation (Upper Jurassic); the Cedar Mountain Formation (Lower–lowest Upper Cretaceous); the Dakota Sandstone, Mancos Shale, and Tarantula Mesa Sandstone (Mesaverde Formation of Billingsley et al. [1998], Graham [2006], and Morris et al. [2010]) of the Mesaverde Group (Upper Cretaceous); and Pliocene-age igneous rocks. Surficial units include a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene). In general, the park's bedrock becomes older going east to west, with Cretaceous and younger Jurassic rocks found in the eastern part of the park, the Lower Jurassic Glen Canyon Group forming the spine of the park, and Triassic and Permian rocks found in the western part of the park (see CARE Figure 3 for a stratigraphic column and CARE Figure 4

for a geological map) (Billingsley et al. 1987; Nelson 1989; Doelling and Willis 1999; Doelling 2004; Graham 2006; Doelling and Kuehne 2007; Martin et al. 2007; McLelland et al. 2007; Sorber et al. 2007; Morris et al. 2010).

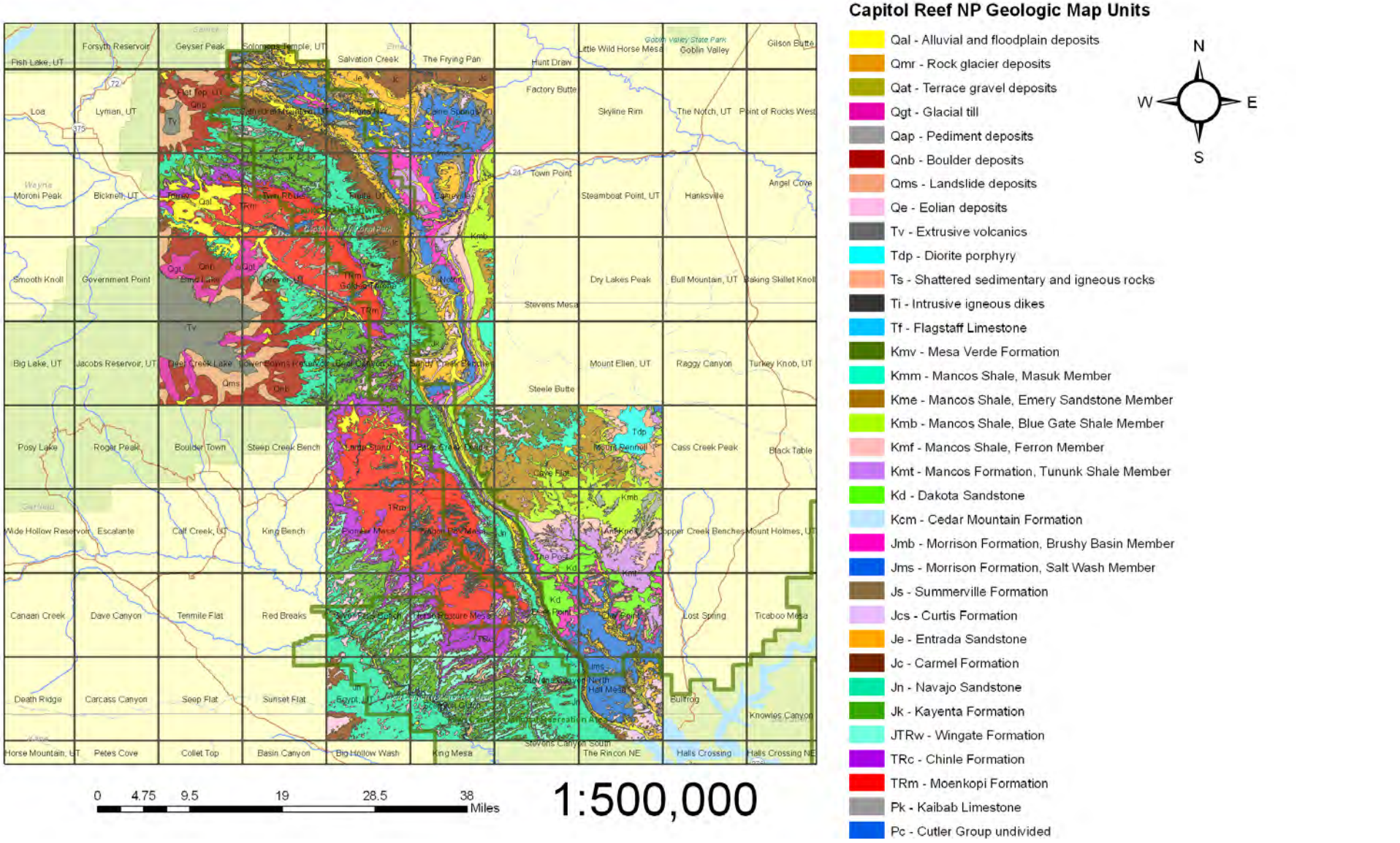
All of these rocks units, with the exception of the Pliocene igneous rocks, are potentially fossiliferous, and many of them have produced fossils within CARE. Paleontological resources at CARE are known from the White Rim Sandstone of the Cutler Group (marine shells and trace fossils), the Kaibab Limestone (marine invertebrates), the Moenkopi Formation (marine invertebrates and terrestrial trace fossils), the Chinle Formation (petrified wood and other plant fossils, bone fragments, trace fossils), the Kayenta Formation (tetrapod tracks), the Navajo Sandstone (stromatolites, a type of layered sedimentary structure formed from sediment trapped by microorganisms), the Carmel Formation (marine invertebrates), the Morrison Formation (petrified wood, dinosaur bones, and trace fossils), the Dakota Sandstone (bivalves), the Mancos Shale (marine invertebrates), and Quaternary sediments (packrat middens) (CARE Table 1). Fossil specimens from CARE are held at several museums as well.



CARE Figure 2. Geography of CARE (NPS).



CARE Figure 3. Stratigraphic column for CARE (from Morris et al. [2010], used with permission of the Utah Geological Association). The Mesaverde Formation is now identified as the Tarantula Mesa Sandstone of the Mesaverde Group.



CARE Figure 4. Schematic geological map of CARE. The Mesaverde Formation is now identified as the Tarantula Mesa Sandstone of the Mesaverde Group. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1049061.

Formation	Age	Fossils Within CARE	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Phytoliths and packrat middens	Alluvial, eolian, fluvial, landslide, and pediment deposits
Pliocene igneous rocks	Pliocene	Unfossiliferous (intrusive igneous rocks)	Not applicable (igneous rocks that solidified at depth)
Tarantula Mesa Sandstone (formerly identified as the Mesaverde Formation)	Late Cretaceous	None to date; root traces and invertebrate traces are most likely	Barrier beach and deltaic settings
Mancos Shale	Late Cretaceous	Plant fossils, bivalves, ammonites, gastropods, and shark teeth	Open marine, shallow marine, and coastal settings
Dakota Sandstone	Late Cretaceous	Bivalves, including a "reef" of *Exogyra* and *Pycnodonte*	Terrestrial (especially fluvial) becoming shallow marine over time
Cedar Mountain Formation	Early–Late Cretaceous	None to date; pollen, spores, plant fossils, and vertebrates are most likely	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	Petrified wood and logs, root traces, invertebrate burrows, and fragmentary bones	Fluvial, floodplain, and lacustrine settings
Summerville Formation	Late Jurassic	None to date; invertebrate and vertebrate trace fossils are most likely	Hypersaline shallow marine and tidal flat settings
Curtis Formation	Late Jurassic	None to date; marine invertebrates are most likely	Shallow marine
Entrada Sandstone	Middle Jurassic	None to date; fossils are very rare in this formation in general	Desert with large eolian sand dunes
Carmel Formation	Middle Jurassic	Bivalves, gastropods, and crinoids	Marine, shoreface, and coastal plain settings
Page Sandstone	Middle Jurassic	None to date; fossils are very rare in this formation in general	Eolian deposition on a coastal plain, with occasional marine incursions
Navajo Sandstone	Early Jurassic	Giant stromatolites	Desert with large eolian sand dunes
Kayenta Formation	Early Jurassic	Tracks of dinosaurs and mammal relatives	Primarily fluvial, with eolian episodes
Wingate Sandstone	Late Triassic–Early Jurassic	None to date; vertebrate tracks are most likely	Desert with large eolian sand dunes
Chinle Formation	Late Triassic	Petrified wood, other plant fossils, invertebrate burrows, bones, and vertebrate tracks and coprolites	Fluvial, floodplain, lacustrine, and eolian settings, becoming drier over time

Formation	Age	Fossils Within CARE	Depositional Environment
Moenkopi Formation	Early–Middle Triassic	Stromatolites, microbial debris, sponges, brachiopods, bivalves, ammonites, gastropods, scaphopods, ostracodes, crinoids, echinoids, burrows, horseshoe crab tracks, and reptile walking and swimming traces	Coastal plain with a shallow marine incursion
Kaibab Limestone	Early Permian	Sponges, bryozoans, brachiopods, bivalves, gastropods, crinoids, and conodonts	Open marine to marginal marine settings
Cutler Group undivided	Early Permian	Shells and invertebrate traces	Coastal eolian sand dunes with occasional marine incursions

CARE Table 1. Concise tabulation of CARE stratigraphy, fossils, and depositional setting; details and references can be found in the text. Igneous rocks are included here for completeness, but are not discussed in detail in the text because they are unfossiliferous.

Sources of Paleontological Resources

The fossils of CARE present opportunities for education, interpretation, and continued or future scientific research in the park.

Cutler Group (Lower Permian)

The Cutler Group of the Four Corners region is composed of several subordinate units (formations or members, depending on the source). At CARE, the Cutler Group includes the Cedar Mesa Sandstone and overlying White Rim Sandstone; at other locations, such as at Canyonlands National Park (CANY), a shale/mudstone unit known as the Organ Rock Shale is present between the two sandstone units, but it pinches out east of CARE. In the absence of the Organ Rock Shale, it is difficult to differentiate the two sandstone units, and they are typically mapped together as undivided Cutler Group rocks (Graham 2006; Morris et al. 2010). For many years, this stratigraphic interval was known as the Coconino or "Coconino" Sandstone at CARE, because it was thought that the rocks correlated to the Coconino Sandstone of the Grand Canyon region (Kirkland 1963; Kamola and Chan 1988).

At CARE, the Cutler Group is only exposed in the deep canyons in the western part of the park. It can be more than 370 m (1,220 ft) thick and is composed of yellow to gray, very-fine- to medium-grained sandstone with prominent cross-bedding reflecting eolian (wind-blown) deposition. The Cedar Mesa Sandstone is at least 240 m (800 ft) thick in the park, and the overlying White Rim Sandstone is 130 m (420 ft) thick; they are regarded as coastal dune units in the CARE area (Morris et al. 2010). The Cedar Mesa Sandstone was deposited during the Early Permian (Chamberlain and Baer 1973), sometime between 299 and 280 Ma. Multiple episodes of dune seas are separated by erosional surfaces thought to be from marine transgressions (Baars 2010). Sand arrived from the northwest (Chamberlain and Baer 1973). The depositional environment of the White Rim Sandstone at CARE has been the subject of detailed study (Huntoon and Kamola 1988; Kamola and Chan 1988). It is thought to be primarily a coastal dune formation at the park, with evidence of periodic marine flooding including reworked sandstone, thin limestone fingers pertaining to the overlying Kaibab Limestone, and marine trace fossils. A shallow epicontinental sea advanced from the west during deposition, as shown by the decreasing marine influence to the east: CANY outcrops of the White Rim Sandstone have only limited reworking at the very top of the formation (Huntoon and Kamola 1988; Kamola and Chan 1988).

The sandstone units of the Cutler Group are not known for their fossils. At CARE, the White Rim Sandstone has abraded marine shells and marine trace fossils in the reworked sandstone, including *Chondrites* and *Thalassinoides* (Huntoon and Kamola 1988). Graham (2006) reported marine trace fossils and shark teeth as fossil resources for the Cutler Group of CARE as a whole. Poorly preserved marine fossils are known from the Cutler Group of the nearby Circle Cliffs (Kirkland 1963, identified as "Coconino" Sandstone; Davidson 1967), including sponge spicules, bryozoans (moss animals), brachiopods (lamp shells), bivalves, gastropods, trilobites, and crinoids (sea lilies) from sandstone-dolomite beds previously assigned to the Kaibab Limestone (Davidson 1967).

Overall, the Cedar Mesa Sandstone's sparse fossil assemblage includes conifer logs (Sumida et al. 1999), root traces, sand-sized fragments of marine invertebrate fossils (Loope 1984), foraminifera (amoeba-like protists that form "shells"), crinoids (Baars 2010), rare casts of marine invertebrates (Kirkland 1963), ghost shrimp burrows and other invertebrate trace fossils (Chamberlain and Baer 1973), bones of unidentified tetrapods (Loope 1984), temnospondyl amphibians, and pelycosaurs (informally known as "mammal-like reptiles") (Sumida et al. 1999), and vertebrate tracks including synapsids (the group including pelycosaurs, other mammal relatives, and true mammals) and lizard-like animals (Lockley et al. 1998). Marine transgressive surfaces have yielded bryozoans, brachiopods, and crinoids (Langford et al. 2007).

The White Rim Sandstone has yielded microbial laminations (Chan 1989), possible impressions of something similar to a kelp-like brown algae (Baars and Seager 1970), root traces, tree trunk traces (Chan 1989), marine shells (Huntoon and Kamola 1988), a crinoid fragment (Steele 1987), echinoids (sea urchins) (Baars 1987), burrows (Huntoon and Chan 1987), and fecal pellets (Steele 1987), with marine fossils more common in the marine-reworked beds.

Kaibab Limestone (Lower Permian)

The Kaibab Limestone is a heterogeneous formation composed mostly of limestone and dolomite (Blakey and Knepp 1989). It is divided into members in Arizona, such as at Grand Canyon National Park (Beus and Billingsley 1989), but this is not done at CARE (Morris et al. 2010). At CARE, it is a cliff-forming unit (Morris et al. 2010) up to about 170 m (550 ft) thick (Martin et al. 2007) but averaging 60 m (200 ft) (Billingsley et al. 1987). This formation is sometimes called the Black Box Dolomite in the CARE area (see for example Hintze and Kowallis [2009]). It is made up of light gray to white chert-rich dolostone with thin beds of sandstone, siltstone, and fossiliferous limestone. Like the Cutler Group, it is exposed in the deep canyons near the western boundary (Morris et al. 2010). The Kaibab Limestone dates to the late Early Permian (Spamer 1984). In the CARE area, the lower contact with the White Rim Sandstone of the Cutler Group is interfingering (Kamola and Chan 1988), while the upper contact with the Moenkopi Formation is disconformable (Morris et al. 2010).

The Kaibab Limestone formed in a variety of settings, from shallow marine to salt flat to coastal plain (Blakey and Knepp 1989). It was deposited during a great marine transgression (Jenson 1986). The sea transgressed from west to east (Blakey and Knepp 1989). At CARE, the rock types indicate normal to marginal marine conditions (Morris et al. 2010), perhaps a shallow sea at or near the shoreline (Smith et al. 1963). Sea-level retreat during the Middle Permian exposed the Kaibab Limestone surface to subaerial erosion, and channels 30 m (100 ft) deep were formed (Morris et al. 2010).

The fossils of the Kaibab Limestone at CARE include sponge spicules (Huntoon and Kamola 1988), bryozoans, brachiopods, bivalves, gastropods, crinoids (Smith et al. 1963; Morris et al. 2010), and conodonts (extinct eel-like chordates) (Huntoon and Kamola 1988). Disarticulated and fragmentary remains of bryozoans, brachiopods, bivalves, gastropods, and crinoids are found in the thin limestone layers. Few are identifiable to genus or species, although the brachiopods *Dictyoclostus bassi* (found in the lower part of the formation) and *Neospirifer pseudocameratus* (usually found in the upper 15 m [50 ft]) have been identified (Smith et al. 1963; Morris et al. 2010). Gregory and Anderson (1939) reported abundant fossils in possible Kaibab Limestone rocks of Capitol Wash and Grand Wash, including bivalves. Limestone interfingerings of the Kaibab Limestone into the underlying White Rim Sandstone at CARE include brachiopods, gastropods, and crinoids (Huntoon and Kamola 1988). In the adjoining Circle Cliffs, Davidson (1967) reported sponges, brachiopods, bivalves, gastropods, and crinoids, with coquina (shell hash) beds up to 15 cm (6 in) thick.

Considering the Kaibab Limestone over its entire area of deposition, the taxonomic list is extensive, including trace fossils, algae (Mather 1970), foraminifera, sponges, corals, conulatans (extinct distant relatives of coral that formed conical structures from rods), bryozoans, brachiopods, bivalves, cephalopods (octopi, squid, and mostly extinct shelled relatives), gastropods, scaphopods (tusk shells), ostracodes (seed shrimp), trilobites, crinoids, echinoids, and conodonts (Kirkland 1963; Hopkins 1990; Spamer 1992), and a few reports of chondrichthyan (cartilaginous fish) teeth, spines, and dermal denticles, and ray-finned fish teeth and tooth plates (Hunt et al. 2005). There are distinct mollusk and open marine faunas (Hopkins 1990). Kaibab Limestone fossils may also be found in rock fragments incorporated into the Chinle Formation (McKee 1937).

Moenkopi Formation (Lower–Middle Triassic)

The Moenkopi Formation is a heterogeneous unit present in several western states. At CARE, it is mostly composed of red and brown sandstone, siltstone, and shale, with the exception of one member (the Sinbad Limestone) which is mostly made of yellow limestone and dolostone (Morris et al. 2010). It reaches a maximum thickness of approximately 300 m (1,000 ft) (Billingsley et al. 1987; Doelling and Kuehne 2007). The lower contact with the Cutler Group and the upper contact with the Chinle Formation are disconformities representing millions of years of time (Morris et al. 2010). The Moenkopi Formation dates to the Early and early Middle Triassic (Heckert et al. 2005a), between approximately 250 and 240 Ma. It has been divided into several members, depending on location. At CARE, four members are present. From oldest to youngest, these are the Black Dragon, Sinbad Limestone, Torrey, and Moody Canyon members (Morris et al. 2010). Aside from the Sinbad Limestone Member, these names were coined by Blakey (1974) and correspond to older stratigraphy as outlined by Stewart et al. (1972a) as follows: the Black Dragon Member is the lower slope-forming member, the Torrey Member is the ledge-forming member, and the Moody Canyon Member is the upper slope-forming member and overlying cliff-forming member. The Moenkopi Formation of central and southeastern Utah is essentially the remains of a large coastal plain mudflat, influenced by and sometimes covered by the neighboring sea. The Black Dragon Member and Sinbad Limestone are interpreted as coastal and marine units, the Torrey Member as a large delta, and the Moody Canyon Member is interpreted as a restricted shallow water to sabkha (salt flat) setting (Blakey 1974).

The Black Dragon Member is a heterogeneous unit of mudstone to sandstone, limestone, dolomite, and gypsum (Blakey 1974). At CARE, it is a reddish slope-forming unit of conglomerate, siltstone, and sandstone, between 15 and 35 m (50 and 110 ft)

thick. The base of the member is a conglomerate with chert fragments from the Kaibab Limestone. Thin beds of limestone and dolostone are common in the upper part of the member, and may have marine fossils. The Black Dragon Member of CARE is interpreted as representing coastal plain and tidal flat deposition (Morris et al. 2010). The upper contact with the Sinbad Limestone is transitional (Stewart et al. 1972a) to intertonguing (Blakey 1974). Fossils are not currently well-known in the Black Dragon Member. Some body fossils are known, such as bivalves, gastropods, a possible ammonite, ostracodes, and a conodont fragment. These body fossils are best known from the San Rafael Swell (Blakey 1974). Trace fossils are known as well (Blakey 1974). As mentioned, marine fossils may be found in the thin carbonate beds at CARE, but they are rare (Morris et al. 2010).

The Sinbad Limestone Member was recognized before the other members, as it is distinctive compared to the surrounding red beds of sandstone, siltstone, mudstone, and claystone (McKee 1954). At CARE, it is 20 to 45 m (70 to 140 ft) thick and is composed of yellow mud-rich limestone and dolostone, with some siltstone and sandstone (Morris et al. 2010). It represents the easternmost marine carbonates of the Early Triassic of the Colorado Plateau. Depositional environments included offshore bars, shallow shelves, and lagoons (Blakey 1974). At CARE, recognized depositional environments include tidal flats, subtidal marine settings, and shoals (Dean 1981). Because it is a marine unit, Lucas et al. (2007a) proposed removing it from the Moenkopi Formation as part of a reorganization to raise the formation to a group and limit it to terrestrial rocks. It grades into the overlying Torrey Member (Stewart et al. 1972a).

The Sinbad Limestone is known for marine invertebrate fossils, and is fossiliferous at CARE and the immediate vicinity. Gregory and Anderson (1939) reported finding bivalves, gastropods, possible ostracodes, and crinoids. Smith et al. (1963) reported bivalves, ammonite cephalopods, gastropods, and echinoids from CARE or the immediate vicinity. Dean (1981), in the course of a study on the depositional settings and petrology of the Sinbad Limestone, took measured sections at nine locations within CARE and three more just outside of the park, and found a fossil assemblage of stromatolites, limy microbial debris, sponges, brachiopods, bivalves, ammonites, gastropods, possible scaphopods, ostracodes, crinoids, echinoids, and burrows. Finally, microbial debris, bivalves, and gastropods have been found in this member in the Circle Cliffs (Davidson 1967). Fossils of the Sinbad Limestone in general include stromatolites, algae, foraminifera, sponges, brachiopods, bivalves, ammonites, gastropods, scaphopods, ostracodes, crinoids, echinoids, conodonts, possible fish material, and marine trace fossils (McKee 1954; Stewart et al. 1972a; Blakey 1974; Dean 1981; Goodspeed and Lucas 2007). Newly-recognized in the San Rafael Swell area are brittle stars and serpulid worm tubes (Hautmann and Nützel 2005).

The Torrey Member is mostly composed of sandstone and siltstone, but many other rock types are present in minor quantities. Often the base is siltstone, followed by ledgey sandstone in the middle, and then an upper sandstone/siltstone section (Blakey 1974). It is 75 to 105 m (250 to 350 ft) thick at CARE, and is composed of red and brown siltstone and sandstone, with minor amounts of conglomerate, dolostone, limestone, and claystone. The member is interpreted as representing tidal flat and coastal plain depositional settings at the park (Morris et al. 2010). It grades into the overlying Moody Canyon Member (Stewart et al. 1972a).

Paleontologically, the Torrey Member is best known for its trace fossils. Numerous tracks and other trace fossils are known from the Torrey Member of CARE, and have been described in several publications, notably Peabody (1956), Lammers (1964), Kirby and McAllister (1998), and Mickelson et al. (2006). Tracks were first found in Moenkopi Formation rocks of the CARE area by CARE custodian Charles Kelly in 1945, which brought paleontologist Charles Camp and Zion National Park (ZION) naturalist Myrl Walker to CARE for further investigations in 1950 (Santucci and Kirkland 2010). Swim tracks and other subaqueous tetrapod traces from the Torrey Member of CARE are some of the first such traces recognized in the world (Kirby and McAllister 1998). Peabody (1956) reported on two localities within CARE. Tracks named *Rotodactylus* and interpreted as being produced by a semi-bipedal tridactyl (three-toed) pseudosuchian (a type of reptile now known to be part of the lineage including modern crocodiles) were found by Charles Kelly near the western boundary of the park, 6.0 km (3.7 miles) west of the headquarters north of the Fruita–Torrey road. These were in Kelly's possession at the time, along with limuloid (horseshoe crab) tracks and other vertebrate tracks, and vertebrate swim marks. Peabody also mentioned the presence of *in situ* swim traces at other locations. The second location described by Peabody is UCMP (University of California Museum of Paleontology, Berkeley) Locality V5366, 145 m (480 ft) up Sulphur Creek from old lime kilns above a ranger station. At this site, tracks (probably the common reptile track *Chirotherium*) were collected for the museum, and were reposited as UCMP 43751 (Peabody 1956).

Lammers (1964) reported that vertebrate tracks are found throughout the Moenkopi Formation of CARE, mostly concentrated at the transition between the units now recognized as the Torrey Member and overlying Moody Canyon Member. These tracks included *Chirotherium*, *Rotodactylus*, lacertoid tracks (from a lizard-like animal), and swimming *Chirotherium* traces. Kirby and McAllister (1998) reported on subaqueous tetrapod traces from three locations in the park: a low ridge west of the headquarters and visitor center bounded on the east and south by Sulphur Creek; a broad slope 0.40 km (0.25 miles) west of the first site, on Route 24; and the base of a small cliff 4.0 km (2.5 miles) west of the second site, south of Route 24 near a dry stream bed. Tracks were plentiful at all three sites.

Mickelson et al. (2006) reported on trace fossils from multiple stratigraphic levels representing delta plains in the Torrey Member of CARE, GLCA, and the San Rafael Swell. Types of fossils specifically reported from CARE include invertebrate burrows (CARE Figure 5) and the reptile ichnogenera *Chirotherium*, *Rhynchosauroides*, and *Rotodactylus* (CARE Figure 6). Other fossils from unspecified locations include horsetails (*Equisetum*), *Arenicolites* (annelid worm trace), a millipede body impression, *Fuersichnus* (possible crustacean or polychaete worm trace), *Kouphichnium* (horseshoe crab trace), *Palaeophycus* (invertebrate burrow), fish bones, *Undichna* (fish fin marks), salamander-like tracks, and rare tetrapod swim marks. Tracks have been found as sandstone casts in mudstone. To date, reptile tracks are dominant (Mickelson et al. 2006). Additionally, a brachiopod was recovered from this unit at or near CARE, and *Chirotherium* and tridactyl reptile tracks (*Rotodactylus*) are found in many places (Smith et al. 1963). Scales of the fish *Boreosomus* were recovered from the Circle Cliffs (Davidson 1967).





CARE Figure 5. Invertebrate trace fossils from the Torrey Member (VLS).
CARE Figure 6. Vertebrate trace fossils from the Torrey Member (VLS).

Fossils from the Torrey Member in general include *Equisetum*, brachiopods, bivalves, ostracodes, fish scales (Blakey 1974), fish bones (Mickelson et al. 2006), a skull of the labyrinthodont amphibian *Parotosuchus* from Hite Crossing (Lucas and Schoch 2002), invertebrate traces, fish swim traces, and amphibian and reptile traces (Mickelson et al. 2006).

The Moody Canyon Member is a heterogeneous unit with red to brown siltstones and mudstones, sandstones, and an upper section of dolostone and gypsum. It is notable for its continuous regular beds (Blakey 1974). It can be best distinguished from the Torrey Member by their differing weathering profiles: the Torrey Member weathers as ledges, while the lower Moody Canyon Member beds weather as slopes and the beds of the upper section weather as cliffs. At CARE, the Moody Canyon Member is 100 to 130 m (320 to 430 ft) thick, 60 to 90 m (200

and 300 ft) of which are made up by the slope-forming section. The Moody Canyon Member of CARE is another tidal flat and coastal plain unit (Morris et al. 2010). The lower slope-forming section may represent a marine transgression (Lammers 1964). Fossils are rare in this unit, with only *Chirotherium* tracks from an outcrop in the Orange Cliffs (Blakey 1974) reported to date.

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location (Stewart et al. 1972b). In CARE, these include from oldest to youngest the Shinarump, Monitor Butte, Petrified Forest, and Owl Rock members (Martin et al. 2007; McLelland et al. 2007; Sorber et al. 2007; Morris et al. 2010). Other members are sometimes reported, particularly the Church Rock Member, which would be the uppermost member if present (Lupe 1984; Billingsley et al. 1987; Doelling and Willis 1999). They can be divided into bentonitic (clay formed from volcanic ash) and red bed parts of the Chinle Formation, with the Petrified Forest Member at the top of the bentonitic portion and the Owl Rock Member as the base of the red bed portion (Stewart et al. 1972b). The formation as a whole thickens from approximately 150 m (500 ft) in the east to 210 m (700 ft) in the west in the park (Billingsley et al. 1987). The lower contact with the Moenkopi Formation and the upper contact with the Wingate Sandstone are disconformities at CARE (Morris et al. 2010). The Chinle Formation is particularly known for petrified wood and other plant fossils, and such fossils have been reported from CARE and the vicinity (Gregory and Anderson; Smith et al. 1963; Ash 1975, 1982, 1993; Lupe 1984; Mathis 2004; Santucci and Kirkland 2010). The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) because this division is known to be in the overlying Wingate Sandstone (Lucas et al. 1997a, 2006a).

The Shinarump Conglomerate Member is one of the most widely exposed units of the Chinle Formation, and is present in Utah, Nevada, Arizona, and New Mexico (Colbert 1972; Ash 1974). It was deposited during the early Late Triassic (Lucas et al. 1997b). At CARE, it is a discontinuous unit up to 30 m (90 ft) thick, composed of white to yellowish-gray conglomerate and conglomeratic sandstone. It fills paleochannels in the underlying Moenkopi Formation, and is interpreted as having been deposited by braided streams (Morris et al. 2010). Shinarump Conglomerate Member braided streams were part of a drainage system flowing north into the center of the depositional basin (Parrish 1989). The climate is interpreted as monsoonal (Demko et al. 2005).

The most common fossils of the Shinarump Conglomerate Member are plants (Colbert and Gregory 1957). Coal, plant fragments (Smith et al. 1963), leaves, twigs, and petrified wood and logs (Gregory and Anderson 1939) are known from this unit at CARE and its surroundings; plant fragments are abundant (Smith et al. 1963). Localities in CARE where plant fossils have been found include Capitol Gorge, Fremont Canyon, Grand Gulch, a tributary canyon of Sulphur Creek approximately 3 km (2 miles) northwest of Fruita (Gregory and Anderson 1939), and Oyler Mine (Mathis 2004). A site just west of Fruita has yielded bennettitales (cycadeoids) and true cycads (Smith et al. 1963). Several plant genera have been identified from the Shinarump Conglomerate of CARE, including *Araucarioxylon* (the namesake petrified wood of Petrified Forest), *Cladophlebis*, *Cyneopteris*, *Equisetites*, *Palissya*, *Pagiophyllum*, *Phleboterpis*, *Sphenozamites*, and *Zamites*. Most of these have been recovered from the northern part of the park (Santucci and Kirkland 2010). Plant fragments, wood fragments, and pollen are also known from the Shinarump Conglomerate Member of the Circle Cliffs (Davidson 1967).

Types of plants found in the Shinarump Conglomerate across its depositional area include horsetails, ferns and fern-like plants, bennettitales, cycads, and conifers and other gymnosperms (Ash 2005a). Some specimens show traces of arthropod attacks (Ash 1996). Other fossils are also known from this unit. Bone fragments are commonly preserved as cobbles, usually too eroded to identify. Some can be identified as belonging to metoposaur amphibians or phytosaurs (crocodile-like reptiles) (Colbert and Gregory 1957; Parrish 1989). Tetrapod footprints, swim traces (Lockley and Milner 2006), and coprolites (fossil feces) are also known (Heckert et al. 2002). For invertebrates, tentaculites (enigmatic possible mollusks, misidentified as shells of the polychaete worm *Spirorbis*), two types of insect eggs, and possible beetle feeding traces have been found (Ash 2005a). Reworked marine fossils from older formations may also be found, such as foraminifera (McKee 1937; Miller and Breed 1962).

The Monitor Butte Member is interpreted as valley fill deposited under a monsoonal climate (Demko et al. 2005). It is a mudstone, siltstone, and sandstone unit (Parrish 1989), with deposition in a system of braided and meandering streams, lakes, and swamps, with volcanic input (Blakey and Gubitosa 1983). At CARE, it is typically purplish-gray and is made mostly of claystones and clay-rich sandstone (Morris et al. 2010). As with the Shinarump Conglomerate, fossil plants are known from the Monitor Butte Member at CARE. The bennettitale *Zamites powelli* was collected from the Waterpocket Fold area in what is now the park by R. C. Moore in 1921. His locality, USGS Locality 10142, is probably in the Monitor Butte Member, 3 km (2 miles) west of the fold and 5.6 km (3.5 miles) northeast of Wagon Box Mesa. Specimens of *Zamites powelli* from this locality are reposited in the National Museum of Natural History (Washington, D.C.) as USNM 168987 through 168990 (Ash 1975). In the nearby Circle Cliffs, phytosaur teeth are known from this unit (Davidson 1967).

The Petrified Forest Member was deposited by streams in a similar way to the Monitor Butte Member, as a

sandstone/mudstone fluvial unit with volcanic input (Blakey and Gubitosa 1983, 1984). It is divisible into smaller units at locations like Petrified Forest National Park (PEFO). Chinle Formation fossil vertebrates are most diverse in the Petrified Forest Member's various units (Parrish 1989). At CARE, it is composed of reddish-orange siltstones and fine-grained sandstones. The upper part of the member is capped by a sandstone cliff or ledge known as the Capitol Reef Bed (Morris et al. 2010). Petrified logs were reported from this member in the Circle Cliffs by Davidson (1967).

The Owl Rock Member is a carbonate and mudstone unit (Heckert and Lucas 2002). It was traditionally interpreted as largely lacustrine because of its carbonate beds (Blakey and Gubitosa 1983; Dubiel 1989; Parrish 1989). This was revised in the 1990s, with the carbonate units instead thought to represent ancient soil horizons from a semi-arid setting (Lucas et al. 1997a; Tanner 2000). Further study concluded that the Owl Rock Member was a combination of both soil horizons and small ponds and lakes, on a fluvial floodplain (Lucas and Heckert 2003; Fraser et al. 2005). The Owl Rock Member represents the beginning of aridification in the Chinle Formation as Pangaea drifted north (Heckert and Lucas 2002; Tanner 2003). The Owl Rock Member of CARE is 45 to 60 m (150 to 200 ft) thick and mostly composed of orange and purple mudstones, siltstones, and sandstone. There are also beds of mottled green to pink limestone up to 3 m (10 ft) thick (Morris et al. 2010). Ash (1982) collected specimens of the unusual plant *Sanmiguelia* from an Owl Rock Member locality approximately 9 km (6 miles) south of CARE headquarters, in the southeast quarter of Section 7, T19S, R7E. Two slabs of plant fossils from this site are in USNM collections, as USNM 298826 and USNM 298827 (Ash 1982).

In addition to the member-specific citations mentioned above, fossils are known from locations in the Chinle Formation at CARE and its immediate vicinity where a member has not been specified (CARE Figure 7). Gregory and Anderson (1939) reported the presence of leaves, twigs, petrified wood, logs, shell fragments, and bone fragments. Smith et al. (1963) reported plant fragments, wood, and logs, with bivalve shells in the middle part of the formation and tooth and bone fragments in the upper part of the formation. Lupe (1984) found burrows and petrified wood throughout the formation at several locations in the park. Ash (1993) reported that the plant assemblage of the park includes horsetails, ferns, cycads and relatives, conifers, and *Sanmiguelia*. Santucci and Kirkland (2010) reported the presence of leaf impressions, petrified wood, bivalves, bones, and trace fossils including tridactyl tracks and coprolites. Finally, there is petrified wood on the Chimney Rock Trail (Mathis 2004).



CARE Figure 7. Petrified log fragments from the Chinle Formation (VLS).

The various members of the Chinle Formation above the Shinarump Conglomerate Member have similar types of fossils, but differ in the genera and species present and which are most commonly found. Plants are represented by charophyte algae (Lucas and Keitzke 1993), diverse palynomorphs (Gottesfeld 1972; Fisher and Dunay 1984; Parker 2004), and leaves, wood, and other remains of lycopods, horsetails, ferns and fern-like plants, cycads, bennettitales, ginkgoes, conifers, and taxa of uncertain classifications (Ash 1972, 1989, 2003, 2005b; Stewart et al. 1972b; Parker and Ash 2004). Fungal remains are also present (Ash 1972; Stewart et al. 1972b). Invertebrates are represented by bivalves, gastropods, ostracodes, conchostracans (clam shrimp) (Breed 1972; Stewart et al. 1972b), crabs (Rinehart et al. 2003), crayfish (Miller and Ash 1988), insects (Breed 1972; Stewart et al. 1972b), and various trace fossils (Caster 1944; Breed 1972; Gillette et al. 2003; Demko et al. 2005), including borings in wood (Walker 1938; Tapanila and Roberts 2012) and a variety of feeding traces on plants (Ash 1996).

The Chinle Formation's vertebrate assemblage has been summarized in a number of documents, including Stewart et al. (1972b), Colbert (1972), Parrish (1989), Long and Murry (1995), Heckert et al. (2005a, 2005b), Irmis (2005a), Parker (2005, 2009), Lucas et al. (2007b), Loughney et al. (2011), and Parker and Martz (2011). Vertebrates include: freshwater sharks; ray-finned fish; lungfish; coelacanths; several types of enigmatic tetrapods known from isolated bones and teeth; metoposaur amphibians; procolophonids (stocky lizard-like herbivorous and insectivorous reptiles); sphenodontians (relatives of the modern tuatara); trilophosaurids (reptiles that resembled large herbivorous lizards); tanystropheids (long-necked piscivorous reptiles); the unusual aquatic reptile *Vancleavea*; drepanosaurids (beaked, possibly aquatic chameleon-like reptiles); several forms of crocodile-like phytosaurs, known under a variety of names; several forms of herbivorous armored aetosaurs, also known under a variety of

names; the small unusual armored *Revueltosaurus*; "rauisuchian" crocodile relatives including poposaurids and rauisuchids, which resembled large carnivorous dinosaurs, and shuvosaurids (toothless forms that resembled ostrich-mimic dinosaurs); "sphenosuchians" (small terrestrial crocodile relatives); lagerpetonids (small bipedal dinosaur relatives); silesaurids (quadrupedal dinosaur relatives); several early dinosaur taxa; and dicynodonts (a type of non-mammalian therapsid, or "mammal-like reptile"). Lungfish, metoposaurs, aetosaurs, and phytosaurs are the most common forms. Vertebrate trace fossils include coprolites (Hunt et al. 1998), swim traces, and footprints possibly attributable to sphenodontians, tanystropheids, phytosaurs, aetosaurs, carnivorous and herbivorous dinosaurs, and mammal relatives (Hunt and Lucas 2007). Bones accumulated in floodplains, bogs, small ponds, and in or near channels (Parrish 1989).

Glen Canyon Group: Kayenta Formation (Lower Jurassic)

The Kayenta Formation is mostly made up of fine- to coarse-grained sandstone with some small amounts of interbedded shale and siltstone, and rare limestone and conglomerate (Vandiver 1936). At CARE it is up to 105 m (350 ft) thick, and is composed of three units: a lower ledge-forming unit of very fine sandstone; a middle cliff-forming unit of very fine sandstone; and an upper slope-forming unit of very fine sandstone and siltstone. Beds of conglomerate, mudstone, and limestone are found throughout. It is interpreted as a fluvial unit at the park, grading out of one eolian formation (the underlying Wingate Sandstone, described under "Potential Sources of Paleontological Resources") and into another (the overlying Navajo Sandstone) (Morris et al. 2010). The middle cliff-forming unit may be an eolian tongue as well (Friz 1980). The transition zone with the Navajo Sandstone at CARE includes a thin tongue of Kayenta Formation rocks in the lower Navajo Sandstone (Doelling and Kuehne 2007). The transition zone with the Navajo Sandstone represents a time when the Navajo Sandstone dunes advanced over the rivers of the Kayenta Formation (Harshbarger et al. 1957). While dating of Glen Canyon Group units has proven difficult to establish, the age of the Kayenta Formation is now thought to be middle Early Jurassic, between approximately 197 and 190 Ma (Lucas et al. 2005). The Kayenta Formation was formed by shifting freshwater braided and meandering streams (Colbert 1974); floodplain deposits are also known (Lucas et al. 2005).

The Kayenta Formation is known for its dinosaur tracks. Examples of these can be found at CARE (Santucci et al. 1998; Santucci and Kirkland 2010), CANY (Santucci et al. 1998; Santucci and Kirkland 2010), and GLCA (Lockley et al. 1998; Santucci and Kirkland 2010), extending through the formation proper into the Kayenta Formation–Navajo Sandstone transition zone. The CARE tracks have not yet been described. Tracks of tritylodontid synapsids (early mammal relatives) are also reported from CARE (Santucci et al. 1998). Tridactyl dinosaur tracks have also been found in the Circle Cliffs, as well as poorly preserved bivalves (Davidson 1967).

Vertebrate body fossils of the Kayenta Formation have been described in many publications, but most are from a silty facies found to the south in Arizona (see Lucas et al. [2005] or the Navajo National Monument summary of the Southern Colorado Plateau Network report for a partial list). Vertebrate body fossils of the sandy facies of the Kayenta Formation, which would be typical of CARE, include fish scales, crocodylomorphs (crocodiles and their relatives), and tritylodontid synapsids such as *Kayentatherium* (Clark and Fastovsky 1986). Kayenta Formation vertebrate trace fossils include coprolites (Luttrell and Morales 1993), therapsid tracks, lizard-like tracks (Hamblin and Foster 2000), and dinosaur tracks such as *Grallator* (small theropod tracks) and *Eubrontes* (large theropod tracks) (Lucas and Tanner 2007). The relevant ichnofossil taxonomy can be complicated. A recent summary for the Chinle Formation and Glen Canyon Group can be found in Lucas et al. (2006b). Kayenta Formation fossils that are not vertebrates or vertebrate traces include microbial limestone (Luttrell and Morales 1993), petrified wood (Jenkins and Shubin 1998), plant impressions (Rigby et al. 1971), freshwater bivalves and gastropods (Harshbarger et al. 1957), ostracodes (Lucas and Tanner 2007), and invertebrate trails and burrows (Luttrell and Morales 1993).

Glen Canyon Group: Navajo Sandstone (Lower Jurassic)

The Navajo Sandstone is a fine- to medium-grained quartz sandstone noted for its prominent cross-bedding formed by dune progression. There are also lenticular beds of cherty limestone (Harshbarger et al. 1957). At CARE, it is between 290 and 425 m (950 to 1,400 ft) thick and is composed of white, tan, and reddish-brown sandstone, grading out of the Kayenta Formation. The upper contact with the Page Sandstone is a disconformity. The Navajo Sandstone forms cliffs, monoliths, and domes in the park, including the domes that help give the park its name (Morris et al. 2010). The bulk of the Navajo Sandstone probably dates to the middle and late Early Jurassic (Lucas et al. 2005), between approximately 190 and 175 Ma.

The Navajo Sandstone represents the largest erg known from North America (Blakey 2008). As discussed in the "Kayenta Formation" section, the upper Kayenta Formation intertongues with the lower Navajo Sandstone as the dunes of the latter overrode the streams of the former. Recent research on geochronology of detrital zircon grains in the Navajo Sandstone suggests that most of the sand was eroded from the ancestral Appalachian Mountains, transported westward by a continental-scale river system to the western shore of Jurassic North America, and was then blown southward and incorporated into the Navajo–Nugget–Aztec dune field (Dickinson and Gehrels 2003, 2009; Rahl et al. 2003). Preservation of this tremendous thickness of wind-blown sand was made possible because of basin subsidence associated with Early Jurassic compressional deformation near the west margin of North America. This

deformation caused the continental interior to flex downward, creating accommodation space for sediment accumulation (Allen et al. 2000).The Navajo Sandstone formed in a tropical arid setting with eolian, playa, and sabkha deposits (Mahaney et al. 2004). Wetter restricted interdunes are also known, and are characterized by mudstone with planar bedding, dolomitic limestone, and sandstone (Irmis 2005b). Interdunes are best known from the vicinity of Moab. At locations such as this, interdunes ranged from fluvial sandstones to lake limestones. During periods of increased precipitation, vegetation spread (Wilkens 2008). Fossils are mostly associated with interdunes (Loope et al. 2004b). As time went on during the deposition of the Navajo Sandstone, the climate became drier (Irmis 2005b), although there were several periods of increased precipitation with durations on the order of tens of thousands of years (Milan et al. 2008).

The Navajo Sandstone is not extensively fossiliferous. To date, fossil organisms and traces represented include stromatolites, root traces (Irmis 2005b), pollen, spores, horsetails, ferns (Loope et al. 2004b), cycadeoids, conifers (Wilkens 2008), petrified wood (Irmis 2005b), possible fungal filaments (Mahaney et al. 2004), worm burrows (Stokes 1978), insect burrows, scorpion tracks (Loope et al. 2004b), unionid bivalves (Wilkens 2008), ostracodes, conchostracans (Harshbarger et al. 1957), a crocodylomorph, the small bipedal carnivorous theropod dinosaur *Segisaurus* (Irmis 2005b), at least two kinds of sauropodomorph dinosaurs ("prosauropods") including *Seitaad* (Sertich and Loewen 2010), a tritylodontid synapsid, and vertebrate trace fossils (Irmis 2005b). The assemblage may be a persistent remnant of the Kayenta Formation assemblage (Winkler et al. 1991). Sandstone dune deposits have mostly yielded vertebrates, while limestone interdune beds have mostly yielded plants and invertebrates (Stokes 1991), although arthropod tracks (*Octopodichnus*) are known from the dunes (Hunt and Lucas 2006). Interdune fossils include stromatolites, root casts, horsetails, ostracodes, and insect traces, with locally common petrified wood (Irmis 2005b).

Vertebrate trace fossils are much more common than body fossils in the Navajo Sandstone, and have been extensively described. More than 60 dinosaur track sites are known in the Navajo Sandstone across its depositional area (Loope 2006; Milan et al. 2008). The various vertebrate trace fossils have gone under different names and interpretations, depending on the author. Recent works (Irmis 2005b; Hunt and Lucas 2006; Loope 2006) seem to have settled on seven or eight primary ichnogenera: *Brasilichnium* (synapsid, but see Milan et al. 2008 for an interpretation as a sauropodomorph), *Batrachopus* (crocodylomorph), *Grallator* (small theropod), *Anchisauripus* (medium theropod), *Eubrontes* (large theropod), *Navahopus* (small ?sauropodomorph), *Otozoum* (larger sauropodomorph), and *Anomoepus* (ornithischian or "bird-hipped" dinosaur) (Irmis 2005b). A few oddities are also known, including lizard-like tracks (*Lacertipus*), a track once thought to be pterosaurian in origin ("*Pteraichnus*"), bird-like *Trisauropodiscus* tracks (Irmis 2005b), undescribed large ornithopod (beaked herbivorous dinosaur) tracks (Santucci and Kirkland 2010), and large vertebrate burrows thought to have been made by mammals and/or mammal relatives (Lucas et al. 2006c; Riese et al. 2011). So far, large traces (*Anchisauripus*, *Eubrontes*, and *Otozoum*) are only known from interdunes, with smaller traces (*Batrachopus*, *Brasilichnium*, and *Grallator*) the only tracks found in dune sands (Loope 2006), although this differentiation may be due to preservational bias (Irmis 2005b). In general, the vertebrate trace fossils are consistent with the known body fossils.

There are a few fossils reported in the Navajo Sandstone at CARE. Giant stromatolites 5 m (15 ft) tall have been described from South Cottonwood Wash and Fivemile Wash. They grew on south-facing interdune margins, and may have formed at the same time, during a period with a relatively high water table (Eisenberg 2003). These mounds are associated with carbonate lenses that may represent oases (Santucci and Kirkland 2010).

San Rafael Group: Carmel Formation (Middle Jurassic)

The Carmel Formation is one of the units responsible for the unique scenery of CARE, with remnants of its heterogeneous strata eroded into unusual forms along the crest of Waterpocket Fold (Morris et al. 2010). For example, Glass Mountain is composed of selenite crystals in this formation (Ege 2005). At the park, it is between 60 and 305 m (200 and 1,000 ft) thick, and is composed of sandstone, siltstone, mudstone, gypsum, and fossiliferous limestone. The clastic rocks (made of rock fragments) are reddish in color, while the limestone and gypsum are white, producing distinctive banding. The heterogeneous rock types, and its fossils, indicate that the Carmel Formation is a coastal unit that was sometimes deposited under water and sometimes deposited under open air (Morris et al. 2010). At CARE, the Carmel Formation includes several members. Recent maps have found it to include one or two basal members separated from the rest of the formation by the intervening Thousand Pockets Member of the Page Sandstone (described below under "Potential Sources of Paleontological Resources") (Doelling and Willis 1999; Doelling 2004; Doelling and Kuehne 2007; Martin et al. 2007; McLelland et al. 2007; Sorber et al. 2007). These maps agree that the upper part of the formation includes the Paria River Member and overlying Winsor Member, but differ on the terminology of the lower part of the formation. Doelling and Willis (1999), Martin et al. (2007), McLelland et al. (2007), and Sorber et al. (2007) describe the lower part as the Judd Hollow Member, while Doelling (2004) and Doelling and Kuehne (2007) divide it into the Co-op Creek and Crystal Creek members.

The Carmel Formation is known for its stratigraphic relationship with the Page Sandstone. The lower Carmel Formation complexly intertongues with the Page Sandstone (Jones and Blakey 1993; Blakey et al. 1996; Doelling and Willis 2006). Where it intertongues with the Page Sandstone, it is a marine

unit, but it becomes terrestrial above the Page Sandstone (Blakey et al. 1983). Clastic rocks such as siltstone become more abundant higher in the unit (Wright and Dickey 1963), reflecting the change from marine to shoreface to coastal plain settings (Blakey et al. 1983). When marine, the water depth was 100 m (330 ft) deep at most (Tang et al. 2000).

The Carmel–Page sequence includes two transgressive-regressive cycles. The various tongues of the Page Sandstone represent eolian and beach sand and sandy, muddy, or limy marginal marine settings, deposited against the Carmel–Twin Creek Seaway represented by the Carmel Formation. This seaway ran through central Utah during the Middle Jurassic. The lower Carmel Formation is more prominent to the west, and the Page Sandstone is more prominent to the east (Blakey et al. 1983). Ash from distant volcanoes is also recorded in these rocks, becoming most prominent at the top of the Page Sandstone and the upper Carmel Formation. This corresponds to volcanic activity on the southern margin of the Colorado Plateau that occurred around 170 Ma (Blakey and Parnell 1995). Ephemeral streams in the Page Sandstone remobilized the volcanic debris (Jones and Blakey 1997). The volcanism and fossil content help to date the formation. The Co-op Creek Limestone Member is regarded as dating from about 169 to 167 Ma, the Crystal Creek and Paria River members date from about 166 to 165 Ma, and the Winsor Member dates from about 164 to 162 Ma (Sprinkel et al. 2011).

Carmel Formation limestone beds at CARE and the vicinity have abundant fossils, described primarily by Imlay (1964) and Sohl (1965). Imlay reported bivalves from three CARE locations. USGS Locality 10436 is just north of Capitol Wash in the northwest quarter of Section 10, T30S, R7E. Specimens were found there by E. Davis in 1920, and include the bivalves *Camptonectes*, *Isognomon*, and *Ostrea*. USGS Locality 18076 is 6 km (4 miles) below Fruita, in Section 17, T29S, R7E. This site was prospected by H. E. Gregory in 1937 and yielded the bivalve *Pronoella*. Finally, USGS Locality 25685 is on the north side of the mouth of Capitol Wash, in the northwest quarter of Section 10, T30S, R7E. It was found by Imlay in 1955, and yielded the bivalves *Ostrea* and *Pronoella* (Imlay 1964). Sohl (1965) reported that *Neritina* and other neritid gastropods are also present at USGS Locality 10436. Other reports from CARE and the area include bivalves and crinoids from Capitol Gorge within CARE and the lower Fremont Canyon near Notom outside of CARE (Gregory and Anderson 1939), and bivalves, gastropods, and crinoids from sites near Teasdale (16 km [10 miles] west of CARE) and Bicknell (17 km [11 miles] west of CARE) (Smith et al. 1963).

Although only marine fossils are known from the Carmel Formation of CARE to date, terrestrial fossils are known elsewhere. Marine fossils are most common in the lower part of the formation, while terrestrial trace fossils are known from the upper part of the formation, particularly from northeastern Utah. The general Carmel Formation marine assemblage include dinoflagellates (a type of plankton that move using whip-like flagella) (Sprinkel et al. 2011), algae, foraminifera, radiolarians, sponges, corals, bryozoans, brachiopods, bivalves, ammonites, gastropods, ostracodes, crinoids, echinoids, sea stars, and trace fossils including pellets and worm tubes (Imlay 1964; Blakey et al. 1983; Nielson 1990; Wilson 1998). The Carmel Formation's marine fossils are sometimes found in association with what had been cavities (Baker and Wilson 1999) or as ostreoliths (oyster balls) (Wilson et al. 1998). Terrestrial fossils include spores, pollen (Sprinkel et al. 2011), and crocodilian and dinosaur tracks (Lockley and Hunt 1995).

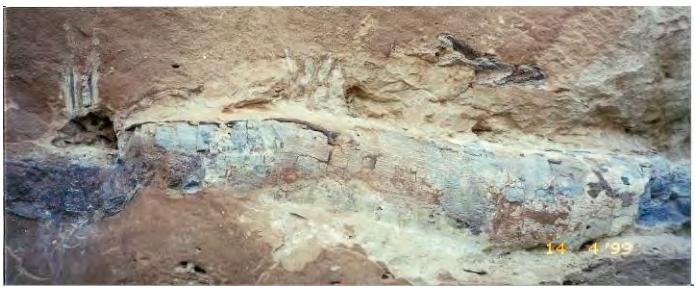
Morrison Formation (Upper Jurassic)

The Morrison Formation is a heterogeneous unit dated to the middle Late Jurassic; near eastern CARE, the Tidwell Member at the base of the formation has been dated to 157.2 ± 1.9 Ma, and rocks of the Brushy Basin Member near the top of the formation have been dated to 151.2 ± 1.8 Ma (Bradshaw and Kowallis 2010). CARE is near the western edge of Morrison Formation deposition (Turner and Peterson 1999). The formation is divisible into several members. From oldest to youngest, the Tidwell, Salt Wash, and Brushy Basin members are reported from CARE, although the Tidwell Member may not be present at the park, and if it is present, it is difficult to recognize and perhaps only 15 to 30 m (50 to 100 ft) thick at most (Morris et al. 2010). Salt Wash Member outcrops at CARE are heterogeneous, with rocks ranging from mudstone and claystone, to sandstone and conglomeratic sandstone, to dolomitic limestone and chert. The Salt Wash Member of the park is interpreted as a meandering stream and floodplain setting, with the sandstones representing channels and the finer rocks representing floodplains. The overlying Brushy Basin Member is similar but with more fine material, and is thought to represent a delta and lake setting (Petersen and Roylance 1982). The Salt Wash Member can be distinguished from the Brushy Basin Member by the greater percentage of sandstone and conglomerate in the Salt Wash Member, and the typical "popcorn" surface of dry Brushy Basin Member outcrops, due to the presence of shrink-swell clays of the smectite group. The Salt Wash Member is 30 to 150 m (100 to 500 ft) thick at the park, and the Brushy Basin Member is 60 to 105 m (200 to 350 ft) thick (Morris et al. 2010). The climate is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004).

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Brachiosaurus*, *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks, although the best finds have been to the north and east of CARE in Colorado, Utah, and Wyoming (Turner and Peterson 1999; Foster 2007). There are several reports of fossils in Morrison Formation rocks at CARE, mostly of fragmentary plant and vertebrate fossils. Root traces and burrows have been

found in the Tidwell and Salt Wash members of southern CARE (Kjemperud et al. 2008). The Salt Wash Member has also produced petrified logs and dinosaur bones in the park (Santucci and Kirkland 2010). Fragmentary bones of herbivorous dinosaurs are common in Salt Wash Member paleochannels, along with less-common petrified wood (Petersen and Roylance 1982). Unfortunately, some bone sites in the Morrison Formation of CARE have been vandalized (Mathis 2004). Smith et al. (1963), reporting on the Morrison Formation of CARE and the vicinity, noted the present of carbonaceous debris, petrified wood, and tree trunks in the Salt Wash Member, and abundant petrified wood and bone fragments in the Brushy Basin Member.

For the most part, Morrison Formation fossils from the vicinity are similar (CARE Figure 8). Large cycad trunks are known from six sites in the Brushy Basin Member within 10 km (6 miles) east and south of the park (Dayvault and Hatch 2005). A few fossils have been reported from the Morrison Formation of GLCA, including termite nests (Engelmann 1999; Santucci and Kirkland 2010), dinosaur bone fragments (Anderson et al. 2010), and the first sauropod track with skin impressions, from the Tidwell Member (Lockley et al. 1998). Pterosaur tracks (*Pteraichnus*) are known from the Tidwell Member of the Del Monte Mines area, and Lost Spring Wash has casts of sauropod tracks under a ledge, in the Salt Wash Member (Lockley et al. 1998). GSENM has yielded plants, nondinosaurian reptiles, and dinosaurs (Gillette 1997). The nearest site with vertebrate fossils identifiable at the genus level is the Hanksville Quarry, 38 km (24 miles) east of CARE. Fossils of *Apatosaurus* and an unidentified theropod were recovered from this site, low in the Brushy Basin Member (Turner and Peterson 1999). There is also a report of three partial reptile skeletons found east of Notom (Gregory and Anderson 1939), but this does not seem to have attracted any further interest. Finally, reworked Paleozoic bryozoans and crinoids are present in Salt Wash Member rocks of the Circle Cliffs (Davidson 1967).



CARE Figure 8. A sauropod humerus found in the Salt Wash Member just outside of CARE (VLS).

The potential for diverse fossils in Morrison Formation rocks at CARE exists, because the Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites, fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans, ostracodes, and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods, gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006); however, some of the traces, such as those attributed to anthozoans, brachiopods, horseshoe crabs, and echinoderms are only known from marine-influenced units that are not found at CARE (Hasiotis 2004). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians, lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods, diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

Dakota Sandstone (Upper Cretaceous)

The Dakota Sandstone (also Formation or Group) is a heterogeneous rock unit with outcrops as far from Utah as Minnesota; because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in Utah may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). The Dakota Sandstone of CARE is 0 to 45 m (0 to 150 ft) thick, and is composed of very-fine- to fine-grained sandstone

with thin beds of coal, carbonaceous clay, and conglomerate (Morris et al. 2010). The Dakota Sandstone of the Colorado Plateau is often divided into three informal members, with a conglomeratic base, a heterogeneous coaly middle, and an upper marine sandstone. This division reflects environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). In the Hanksville area to the east, the Dakota Sandstone formed a large wedge of sediment into the advancing sea that later subsided and was covered by an oyster bank (Lawyer 1972). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma. In GSENM to the south, the middle of the formation was deposited around 96.0 Ma (Titus et al. 2005). The lower contact with the Cedar Mountain Formation is disconformable (Morris et al. 2010).

The fossils of the Dakota Sandstone are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. The Dakota Sandstone of CARE is locally fossiliferous, including a dense bivalve reef dominated by the oysters *Exogyra* and *Pycnodonte* (Santucci and Kirkland 2010). This site is known as Oyster Reef, and can be observed within CARE a few yards east of Notom–Bullfrog Road roughly 39 km (24 miles) from the turnoff of Highway 24 (Morris et al. 2010). This reef also extends outside of the park (Mathis 2004). Other fossils of the CARE area include coal, and wood and bone rounded into pebbles (Gregory and Anderson 1939).

The best-known terrestrial fossils from the Dakota Sandstone are plant fossils (particularly angiosperm leaves) and vertebrate tracks. The Dakota Sandstone is particularly noted for its paleobotanical material. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Sandstone (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Sandstone of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Sandstone for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). The floral assemblage of Utah outcrops includes coal (Molenaar and Cobban 1991), palynomorphs (moss and fern spores, and gymnosperm and angiosperm pollen; Nichols 1997), the tree fern *Tempskya* (Ash and Read 1976), *Equisetum*, possible gymnosperms, and angiosperms (Tidwell et al. 1976). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks come from Dakota Group rocks that are several million years older than the rocks at GLCA, and include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Formation rocks in Utah reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals; locations included areas near Bryce Canyon National Park (BRCA; see the BRCA summary in this report for more details) (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees and wasps (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Sandstone marine fossils include algae, foraminifera (Tibert et al. 2003), dinoflagellate cysts (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Sandstone may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera, corals, bryozoans, brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Dakota Sandstone outcrops of the nearby Kaiparowits Plateau and the rest of GSENM have yielded diverse fossils, and provide a model for what might be found at CARE. These include algae, foraminifera (Tibert et al. 2003), coal, palynomorphs, petrified wood (Doelling 1977), leaves (Erwin 1997), bivalves, gastropods, ostracodes (Tibert et al. 2003), ammonites (Foster et al. 2001), a dragonfly nymph (Titus et al. 2005), invertebrate traces (Tibert et al. 2003), sharks, rays, ray-finned fish, lungfish, amphibians, turtles, lizards, several types of crocodilians and crocodile relatives, small theropods such as dromaeosaurids and troodontids, tyrannosaurids, armored dinosaurs, hypsilophodonts (small bipedal herbivorous dinosaurs), hadrosaurids (duckbills), mammals (Eaton and Cifelli 1997; Eaton et al. 1999; Eaton and Kirkland 2001, 2003; Titus et al. 2005), and footprint sites (Hamblin and Foster 2000). Most fossils there were found in floodplain settings of the middle unit, with some from lacustrine rocks (Eaton and Cifelli 1997; Eaton et al. 1999), and algae, foraminifera, gastropods, bivalves, ostracodes, and burrows from the upper part (Tibert et al. 2003).

Mancos Shale (Upper Cretaceous)

The Mancos Shale is a widespread unit that marks the advances and retreats of the Western Interior Seaway. Its name is something of a misnomer at CARE, because a significant amount of it is made up of sandstone. The formation is divisible into a number of members depending on location. At CARE, it is divided from oldest to youngest into the Tununk, Ferron Sandstone, Blue Gate Shale, Muley Canyon, and Masuk members (Morris et al. 2010). In the nearby Henry Mountains, the Mancos Shale includes parts of four transgressive-regressive cycles. The Tununk and Ferron members represent most of the cycle began during the deposition of the Dakota Sandstone, the Blue Gate Shale includes two cycles, and the Muley Canyon and Masuk members include part of a cycle ended during the deposition of the overlying Tarantula Mesa Sandstone of the Mesaverde Group (Peterson et al. 1980). This cycle marked the last regression of the epicontinental Western Interior Seaway (Eaton 1990).

The Tununk Member of CARE is between 160 and 220 m (540 and 720 ft) thick, and is composed of bluish-gray slope-forming shale with interspersed beds of mudstone, siltstone, and sandstone. It is interpreted as representing open marine deposition (Morris et al. 2010). An altered ash layer present in the Tununk Shale of the Caineville Reef area just east of CARE has been dated to 93.46 ± 0.6 Ma and 93.25 ± 0.55 Ma. Shell beds (of *Pycnodonte*) and shark teeth are present at this location (Kirkland and Madsen 2007). Volcanic ash may have provided nutrients for organisms (Bagshaw 1977). The upper part of the member intertongues with the lower Ferron Sandstone (Billingsley et al. 1987). The water depth during deposition in the area was 90 to 180 m (300 to 600 ft), with normal marine conditions (Peterson et al. 1980). Deposition was in advance of deltas, with several minor sea-level fluctuations (Sethi and Leithold 1997).

The fossils of the Tununk Member are largely derived from marine mollusks, and some, like the bivalve *Gryphaea*, are common (Mathis 2004). It has yielded coccoliths (microscopic calcium carbonate plates from algae) from fecal pellets (Leithold 1993), foraminifera, bivalves, ammonites (Peterson et al. 1980), gastropods (Santucci and Kirkland 2010), ostracodes (Peterson et al. 1980), burrows (Molenaar and Cobban 1991), shark teeth (Santucci and Kirkland 2010), and scales (Morris et al. 2010) and teeth (Bagshaw 1977) of bony fish. Bivalves, ammonites, gastropods, and shark teeth are present in this member in and around CARE (Santucci and Kirkland 2010). In the vicinity, bivalves have been found in this unit in the Circle Cliffs (Davidson 1967), and foraminifera, bivalves (including *Gryphaea* and *Inoceramus*), ostracodes, and fish teeth have been found near Caineville (Bagshaw 1977).

The Ferron Sandstone of CARE is between 60 and 115 m (205 and 385 ft) thick, and is a cliff-forming marine sandstone unit deposited in middle and lower shoreface settings created by deltas (Morris et al. 2010). It is a sandstone tongue into the marine Mancos Shale (Dickinson and Gehrels 2008), representing the culmination of a regressive phase (Molenaar and Cobban 1991). At some locations it is divisible into smaller units (Cotter 1975). A variety of deltaic and coastal depositional settings are known. Two large deltas were present in central Utah, the northern Vernal Delta and the southern Last Chance Delta (Cotter 1975). Five principal deltaic cycles are known from the unit, as well as multiple ash beds (Crowley et al. 1994). To the northeast in the Uinta Basin, the Ferron Sandstone dates to approximately 89.7 Ma (Molenaar and Cobban 1991). Its importance to the fossil fuel community has made it exceptionally well-known (Morris et al. 2010).

The fossils of the Ferron Sandstone include both marine and terrestrial taxa. The assemblage includes organic debris (Cotter 1975), coal, palynomorphs (representing acritarchs [miscellaneous organic microfossils], dinoflagellates, green algae, ferns, sphagnum mosses, gymnosperms, and angiosperms) (Crowley et al. 1994), plant fragments, seed cones, leaves (Ursek 1979), ferns, gymnosperms, angiosperms (Young 1987), angiosperm logs (Ursek 1979), petrified wood (Smith et al. 1963), root traces (Ryer 1981), bryozoans (Hill 1982), bivalves, ammonites (Molenaar and Cobban 1991), gastropods (Ryer 1981), echinoids (Hill 1982), marine invertebrate burrows and trails (Cotter 1973), possible cephalopod coprolites (Ursek 1979), sharks, rays (Becker et al. 2010), bony fish (Becker and Linteris 2009), and ornithopod dinosaur tracks (Jones 2001). Marine fossils such as *Inoceramus* are found throughout the member; coal and plant fragments appear in the upper part (Peterson et al. 1980). Coal, petrified wood, and leaf impressions are found in the Ferron Sandstone of the CARE area, particularly in the upper part of the unit (Smith et al. 1963). Nearby, near Caineville, coal, plant fragments, seed cones, leaves, angiosperm logs, a shark tooth, and fish scales have been found in the upper Ferron Sandstone (Ursek 1979).

The Blue Gate Shale of CARE is between 370 and 460 m (1,200 and 1,500 ft) thick. It is composed of bentonitic clay, mudstones, and minor limestones. Like the Tununk Member, it weathers as slopes and is interpreted as an open marine unit (Morris et al. 2010), with depths of 60 to 120 m (200 to 400 ft) (Peterson et al. 1980). The upper part of the member intertongues with the lower Muley Canyon Member (Billingsley et al. 1987 [as the Emery Sandstone Member]).

Fossils of the Blue Gate Shale mostly come from marine invertebrates. The unit has yielded algal biscuits (lumps of calcium carbonate produced by algae) (Cross 1976), bivalves, ammonite and belemnite cephalopods, gastropods (Eaton 1990), trace fossils (Riemersma and Chan 1991), and shark teeth (Santucci and Kirkland 2010). Bivalves, ammonites, gastropods, and shark teeth are present in this member in and around CARE (Santucci and Kirkland 2010), and cephalopods are known from the Circle Cliffs (Davidson 1967). Shells can be abundant enough in the upper part of the member to form coquinas (Riemersma and Chan 1991).

The Muley Canyon Member of CARE is 90 to 120 m (300 to 400 ft) thick, and is composed of fine-grained sandstone and carbonaceous shale. Its depositional setting changed from marine shoreface to coastal plain over time (Morris et al. 2010). This unit was formerly known as the Blue Gate Sandstone Member or Emery Sandstone Member (Smith 1983). The Muley Canyon Member intertongues with the Masuk Member (Billingsley et al. 1987), and correlates to the Star Point Sandstone of the Mesaverde Group to the northeast in Castle Valley (Peterson et al. 1980), and the Drip Tank Member of the Straight Cliffs Formation in GLCA (Eaton 1990). It dates to the early Campanian stage of the Late Cretaceous (Eaton 1990), sometime between approximately 83.5 Ma and 79 Ma.

The fossil assemblage of the Muley Canyon Member is not as extensive as those of some other members. Coal, plant fragments (Peterson et al. 1980), petrified wood, leaf impressions (Smith et al. 1963), and trace fossils are known from the unit (Peterson et al. 1980). Coal, petrified wood, and leaf impressions have been found in the CARE area, particularly in the upper part of the unit (Smith et al. 1963); similar to the Ferron Sandstone, terrestrial fossils appear in the upper part of the member (Peterson et al. 1980).

Finally, the Masuk Member of CARE is 200 to 230 m (650 to 750 ft) thick, and is composed of alternating slope-forming mudstones and cliff-forming sandstones. These alternating beds are interpreted as stemming from sea level fluctuations; sometimes the area was a shallow marine setting, and other times it was a coastal plain (Morris et al. 2010). It dates to a younger part of the early Campanian stage than the Muley Canyon Member (Eaton 1990). In the Henry Basin, it can be divided into three informal units: a lower sandstone and coal unit; a middle mudstone and carbonaceous shale unit, and an upper sandstone and carbonaceous siltstone unit (Eaton 1990). The Masuk Member correlates to the Blackhawk Formation of the Mesaverde Group to the northeast in Castle Valley (Peterson et al. 1980).

The Masuk Member has both terrestrial and marine fossils. Its fossil assemblage includes palynomorphs, coal, wood, root traces (Eaton 1990), marine shells (Davidson 1967), brackish-water bivalves (Peterson et al. 1980), freshwater bivalves and gastropods, crabs, burrows, hybodont sharks, rays, gars, bowfins, teleost fish, salamanders, frogs, turtles, crocodiles, dinosaurs (theropods, nodosaurids, pachycephalosaurids or "bone heads" [Eaton 1990], ceratopsians or "horned dinosaurs" [Peterson et al. 1980], and hadrosaurids, represented by teeth), and multituberculate and marsupial mammals (Eaton 1990). Petrified wood has been found in this unit in the CARE region (Smith et al. 1963). In the Circle Cliffs, marine shells and shark teeth have been found in the lower, shallow-water part of this member (Davidson 1967).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary deposits of CARE include boulder deposits, pediment deposits, terrace gravels, colluvial and alluvial deposits, and eolian deposits, with a mix of rock sizes from boulders to clay (Graham 2006). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). To date, CARE has only paleoecological materials, but some large vertebrate specimens are known from the immediate vicinity. Summaries of Quaternary fossil sites in southern Utah and the vicinity include Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008).

22 packrat middens have been described from CARE (Cole et al. 1997; Cole and Murray 1999). The majority of these were constructed after 1400 AD, but ten samples have yielded more ancient dates. Two samples from the large midden denoted Hall's Canyon #2 have been dated to >39,600 years old (older than the maximum age for radiocarbon dating) and 28,050 ± 2,600 radiocarbon years before present (38,560–27,570 calendar years before present) (CARE Figure 9).



CARE Figure 9. The large ancient midden Hall's Canyon #2. (photo courtesy Kenneth Cole, Northern Arizona University).

Three more CARE middens are over 3,000 years old: Fremont River 1 (7,010 ± 105 radiocarbon years before present, 7,980–7,740 calendar years before present), Hartnet Draw 6 (5,450 ± 90 radiocarbon years before present, 6,360–6,240 calendar years before present), and Hartnet Draw 5 (3,615 ± 70 radiocarbon years before present, 4,130–3,890 calendar years before present). The other five were constructed between 3,000 and 600 years ago (from oldest to youngest, Hartnet Draw 9, Hartnet Draw 9, Long Leaf Flat 1B and approximately contemporaneous Hartnet Draw 1A, and Hartnet Draw 8) (Cole and Murray 1999). Numerous fossil middens are also present along the Fremont River between park headquarters and Deep Creek, but

they are difficult to access, and only one has been dated (Cole and Murray 1999). The packrat middens of CARE have pollen, needles, other plant macrofossils, and packrat droppings (Cole et al. 1997; Cole and Murray 1999). Comparisons of midden plant assemblages from before grazing activities to plants from more recent middens have been of particular interest (Cole and Henderson 1993; Cole et al. 1997; Cole and Murray 1999). There is the possibility of finding additional fossils in alcoves and rock shelters within CARE, particularly packrat middens. Santucci et al. (2001) presented an overview and cited selected examples of National Park Service fossils found in caves, including at Arches National Park (ARCH), GLCA, and Natural Bridges National Monument (NABR).

A few other paleontological resources or potential paleontological resources are known from the park and immediate vicinity. Opal phytoliths (silica structures found in some plant tissues) are known from four sites within CARE (Cathedral, Hartnet, Post, and Surprise sites), dating to between approximately 800 and 250 years old (Fisher et al. 1995), suggesting the presence of older phytoliths in the park's surficial sediments. Thin peat layers are present in alluvial deposits of this age in the CARE region (Graham 2006). Holocene pollen and plant macrofossils are known from Fryingpan Lake on the Fish Lake Plateau, 24 km (15 miles) northwest of CARE, and Posy Lake on the Aquarius Plateau, 47 km (29 miles) west of CARE (Shafer 1989). Faunal remains are known from the Pleistocene and Holocene. From the Pleistocene, mammoth fossils are known from Boulder Mountain to the west (Santucci and Kirkland 2010). Holocene alluvium in the region has yielded gastropods, a bighorn sheep skull, an elk antler, a jawbone possibly from a deer, indeterminate bones, and hearth sites with charcoal and artifacts (Smith et al. 1963).

The most diverse fossil sites within 100 km (60 miles) of CARE are Bechan Cave within GLCA, and Cowboy Cave, over 75 km (46 miles) east of CARE. Bechan Cave is the best-known and most diverse of several Quaternary alcove and rock shelter sites at GLCA. It is found near the confluence of the Colorado River and Escalante River. Containing 300 m^3 (10,600 ft^3) of fossil dung among other fossil resources, Bechan Cave has been described in a number of publications. Its fossils include: fungal spores in droppings (Davis 1987; Davis and Shafer 2006); pollen (Davis 1990); charcoal, wood (Agenbroad and Mead 1989), seeds and plant fragments (Davis et al. 1984); insects (Elias et al. 1992); dung assigned to packrats, cottontail rabbits, Shasta ground sloths (*Nothrotheriops shastensis*), mammoths, possibly horses, the extinct shrub ox *Euceratherium collinum*, bighorn sheep, and extinct mountain goats; hair from shrews, packrats and other rodents, bats, coyotes, bear, ground and mylodont sloths (*Glossotherium* and *Nothrotheriops*), mammoths, horses, bison, bighorn sheep, and deer; bones from toads, snakes, grouse-sized birds, a variety of rodents, rabbits, and *Euceratherium*; and packrat middens covering the last 12,000 years (Agenbroad and Mead 1987, 1990; Mead and Agenbroad 1992; Santucci et al. 2001). Human artifacts are also known from younger levels of the cave (Agenbroad et al. 1989). Fourteen plant samples and seven dung samples from mammoths and shrub oxen have been dated from Bechan Cave, yielding a range that spans from 13,505 ± 580 radiocarbon years before present (17,835–14,460 calendar years before present) to 11,630 ± 150 radiocarbon years before present (13,800–13,215 calendar years before present) (Mead and Agenbroad 1992).

Cowboy Cave and its contents were extensively described by Jennings (1980). It is notable for having both extensive Pleistocene material and four distinct Holocene cultural layers above, spanning the last 13,000 years at least (humans present only since approximately 9,000 years ago). In the Pleistocene layer, various types of mammal dung and hair were found with packrat middens and a handful of bones, mostly bison, but also two juvenile mammoth tusks and a possible elk bone (Barnett and Coulam 1980; Hansen 1980; Jennings 1980; Spaulding and Peterson 1980). The cultural levels show human use of stone, bones, hair, hides, shells, plant parts, and other local organic materials to make tools, ornaments, bags, and other artifacts (Hewitt 1980; Hull 1980a, 1980b; Janetski 1980; Lucius 1980; Weder 1980). Prehistoric dung at the cave mostly came from bison, but also includes feces from Shasta ground sloths, elephants, horses, elk, elk or camel (Mead and Agenbroad 1992), rabbits, rodents (Hansen 1980), and humans (Hogan 1980).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within CARE. However, they are known to preserve fossils elsewhere, and future field investigations within the park may recover fossils from one or more of them.

Glen Canyon Group: Wingate Sandstone (Upper Triassic–Lower Jurassic)

The Wingate Sandstone is an eolian sandstone unit that spans the Triassic–Jurassic boundary (Lucas et al. 2006a). Limestone lenses are also present (Harshbarger et al. 1957). At CARE, it is the basal unit of the Glen Canyon Group. The park's exposures are up to 105 m (350 ft) thick, and are composed of resistant sandstone that weathers salmon in color. The lower contact with the Chinle Formation is a disconformity, and the upper contact with the Kayenta Formation is gradational at the park (Morris et al. 2010). The Wingate Sandstone is the eastern dry eolian equivalent of the western Moenave Formation wet eolian rocks (Lucas and Tanner 2007), and interfingers at its western contact with the Dinosaur Canyon Member of the Moenave Formation (Donohoo-Hurley et al. 2007). Wingate Sandstone dunes were driven by winds from the north and northwest (Dubiel et al.

1989). After much debate concerning the age of the Wingate Sandstone and other Glen Canyon Group units, it was thought that the Wingate Sandstone was entirely Jurassic in age, but the discovery of a skull of the phytosaur *Redondasaurus* at the base of the formation in Lisbon Valley shows that at least some of the unit is Late Triassic in age (Lucas et al. 1997a). Trace fossils suggest that the Triassic–Jurassic boundary is in the middle to upper 75 m (246 ft) of the formation (Lucas et al. 2006a).

The Wingate Sandstone has had a complicated history (Colbert and Mook 1951). It was originally defined to contain much the same rocks as what was later named the Entrada Sandstone, but the Entrada Sandstone eventually took over most of these rocks, including the original type section for the Wingate Sandstone (at Fort Wingate, New Mexico). Furthermore, the Wingate Sandstone was later divided into the lower Rock Point Member and the upper Lukachukai Member, but where the Rock Point Member has been assigned to the Chinle Formation, the Lukachukai Member is abandoned as redundant (Dubiel 1989).

Fossils are not currently known from the Wingate Sandstone of CARE. In general, fossils are uncommon in the Wingate Sandstone, with the phytosaur skull mentioned above being the only body fossil of note, but trace fossils are known from multiple levels in Utah, Arizona, and Colorado. Trace producers include invertebrates (Odier et al. 2004), mammal-like animals, nondinosaurian reptiles, and theropod, prosauropod, and possibly sauropod dinosaurs (Odier et al. 2004; Lucas et al. 2006a, 2006b). Mammal-like tracks are often found nearly alone and are confined to dunes (Odier et al. 2004). There appears to be distinct Late Triassic and Early Jurassic assemblages, with mammal-like tracks limited to the Late Triassic and prosauropod tracks appearing in the Early Jurassic (Lucas et al. 2006a, 2006b). Several track sites are known to the south in GLCA (Lockley et al. 1998) and GSENM (Hamblin and Foster 2000).

San Rafael Group: Page Sandstone (Middle Jurassic)

The Page Sandstone is another eolian sandstone, very similar to the Navajo Sandstone. It was originally thought to be the upper part of the Navajo Sandstone (see for example Wright and Dickey 1963), and is sometimes still described as such (Lucas et al. 2005). The Page Sandstone of CARE is between 0 and 30 m (0 and 100 ft) thick, and is composed of sandstones that are interpreted as having been deposited on an arid coastal plain with periodic marine incursions (Morris et al. 2010). It is separated from the Navajo Sandstone by a disconformity; at GLCA to the south, this is marked by a thin chert lag, but this is not true everywhere (Anderson et al. 2010). Volcanic deposits indicate that it dates to the Middle Jurassic, between about 171.5 and 169.5 Ma (Dickinson et al. 2010).

As discussed above in the "Carmel Formation" section, the Page Sandstone complexly intertongues with the lower Carmel Formation. At CARE, the largest subdivisions in ascending order are the Harris Wash and Thousand Pocket members (sometimes called tongues), separated by the lower part of the Carmel Formation (Doelling and Willis 1999; Doelling and Kuehne 2007; Martin et al. 2007; McLelland et al. 2007; Sorber et al. 2007; Morris et al. 2010). To maintain consistency with the CARE GRI report (Graham 2006), the Judd Hollow Member will be discussed here with the Page Sandstone. The Judd Hollow and Thousand Pockets members overlap in southern CARE, with the limestone and dolostone of the Judd Hollow Member wedging out to the southeast and the sandstone of Thousand Pockets Member wedging out to the northwest (Wright and Dickey 1963).

Fossils are extremely rare in the Page Sandstone; however, wood is reported from the Page Sandstone at GLCA (NPS 1999b). Otherwise, to date there have only been reports of echinoderm fragments (Blakey et al. 1983), and a single possible theropod footprint from the Thousand Pockets Member in GSENM (Foster et al. 2001). Rocks of the intertongued Judd Hollow Member of the Carmel Formation from near the Cockscomb of GSENM have yielded dinoflagellates, spores, and abundant and diverse pollen (Ash and Litwin 2002).

San Rafael Group: Entrada Sandstone (Middle Jurassic)

The Entrada Sandstone is dominated by red sandstone, with some silt (Harshbarger et al. 1957). It is divided into multiple members depending on location. It lacks age-diagnostic fossils and is regarded as late Middle Jurassic in age (Lucas et al. 1985), approximately 165 Ma. The Entrada Sandstone of CARE is between 120 and 275 m (400 and 900 ft) thick and is composed of reddish-orange siltstone and sandstone. Entrada Sandstone rocks change from slope-forming siltstone in southern CARE to slope- and cliff-forming sandstone and siltstone in northern CARE. Like the Carmel Formation, weathering of the Entrada Sandstone has produced some of CARE's unique scenery, in this case the monoliths of Cathedral Valley. While usually thought of as an eolian sandstone unit, the Entrada Sandstone of CARE is interpreted as representing tidal flat deposits (Morris et al. 2010). The majority of the Entrada Sandstone represents a semi-arid to arid dune field (Ekdale and Picard 1985) that was hot and sparsely vegetated. It was the last major erg of the area, and formed when southern Utah was about 22° N (Loope 2008). The lower contact with the Carmel Sandstone is conformable at CARE, and the upper contact with the Curtis or Summerville formations is disconformable (Morris et al. 2010).

Fossils are not known from the Entrada Sandstone at CARE, and fossils are generally rare in the formation. Many reports of fossils from the Entrada Sandstone are from the Moab Member, which is now regarded as part of the Curtis Formation (Doelling 2010). These include the well-known megatracksite of the Moab area (Lockley 1991; Lockley et al. 2007) and the crocodylomorph *Entradasuchus* (Hunt and Lockley 1995). Trace fossils have been reported from the Escalante Member in

south-central Utah, including rare root traces, insect burrows (Loope 2008), vertebrate burrows, traces of sand-swimming vertebrates (Graverson et al. 2007), and theropod (large *Megalosauripus* and smaller *Therangospodus*) and sauropod (cf. *Brontopodus*) dinosaur tracks (Foster et al. 2000). At GLCA, some theropod tracks from Lake Powell have been reassigned to the Entrada Sandstone from the Navajo Sandstone (Lockley et al. 2005). Other nearby trace fossil finds include insect burrows, large vertebrate burrows, dinosaur tracks, and sand-swimming traces from near Escalante in GSENM (Loope 2004). Hoggan (1970) reported bivalves and crinoids in marine facies of the Entrada Sandstone in the Vernal area.

San Rafael Group: Curtis Formation (Upper Jurassic)

The Curtis Formation is a shallow marine unit representing part of the last marine transgression of the Jurassic at CARE. The Curtis Formation of CARE is thickest in the northern part of the park, up to 55 m (175 ft), thinning to a feather edge or even disappearing in the southern part of the park. It is made up of light grayish-green sandstone and siltstone with some sandy limestone beds, and caps the monoliths and cathedral in northern CARE (Morris et al. 2010). Its lower contact with the Entrada Sandstone is disconformable in central Utah, and its upper and lateral contacts with the Summerville Formation are conformable to intertonguing (Blakey 1989). The Curtis Formation was previously thought to be late Middle Jurassic in age (Lucas and Anderson 1997), but is now known to date to the early Late Jurassic (Oxfordian stage), sometime between approximately 161 and 156 Ma (Wilcox 2007; Hintze and Kowallis 2009).

The sea that deposited the Curtis Formation was on the order of 30 m (100 ft) deep at its greatest in Utah, and retreated during deposition (Hoggan 1970). When it first advanced, the sea caused the collapse of the massive dunes of the Entrada erg (estimated as up to 100 m [300 ft] tall), leading to the formation of a disrupted basal sandstone layer for part of the Curtis Formation (Eschner and Kocurek 1986). Tidal deposits are also present in central Utah (Kreisa and Moila 1986).

The Curtis Formation is not known to be fossiliferous at CARE. It has a marine fossil assemblage, better known in northeastern Utah (Hoggan [1970], although it is likely that at least some of the fossils pertain to an overlying unit; see the Dinosaur National Monument summary in this report for more details) than in the vicinity of CARE. The assemblage includes dinoflagellates (Evitt 1961), foraminifera, coral-like organisms, brachiopods, bivalves, ammonite and belemnite cephalopods, gastropods, serpulid worms, possible crustaceans, ostracodes, crinoids, echinoids, holothurians (sea cucumbers), invertebrate trace fossils, and fragments of vertebrates (Hoggan 1970). Some burrows are present in the basal sandstone (Eschner and Kocurek 1986). Terrestrial fossils, such as those found in the Moab Member, are unlikely to be found at CARE because of the lack of Curtis Formation terrestrial deposition.

San Rafael Group: Summerville Formation (Upper Jurassic)

The Summerville Formation is 15 to 75 m (50 to 250 ft) thick at CARE and is composed of thin-bedded siltstones and mudstones, smaller quantities of greenish-gray limestone and sandstone, and gypsum beds up to 9 m (28 ft) thick (Morris et al. 2010). It grades vertically and horizontally into the underlying upper Curtis Formation (Blakey 1989), and has a disconformable upper contact with the overlying Morrison Formation (Kjemperud et al. 2008).

In the past, the Summerville Formation was usually thought to have been deposited during the late Middle Jurassic (Steiner 2003), but like the underlying Curtis Formation, it is now known to date to the Oxfordian stage of the early Late Jurassic (Wilcox 2007; Hintze and Kowallis 2009). The Tidwell Member is sometimes transferred to the Summerville Formation, making it the uppermost member (Lucas and Anderson 1997), but this has not been widely adopted. If the Tidwell Member is included, the lower part becomes the Beclabito Member, and the formation is extended into the earliest Late Jurassic (Lucas and Anderson 1997). This debate may not have much meaning for CARE, because of the issues surrounding the Tidwell Member at the park as discussed above (Morris et al. 2010). Deposition was affected by occasional flooding from the regressing Sundance Sea (Petersen and Pack 1982). Aridity decreased during deposition (Lockley and Wright 2003), as Pangaea continued to drift north (Lockley and Mickelson 1997). The Summerville Formation of CARE is interpreted as representing hypersaline shallow marine and tidal flat deposits (Mathis 2004; Morris et al. 2010).

The Summerville Formation is not known to be fossiliferous at CARE. Summerville Formation fossils include plant debris (Petersen and Pack 1982), a single sauropod tail vertebra (Lucas and Heckert 2000), invertebrate traces (Lockley and Wright 2003), and a variety of vertebrate traces including pterosaur tracks (Lockley et al. 1995; Lockley and Wright 2003; Mickelson et al. 2004; Lockley et al. 2007), theropod tracks (Lockley et al. 1996), and sauropod tracks from near the top of the Summerville Formation at Moab (Lockley 1991). Twigs and plant debris are known from channel sandstones just east of CARE, but otherwise the Summerville Formation of the CARE area is thought to have been a hostile environment (Petersen and Pack 1982). If the Tidwell Member is included in the Summerville Formation, the fossil assemblage is broadened to include root traces (Kjemperud et al. 2008), algal and microbial laminations, charophytes, gastropods, ostracodes (Scott et al. 2001a), burrows (Kjemperud et al. 2008), horseshoe crab traces (Scott et al. 2001b), a lungfish tooth plate (Imhof and Trujillo 2005), indeterminate reptile bones (the putative pterosaur *Utahdactylus*; Bennett 2007), the poorly-known sauropod *Dystrophaeus* (Gillette 1996; Lucas and Anderson 1997), and pterosaur and dinosaur tracks (Lockley et al. 1998). Because some of the Tidwell Member root traces and burrows are known

from southern CARE (Kjemperud et al. 2008), this would make the Summerville Formation fossiliferous within the park.

Cedar Mountain Formation (Lower–lower Upper Cretaceous)

The Cedar Mountain Formation is similar to the underlying Morrison Formation, and the two can be difficult to distinguish. It is between 0 and 51 m (0 and 166 ft) thick at CARE, typically made of slope-forming sandstone and mudstones that are relatively more pastel-colored than those of the Morrison Formation. In other locations, the contact between the two formations is marked by the Buckhorn Conglomerate Member of the Cedar Mountain Formation, but this unit is nearly absent from CARE (Morris et al. 2010). The Cedar Mountain Formation has been divided into several members. Of these, the only members present in the vicinity of CARE are, from oldest to youngest, the Buckhorn Conglomerate, Ruby Ranch, and Mussentuchit Member. The formation is only present in the northeast section of the park (Kirkland and Madsen 2007). It is interpreted as a fluvial terrestrial formation (Morris et al. 2010). Mudstones represent overbank deposits, and sandstones represent the rivers and streams (Craig 1981). The climate became more humid over time, with high temperatures (Garrison et al. 2007). Deposition of the Cedar Mountain Formation occurred at the same time as the early Sevier Orogeny to the west (Fiorillo 1999).

The Cedar Mountain Formation is the lateral equivalent of the Burro Canyon Formation, as found at Colorado National Monument (COLM) and Hovenweep National Monument (HOVE). Generally, the Burro Canyon Formation is present east of the Colorado River, and the Cedar Mountain Formation is present west of it (Dickinson and Gehrels 2008). The two intertongue in places (Dayvault and Hatch 2005). The upper contact with the Dakota Sandstone is a disconformity (Morris et al. 2010), while the lower contact with the Morrison Formation is usually described as a disconformity but may not be disconformable everywhere (see below).

The dating of the Cedar Mountain Formation is complex. The upper contact is established at approximately 97 Ma (Britt et al. 2009), but the internal dating of the various members and the lower contact are uncertain. For example, the Mussentuchit Member was thought to have begun deposition at approximately 98.37 ± 0.07 Ma (Kirkland and Madsen 2007). Then several quarries were found to be placed incorrectly in the Ruby Ranch Member and were actually from the Mussentuchit Member, including the Price River Two Quarry, which was dated to between 116 and 109 Ma by detrital zircons (Carpenter et al. 2008; although note that Kirkland and Madsen [2007] suggest that this age is too old and derives from older reworked zircons). At the base of the formation, the Buckhorn Conglomerate is usually thought of as dating to the Barremian stage of the Early Cretaceous (Dickinson and Gehrels 2008), approximately 130 to 125 Ma. However, its age is controversial (Kirkland and Madsen 2007). There are places where it appears to intertongue with the Brushy Basin Member of the Morrison Formation, which would make it at least partially Late Jurassic in age (Kirkland et al. 1999). Additionally, there are reports of basal Cretaceous ostracodes in the lowest Cedar Mountain Formation of the San Rafael Swell (Sames et al. 2010). Internal stratigraphy can also be problematic, as several fossiliferous quarries have been transferred from one member to another (Kirkland and Madsen 2007; Carpenter et al. 2008). Together, these suggest that the basal age and stratigraphic relationships of the formation are still incompletely known.

For many years, the fossils of the Cedar Mountain Formation were poorly known. This began to change with a series of publications on fossil plants in the late 1970s and early 1980s (Tidwell et al. 1976; Thayn et al. 1983, 1985; Thayn and Tidwell 1984; Tschudy et al. 1984; Tidwell and Thayn 1985). Then, during the later 1980s and 1990s, rich vertebrate localities were found at several levels in the formation, and it became an important stratigraphic unit for the study of Cretaceous vertebrates, fitting between the well-studied Morrison Formation of the Upper Jurassic and various Upper Cretaceous formations of the American West. The Cedar Mountain Formation is not currently known to be fossiliferous at CARE, although turtle and dinosaur fossils are known just north of the park (Santucci and Kirkland 2010).

Cedar Mountain Formation (Lower Cretaceous): Buckhorn Conglomerate Member

The Buckhorn Conglomerate Member represents the base of the Cedar Mountain Formation in its western depositional area, and is interpreted as a river system flowing northeast along the western flank of the San Rafael Swell (Kirkland and Madsen 2007). It is thought to be at least partially equivalent to the Yellow Cat and overlying Poison Strip Sandstone members of eastern Utah (Kirkland and Madsen 2007), and may intertongue with the Yellow Cat Member (Dickinson and Gehrels 2008).

The Buckhorn Conglomerate Member is sparsely fossiliferous. It has yielded cycad stems (Furniss and Tidwell 1972), petrified logs, poorly preserved mollusk shells, fragmentary dinosaur bones (Stokes 1950), a partial skeleton of an ankylosaurian dinosaur (which would suggest a Cretaceous age; Kirkland and Madsen 2007), and reworked Paleozoic fossils (Kirkland and Madsen 2007).

Cedar Mountain Formation (Lower Cretaceous): Ruby Ranch Member

The Ruby Ranch Member is mostly composed of mudstone, with minor sandstone and limestone (Kirkland et al. 1999). It is notable for its abundant carbonate nodules, representing ancient soil horizons (paleosols) and ephemeral ponds. The surrounding sandstones were deposited by low-sinuosity rivers (Kirkland and Madsen 2007). The prevailing climate of the member is thought to have been semi-arid and monsoonal, like the Morrison Formation (Kirkland et al. 1999).

Overall, the Ruby Ranch Member's fossil assemblage includes pollen (Kirkland et al. 1999), root traces, arthropod trace fossils (Tremain 2003), hybodont sharks (Kirkland et al. 1999), the sphenodont *Toxolophosaurus*, crocodylomorphs, several types of theropods (an unnamed carnosaur, *Acrocanthosaurus*, and dromaeosaurids), brachiosaurid sauropods, nodosaurids, a new species of the polacanthid *Gastonia*, the long-tailed ornithopod *Tenontosaurus* (mostly known as a prey item of *Deinonychus*) (Kirkland and Madsen 2007), and tracks of theropods, sauropods, ankylosaurians, and ornithopods (Lockley et al. 2004). Additionally, a site in Utah near the border with Colorado (the Hotel Mesa Quarry) has yielded fossils of sharks, bony fish, lungfish, various crocodile relatives, theropods, the sauropod *Brontomerus*, and ornithopods (beaked bipedal dinosaurs) (Taylor et al. 2011). It has been attributed to both the Burro Canyon Formation (Kirkland and Madsen 2007) and the Ruby Ranch Member of the Cedar Mountain Formation; by geography, it is within the Burro Canyon Formation (Taylor et al. 2011). The dinosaurs of the Ruby Ranch Member are similar to those of the Arundel Formation of Maryland, Cloverly Formation of Montana and Wyoming, and Trinity Group of Oklahoma and Texas (Kirkland et al. 1999). However, with the reassessment of quarry stratigraphic position it is not clear how many of these fossils are actually from the member. There is at least a large unnamed nodosaurid present in the middle of the unit (Carpenter et al. 2008), and the unnamed carnosaur and brachiosaurid ("*Astrodon*") are known from the Long Walk Quarry low in the member (Kirkland and Madsen 2007).

Cedar Mountain Formation (Lower Cretaceous): Mussentuchit Member

The contact of the Ruby Ranch Member and overlying Mussentuchit Member is often marked by a black chert pebble lag. Unlike the Ruby Ranch Member, the Mussentuchit Member has few carbonate nodules, instead having an abundance of smectite, an alteration product of volcanic ash (Kirkland and Madsen 2007). This is due to an increase in volcanism to the west (Kirkland et al. 1999). It also differs by having local lignite deposits (Kirkland and Madsen 2007); coal is generally lacking in the Cedar Mountain Formation, unlike several succeeding formations. It has been proposed that it would make more sense to include the Mussentuchit Member in the Dakota Sandstone instead, but this would have the downside of making previous work on the Cretaceous geology and paleontology of Utah more confusing (Kirkland et al. 1999). The climate is interpreted as subtropical to tropical, with deposition near streams and rivers (Fiorillo 1999).

The fossil assemblage of the Mussentuchit Member was thought to be dramatically different from that of the Ruby Ranch Member (Kirkland et al. 1999), but that was before the reassignment of several quarries from the Ruby Ranch Member to the Mussentuchit Member. The fossils of the Mussentuchit Member are important for studies of paleobiogeography because they include the first appearances of several Asian taxa. They also include the first representatives of groups that would dominate North America during the Late Cretaceous (Kirkland and Madsen 2007). The sauropods of the Mussentuchit Member mark the last in North America before the "sauropod hiatus" that lasted until near the end of the Cretaceous (Maxwell and Cifelli 2000). The Mussentuchit Member is also known for the oldest North American snake fossils (Kirkland and Madsen 2007), its diverse armored dinosaurs (*Animantarx*, *Cedarpelta*, and *Peloroplites*) (Carpenter et al. 2008), the hadrosaur-like iguanodont *Eolambia*, and *Kokopellia*, the oldest and most basal marsupial (Kirkland and Madsen 2007). Plant fossils and microvertebrate materials (teeth and other small bones recovered through screenwashing) are particularly well-known from this member (Kirkland and Madsen 2007).

The floral assemblage of the Mussentuchit Member includes diatoms (algae with silica "shells") (Suarez et al. 2007), coal (Suarez et al. 2012), palynomorphs (including algal cysts, fern spores, conifer pollen, and angiosperm pollen) (Tschudy et al. 1984; Garrison et al. 2007), phytoliths (Suarez et al. 2007), the tree fern *Tempskya*, cycadeoids, conifer wood, platinoid flower impressions, wood from dicotyledonous angiosperms (Tidwell and Thayn 1985), and plant debris (Garrison et al. 2007). The plant discoveries mentioned earlier appear to be from the Mussentuchit Member, so they are included here. Invertebrate fossils known from the member include bivalves, gastropods, and invertebrate trace fossils (Garrison et al. 2007). Fish are diverse, including several types of chondrichthyans (hybodonts, carpet sharks, skates, and rays), several types of ray-finned fish (bowfins, gars, pycnodonts, and indeterminate ray-fins), and the lungfish *Ceratodus* (Cifelli et al. 1999). Finally, the tetrapod assemblage includes amphibians (albanerpetontids, anurans, and salamanders), turtles, diverse lizards (including teiids, monitor relatives, and possible skinks and helodermatids), snakes (Cifelli et al. 1999), several types of crocodylomorphs (Suarez et al. 2012), pterosaurs (Carpenter et al. 2008), small and large theropod dinosaurs (dromaeosaurids, ornithomimids, troodontids, tyrannosaurid relatives, and problematic taxa), brachiosaurid sauropods, some of the earliest fossils of pachycephalosaurid and ceratopsian dinosaurs in North America, hypsilophodont and iguanodont ornithopods, birds, diverse early mammals (including marsupials, multituberculates, symmetrodonts, tribotheres, and triconodonts) (Cifelli et al. 1999; Kirkland and Madsen 2007), dinosaur tracks (Garrison et al. 2007), and dinosaur eggshells (Fiorillo 1999).

Mesaverde Group: Tarantula Mesa Sandstone (Upper Cretaceous)

The Tarantula Mesa Sandstone of CARE is at the western extreme of the depositional area of the larger Mesaverde Group, a widespread geologic unit. This group is well-exposed at Chaco Culture National Historical Park (CHCU) and Mesa Verde National Park (MEVE), where its rocks record the waning Western Interior Seaway. At CARE, the Mesaverde Group includes is represented by the Tarantula Mesa Sandstone. This formation is 90 to 120 m (300 to 400 ft) thick, and is composed of brown to gray sandstone with interbedded shale and some

pebble conglomerates in the upper part of the formation (Morris et al. 2010). It is interpreted as having been deposited in barrier beach and deltaic settings (Morris et al. 2010) as the Western Interior Seaway continued the regression begun during the Muley Canyon and Masuk members of the Mancos Shale (Peterson et al. 1980). The contact with the underlying Masuk Formation is transitional in the CARE area (Smith et al. 1963). The Tarantula Mesa Sandstone of the CARE area is correlated to the Castlegate Sandstone of the Mesaverde Group to the northeast, and probably dates to the late early Campanian stage (Peterson et al. 1980), around 80 Ma. Formerly, this unit was known as the Mesaverde Formation in the CARE area, but Smith (1983) formally abandoned the old name and renamed this strata the Tarantula Mesa Sandstone (see also Peterson et al. 1980; Eaton 1990; not the same as the overlying "beds on Tarantula Mesa" [Smith 1983; Eaton 1990]).

While the Mesaverde Group as a whole is fossiliferous, the coastal margin sandstones of the Tarantula Mesa Sandstone are not known for their fossils, and the formation is not known to be fossiliferous at CARE. Root traces and general bioturbation have been reported from the lower part of the formation (Smith 1983), as well as burrows (Peterson et al. 1980). In general, noncoaly Mesaverde Group units, like the correlative Castlegate Sandstone, are sparsely fossiliferous. They may yield fossils including charophytes, gymnosperms, angiosperms, bivalves, gastropods, fish (such as gars), turtles, crocodilians, bone fragments and teeth of typical North American Late Cretaceous dinosaurs (tyrannosaurids, ceratopsids, and hadrosaurids), dinosaur eggshells, mammals (including multituberculates), and trace fossils (Young 1987). An even less extensive assemblage would be likely for the Tarantula Mesa Sandstone, given the depositional setting.

Park Collections

Cameron Cox (CARE archeological technician, pers. comm., March 2010) reported 40 catalog numbers for fossils in CARE collections. The records include 14 catalog numbers for vertebrate tracks (*Chirotherium*), five for bivalves, four for brachiopods, four for cycads, two for gastropods, and two for unknown fossils. Taxonomic descriptions for nine catalog numbers currently at the Superintendent's office were unavailable. The specimens were added over nine accessions. Accession 1, from 1950, came from Charles Camp, from collections in the Moenkopi and Chinle formations. Accession 4, from 1954, consists of cycad leaf fossils from a park visitor. Accession 7, from 1956, came from trackways of lacertoids, *Chirotherium*, and *Rotodactylus* from former superintendent Charles Kelly. Accession 16, from 1959, consists of a clam fossil from A. L. Inglesby of Fruita, Utah. Accession 18, from 1961, includes polished rock and petrified bone specimens from A. L. Ingelsby. Accession 40 predates 1978 and includes geological specimens. Accession 60, from 1986, consists of a fossil bone delivered by a park technician. Similarly, Accession 61, consisting of a mold of a fossil gastropod, was delivered by a park employee in 1987. Finally, Accession 178 was added in 2000, and consists of *Chirotherium* tracks from a former exhibit (C. Cox, pers. comm., March 2010).

At least two other institutions have CARE specimens in their collections. As discussed under the Moenkopi Formation section, the University of California Museum of Paleontology at Berkeley has a trackway from CARE in its collections, catalogued as UCMP 43751 (Peabody 1956). As mentioned in the Chinle Formation section, specimens of the bennettitale *Zamites powelli* were collected from Waterpocket Fold by R. C. Moore in 1921, and are now in the collections of the National Museum of Natural History (Washington, D.C.) (USNM 168987, 168988, 168989, and 168990) (Ash 1975). Specimens of the problematic plant *Sanmiguelia* were collected by Sidney Ash from the Chinle Formation of the park and are also now in USNM collections (two slabs catalogued as USNM 298826 and 298827) (Ash 1982).

Cultural Resource Contexts

Many archeological sites are present in CARE (Smith et al. 1963), so it would not be surprising to find paleontological resources associated with or as cultural resources. For example, charcoal and carbonized wood are present at some hearth sites in the CARE area (Smith et al. 1963). As noted earlier, nearby Cowboy Cave was occupied by humans over multiple intervals spanning approximately 9,000 years, with numerous artifacts made from the contemporaneous biota, providing a model for potential finds at CARE. Stokes (1972) interpreted three petroglyphs in southeastern Utah, including one from the eastern foothills of nearby Thousand Lake Mountain, as representing proboscideans (possibly mastodons), but such identifications are extremely controversial. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for CARE during September 1999 (NPS 1999a), and produced a geologic resource evaluation report for CARE in 2006 (Graham 2006). The participants of the scoping meetings developed a series of recommendations for future work including continuation of paleontological resource inventories, increased protection, staff training, and the completion of a paleontological resource management plan.

- The park should consider future field inventories for paleontological resources to more fully document *in situ* occurrences of fossils. The park may consider a formal site documentation and condition assessment for significant fossil localities. Vandalism is a particular issue of concern; Mathis (2004) mentioned vandalism of Morrison Formation fossil sites in the park. Monitoring of significant sites should be undertaken at least once a year in the future. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- Within the NCPN, the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and increase understanding of this important fossil-bearing unit. A similar USGS–NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Cedar Breaks National Monument

Cedar Breaks National Monument (CEBR), Utah, was established to protect Cedar Breaks, a vast and scenic natural amphitheater than spans approximately 4.8 km (3 miles) and is over 610 m (2,000 ft) deep (CEBR Figure 1). It was proclaimed August 22, 1933, and had boundary changes March 7, 1942 and June 30, 1961. CEBR currently encompasses 2,490.68 ha (6,154.60 acres), all of which are under federal administration (CEBR Figure 2).

Geologic Background

CEBR is located near the southeastern corner of Iron County in southwestern Utah. It is surrounded by a parcel of Dixie National Forest. Brian Head Peak is just to the north. Cedar City is 11 km (7 miles) west of CEBR; CEBR and Cedar City are connected by Cedar Canyon (sometimes called Coal Canyon) and its subsidiary canyons. Cedar City is also just west of the prominent Hurricane Cliffs. The St. George area is over 85 km (53 miles) to the southwest. Structurally, CEBR is within the western Markagunt Plateau. It is geologically similar to Bryce Canyon National Park (BRCA; see the BRCA summary in this report for more details), 48 km (30 miles) to the east. In particular, both parks owe much of their scenery to differential erosion of the Claron Formation (Hatfield et al. 2010).



CEBR Figure 1. The natural amphitheater Cedar Breaks, eroded through the Claron Formation (VLS).

The geology and paleontology of CEBR are discussed in a number of documents, including: Gregory (1950); Schneider (1967); Santarcangelo et al. (1975); Mullett et al. (1988); Moore and Nealey (1993); Sharpe (1993); Bown et al. (1997); Sable and Maldonado (1997); Anderson et al. (1999); National Park Service (NPS) (1999); Hatfield et al. (2004, 2010); Eaton et al. (2001); Koch and Santucci (2002); Moore et al. (2004); Thornberry-Ehrlich (2006); Santucci and Kirkland (2010); Spangler and Green (2010); Biek et al. (2011); and Rowley et al (in prep.). The Geologic Resources Division of the NPS coordinated a geologic resources inventory scoping session for CEBR during July 1999 (NPS 1999), and produced a geologic resources evaluation report for the monument in 2006 (Thornberry-Ehrlich 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The oldest rocks exposed at CEBR were deposited around 90 Ma (million years ago), during the Late Cretaceous, as a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway retreated from southwestern and south-central Utah (see the appendix for a geologic time scale). The Western Interior Seaway extended over the central United States and Canada during the Late Cretaceous, but it only encroached briefly on southwestern Utah. At its maximum extent it reached about as far west as Cedar City (Laurin and Sageman 2007: Biek et al. 2011). Instead of the marine rocks that dominate areas to the east, continental beds deposited by fluvial processes are abundant in the Cretaceous strata of CEBR (from oldest to youngest, the Straight Cliffs, Wahweap, and Grand Castle formations) (Hatfield et al. 2010; Biek et al. 2011). A lengthy mountain-building event known as the Sevier Orogeny was active to the west concurrently with the Cretaceous deposition, forming large thrust sheets of rocks (Hatfield et al. 2010) which eroded and contributed to the Cretaceous formations (Goldstrand and Mullett 1997; Eaton et al. 2001). As the Sevier Orogeny was nearing its end, the Laramide Orogeny was beginning. This mountain building event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). Basin and Range extension in the CEBR area began by around 20 Ma, but most Basin and Range faulting did not occur until about 10 Ma (Hatfield et al. 2010).

CEBR has a hiatus of uncertain duration between the Upper Cretaceous rocks and the overlying Claron Formation. Much of the exposed rock at CEBR belongs to the Paleocene–Eocene-age Claron Formation and the Eocene–Oligocene-age Brian Head Formation, a pair of fluvial and lacustrine formations (Hatfield et al. 2010). The Claron Formation was deposited as part of a depositional system that transitioned from predominantly fluvial to predominantly lacustrine (Davis and Pollock 2010), and is related to a series of lakes that formed in central and eastern Utah during the Paleogene (Davis et al. 2009). Erosion and weathering of the colorful Claron Formation

produced CEBR's outstanding scenery. Deposition of the Brian Head Formation occurred at the same time as the onset of volcanic activity to the north, in the Marysvale volcanic field (see below). Still younger rocks and sediments at CEBR include weathered remains of the Markagunt Megabreccia and a variety of Quaternary sediments (Hatfield et al. 2010; Biek et al. 2011).

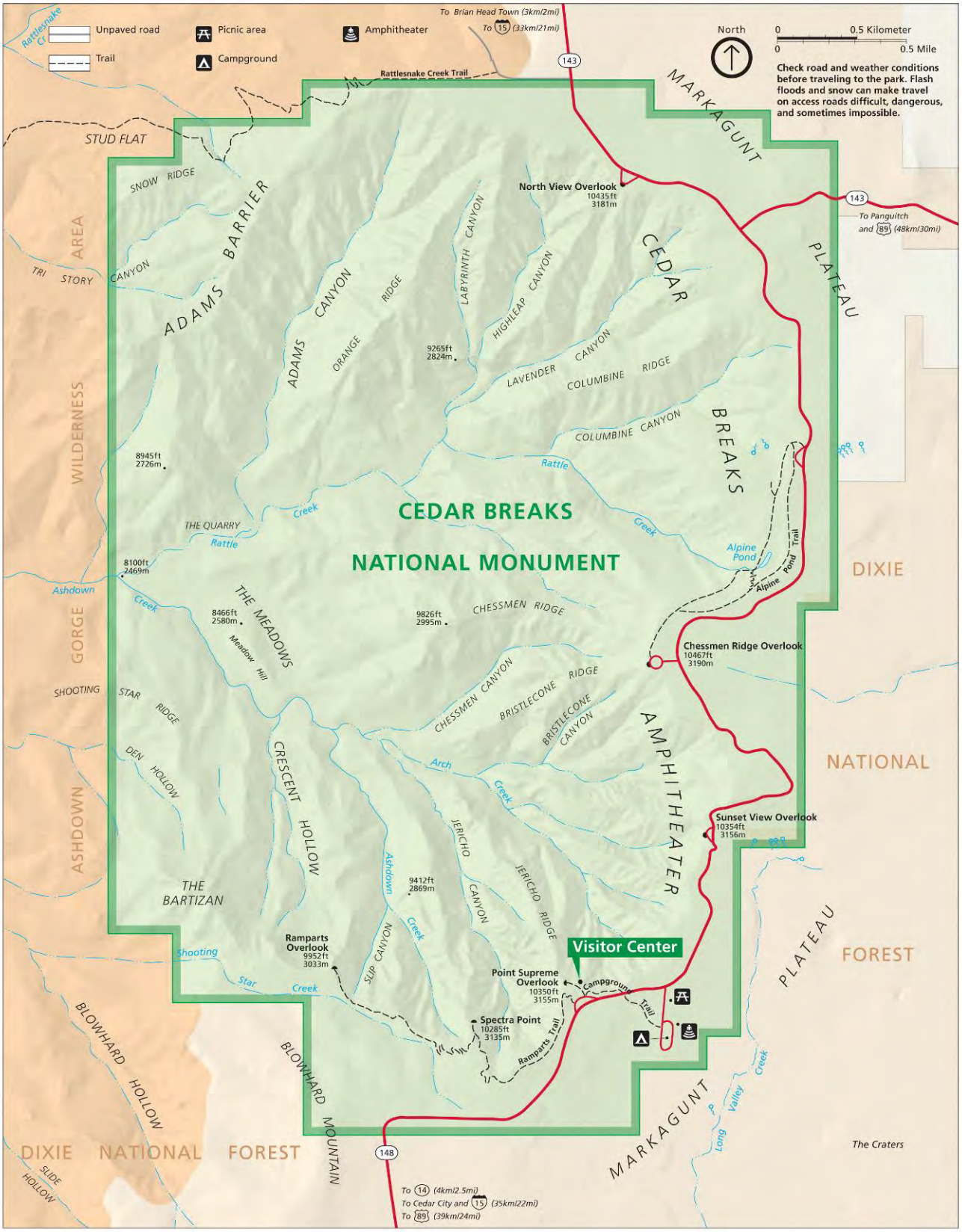
In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent widespread igneous activity during the late Eocene, Oligocene, and Miocene. The Cenozoic igneous activity of south-central Utah is dominated by the Marysvale volcanic field. Named for Marysvale, just over 100 km (about 60 miles) north-northeast of CEBR, Marysvale volcanics extend from Salina and the Fish Lake Plateau in the northeast to the Markagunt Plateau in the southwest (Rowley et al. 1975). Volcanism began in the Marysvale field by 36 Ma (represented in CEBR by the volcaniclastic [made of volcanic rock fragments] strata of the Brian Head Formation) and tailed off by around 14 Ma, but small-scale basaltic eruptions occurred as recently as the latest Pleistocene or possibly early Holocene (Rowley et al. 1994; Biek et al. 2011). Brian Head Peak, immediately north of CEBR, includes two ash-flow tuffs that erupted from calderas near the Utah–Nevada border. The lower ash-flow tuff is the Baldhills Tuff Member of the Isom Formation, erupted around 27 Ma. It is capped on Brian Head Peak by an ash-flow tuff from about 23.8 Ma (the Leach Canyon Formation) (Hatfield et al. 2010). More recently, during the Pliocene and Quaternary, small basaltic volcanic eruptions occurred on the on the Markagunt Plateau immediately east of CEBR, marking the northern edge of the Western Grand Canyon basaltic field (Johnson et al. 2010; Biek et al. 2011, in prep.). Many sinkholes and other karst features exist in these flows and adjacent plateau-top exposures of the Claron Formation, though actual caves are few (Spangler 2010; Spangler and Green 2010).

The Baldhills Tuff Member is found in CEBR as the local primary component of the Markagunt Megabreccia (Hatfield et al. 2010), or in debris derived from weathering of the megabreccia (Biek et al. 2011). Interpretation of the Markagunt Megabreccia is controversial. The latest research, based on information from earlier studies and new mapping of the Panguitch 30' x 60' quadrangle (which includes CEBR), suggests that the megabreccia is the debris of a gigantic gravity slide that resulted from catastrophic failure of the south flank of the Marysvale volcanic field about 20 Ma (see Biek et al. [2011] and references therein for a discussion of this enigmatic deposit). The trigger for the gravity slide is unknown, but may be related to inferred shallow igneous intrusions at the northeast end of the igneous complex known as the Iron Axis. To the southwest in and near the Pine Valley Mountains, Hacker et al. (2007) documented gravity slides that resulted from emplacement of shallow intrusions called laccoliths, which caused doming of the overlying rocks and subsequent failure on oversteepened slopes. With respect to CEBR, volcanic ash beds in the Brian Head Formation may have served as slip surfaces. Hatfield et al. (2010) interpreted the Markagunt Megabreccia as present at CEBR, but Biek et al. (2011) suggested that the Isom Formation boulders and blocks within CEBR simply represent debris eroded from the megabreccia.

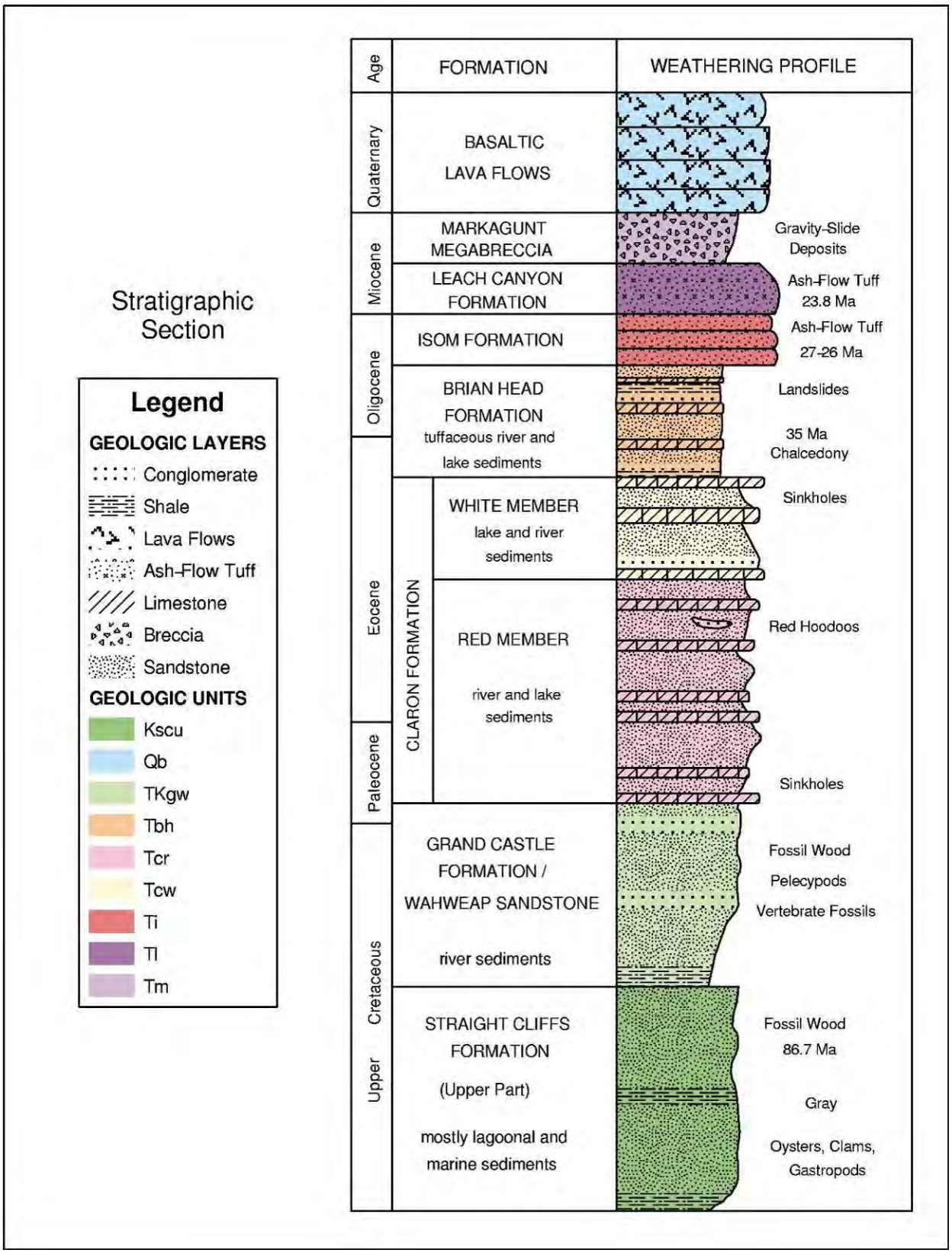
As the Pleistocene began approximately 2.6 Ma and continental glaciers formed over wide areas of the Northern Hemisphere, alpine glaciation occurred in Utah where elevations and precipitation were sufficient (Richmond et al. 1965). Two small glaciers formed during the Last Glacial Maximum (about 21,000 years ago) on the southeast side of Brian Head Peak (Mulvey et al. 1984; Madsen et al. 2002a; Biek et al. 2011). Deglaciation happened by over 14,400 years ago (Mulvey et al. 1984; Anderson et al. 1999). People were present in the Colorado Plateau by 12,000 years ago (Davis and Pollock 2010). Grand Staircase–Escalante National Monument to the southeast has evidence of the Clovis culture, which existed before 11,000 years ago (Agenbroad 1997).

Rock units exposed at CEBR include, from oldest to youngest, the Straight Cliffs Formation (Upper Cretaceous), Wahweap Formation (Upper Cretaceous), Grand Castle Formation (Upper Cretaceous–Paleocene?), Claron Formation (Paleocene–Middle Eocene), Brian Head Formation (Upper Eocene–Lower Oligocene), debris from the Miocene-age Markagunt Megabreccia, and various Quaternary-age sediments (Pleistocene–Holocene) (see CEBR Figure 3 for a stratigraphic column and CEBR Figure 4 for a geological map) (Moore and Nealey 1993; Moore et al. 2004; Thornberry-Ehrlich 2006; Hatfield et al. 2010; Biek et al. 2011). The Wahweap–Grand Castle interval has historically been problematic. Previously, the two formations were lumped as the informal "formation of Cedar Canyon" (Moore et al. 2004; Thornberry-Ehrlich 2006; Hatfield et al. 2010). It is now known that the strata beneath the Claron Formation at CEBR belong to the John Henry Member and Drip Tank members of the Straight Cliffs Formation (the latter of which is equivalent to the lower conglomerate member of the Grand Castle Formation), and the overlying Wahweap Formation and middle sandstone and upper conglomerate members of the Grand Castle Formation (Biek et al. 2011). The Claron Formation and Quaternary deposits are known to be fossiliferous within CEBR, and the Straight Cliffs Formation, Wahweap Formation, Grand Castle Formation, and Brian Head Formation are potentially fossiliferous within CEBR. Several of these formations have yielded fossils just outside of CEBR. The Markagunt Megabreccia is composed of faulted blocks of primarily volcanic rocks and has never been reported as fossiliferous (CEBR Table 1).

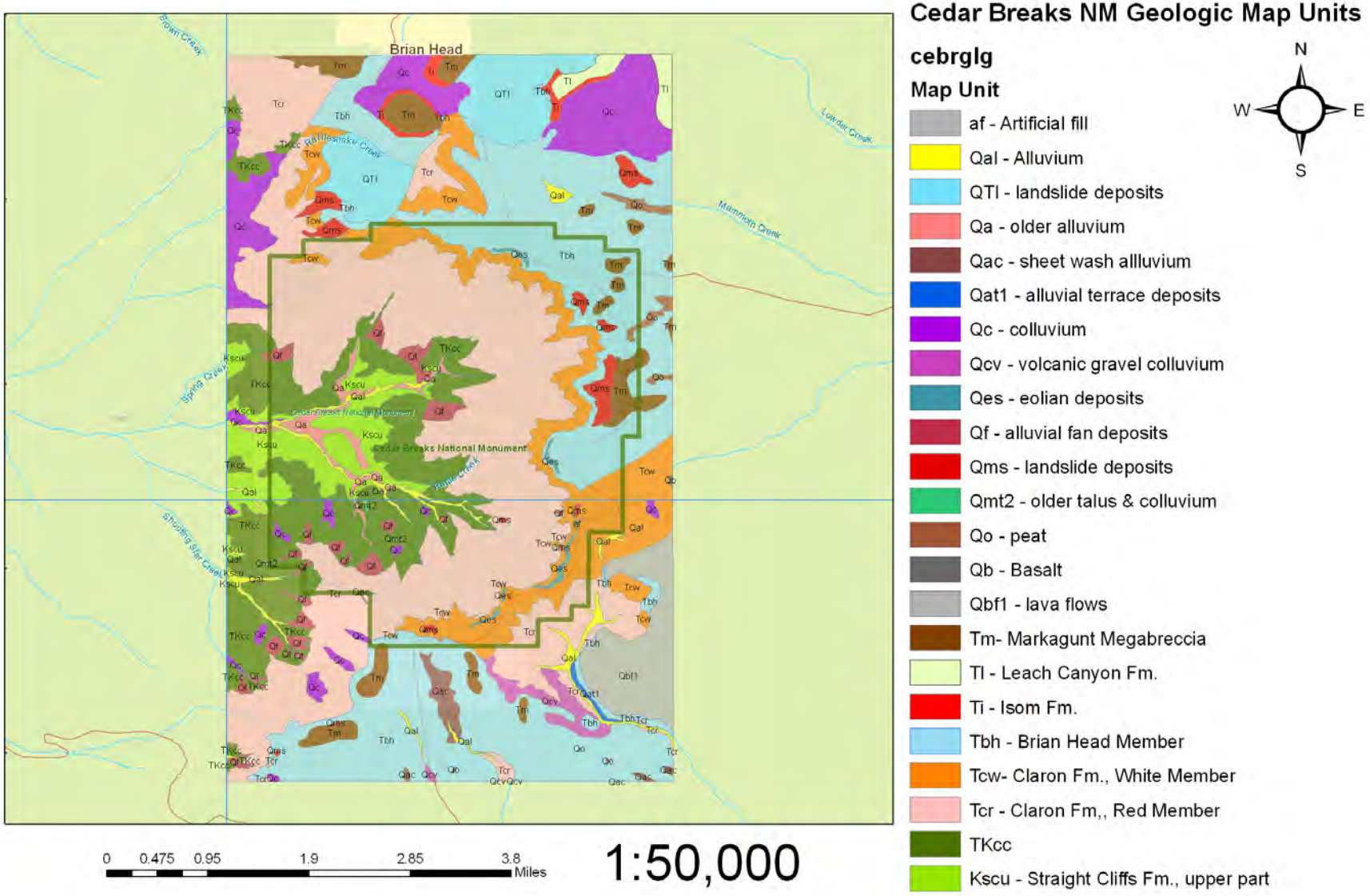
As with BRCA, the formational nomenclature for the rocks present at CEBR follows a confusing history of development (Biek et al. 2011) as described below. This can cause confusion when using older literature. Older usages are explained in the appropriate sections.



CEBR Figure 2. Geography of CEBR (NPS).



CEBR Figure 3. Stratigraphic column for CEBR (from Hatfield et al. [2010], used with permission of the Utah Geological Association).



CEBR Figure 4. Schematic geological map of CEBR. Recent mapping has revised some of the contacts and terminology, particularly for the pre-Claron Formation units. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1039076.

Formation	Age	Fossils Within CEBR	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Spores, pollen, conifer needles, wood, and mollusks	Alluvial, bog, colluvial, eolian settings, landslide, and talus deposits
Markagunt Megabreccia debris	Miocene	None to date; fossils are unlikely because the rocks are derived from a volcanic formation	Eroded gravity slide of volcanic rocks
Brian Head Formation	Late Eocene–Early Oligocene	None to date; root traces, freshwater invertebrates, and invertebrate traces are most likely	Fluvial and lacustrine settings, with significant input of volcanic debris
Claron Formation	Paleocene–Middle Eocene	Microbial structures, seeds, gastropods, invertebrate traces, and fish fragments	Fluvial, alluvial, and lacustrine settings
Grand Castle Formation	Late Cretaceous–Paleocene?	None to date; plant fossils are most likely	Fluvial settings
Wahweap Formation	Late Cretaceous	None to date; plants, freshwater invertebrates, and microvertebrate fossils (teeth, small bones, and other fossils recovered by screenwashing) are most likely	Fluvial settings
Straight Cliffs Formation	Late Cretaceous	None to date; spores, pollen, plant fragments, bivalves, gastropods, and microvertebrate fossils are most likely	Fluvial settings

CEBR Table 1. Concise tabulation of CEBR stratigraphy, fossils, and depositional setting; details and references can be found in the text. Debris of the Markagunt Megabreccia is included here for completeness, but is not discussed in detail in the text because of the improbability of fossils.

Sources of Paleontological Resources

The fossils of CEBR present opportunities for education, interpretation, and continued or future scientific research in the monument.

Claron Formation (Paleocene–Middle Eocene)

The Claron Formation of southwestern Utah has a complicated nomenclatural history involving several informal formation names and differing conceptions of what these formations encompassed. The Claron Formation was originally identified as the Wasatch Formation, better known from northern Utah and Wyoming. This is particularly apparent in the works of geologist Herbert Gregory. Additionally, in some of his publications (Gregory 1945, 1949) the unit now known as the upper white limestone member of the Claron Formation was included in the Brian Head Formation (see below). The use of the Wasatch Formation name for these rocks persisted into the 1970s (Bowers 1972). The modern lower (sometimes called pink or red) member of the Claron Formation was informally called the Cedar Breaks Formation from exposures at CEBR (Schneider 1967), which did not attain wide usage. Anderson and Rowley (1975) returned the upper white limestone to the Claron Formation and included all of Gregory's original Brian Head Formation rocks in the Claron Formation. The situation increased in complexity during the 1990s. First, a third formation (the Grand Castle Formation, also discussed below) was named from former lower Claron Formation rocks (Goldstrand 1994; Goldstrand and Mullett 1997). Then, the Brian Head Formation was brought back into usage, limited to rocks above the Claron Formation's upper white limestone member (Anderson 1993; Sable and Maldonado 1997). While it is necessary for the researcher wishing to examine the literature on the Claron Formation to keep in mind the dramatically shifting nomenclature, fortunately the rocks themselves are usually simple to identify in publications from their descriptions: the modern lower Claron Formation is almost always identified as a distinctive pink or red unit, and the modern upper Claron Formation is almost always identified as a white limestone. Biek et al. (2011, in prep.) discussed the Claron Formation on the Markagunt and Paunsaugunt plateaus and noted that the formation is restricted to non-volcanic strata bounded two unconformities: a major Late Cretaceous–early Cenozoic unconformity at the base of the formation, and the unconformity associated with the conglomerate of Boat Mesa at the top.

The Claron Formation of CEBR and the vicinity includes a lower red member and upper white member; Biek et al. (2011) further subdivided the white member into four units. The red member is composed of alternating beds of red to pink sandy limestone, calcareous (lime-rich) mudstone, siltstone, and

sandstone, with pebble conglomerate as a minor component. The white member is known for its two white limestone cliffs, but also includes calcareous mudstone, siltstone, and sandstone similar to that of the lower red member (Moore et al. 2004; Biek et al. 2011; Rowley et al. in prep.). The red member is about 300 m (1,000 ft) thick in CEBR, and the white member is about 135 m (440 ft) thick, making the thickness of the entire formation about 435 m (1,440 ft) in CEBR (Biek et al. 2011). Erosion of the white member cuts downsection to the south (Moore et al. 2004). The contact with the overlying Brian Head Formation, which corresponds to the top of a thin pebble conglomerate (possibly correlative to the conglomerate of Boat Mesa), is a disconformity (Hatfield et al. 2010; Biek et al. 2011). The Claron Formation is exposed through much of CEBR (Hatfield et al. 2010). The Pink Cliffs of CEBR (Moore et al. 2004) and much of the monument's scenic beauty are due to the erosion of the Claron Formation (Hatfield et al. 2010).

Historically, the Claron Formation was regarded as largely lacustrine, but fluvial processes are now more widely recognized. The lower member is mostly fluvial and alluvial in depositional environment, whereas the upper white member represents both lacustrine and low-energy fluvial environments (Goldstrand 1992; Bown et al. 1997; Ott 1999; Davis et al. [2009] erroneously regarded both members as shallow lacustrine). The lake transitioned from closed to open drainage around 42 Ma (Davis et al. 2009). Laramide-age uplifts to the northeast and northwest contributed sediment to the formation (Goldstrand 1992). The Claron Formation has been heavily affected by pedogenic (soil-forming) processes, which are thought to have destroyed most of its original depositional features and fossils (Mullett et al. 1988; Mullett 1989; Feist et al. 1997). Plant growth is interpreted as moderate, with moist soils that underwent periodic drying (Bown et al. 1997). The reddish coloration of the lower member is from iron staining, while the white of the upper member is thought to have resulted from leaching (Mullett 1989). Dissolution of Claron Formation limestone can produce sinkholes, which are abundant on the Markagunt Plateau (Moore et al. 2004), and shallow caves (Hatfield et al. 2010; Spangler 2010).

The age of the Claron Formation has been difficult to establish, because of its paucity of fossils, lack of volcanic debris, and the fact that it is bounded by unconformities; it is now widely considered to be Paleocene to latest middle Eocene in age (Goldstrand 1994; Eaton et al. 2001; Biek et al. 2011, in prep.). The recent discovery of a rodent jaw in the middle of the upper (white) member indicates that it includes rocks of middle Eocene age (Eaton et al. 2011). The upper age limit of the Claron Formation is also constrained by overlying volcaniclastic strata of the Brian Head Formation, which have been dated to 36 Ma (Biek et al. 2011). Older reports of Oligocene-age rocks in the Claron Formation (Bowers 1990) are based on rocks now included in the Brian Head Formation.

The Claron Formation is sparsely fossiliferous at CEBR. The lower member has uncommon microbial structures, rare seeds (possibly of hackberry or the genus *Prunus*, which includes plants that produce almonds, cherries, peaches, and plums), and possible invertebrate burrows (Schneider 1967), and beds near the top of the formation in Jericho Canyon have yielded fossils of the gastropod *Helix spatiosa* and the trace fossil *Celliforma*, interpreted as a bee larval cell (Gregory 1950). Fossils from unspecified members include freshwater gastropods and scraps of fish (Santucci and Kirkland 2010).

Fossils have historically been difficult to find in the Claron Formation, because of pedogenic and diagenetic processes. The most abundant fossils, including areas outside of CEBR, are microbial structures, mollusks, and arthropod traces. Microbial body fossils and traces from the Claron Formation include oncolites, stromatolites (Taylor 1993), and charophytes (Moore and Nealey 1993; Moore et al. 2004). These fossils are mostly found in the upper member (Moore and Nealey 1993; Moore et al. 2004). Plant fossils include palynomorphs (organic microfossils such as pollen and spores) (Goldstrand and Mullett 1997; Foster et al. 2001), leaf impressions (Gregory 1950), seeds (Schneider 1967), and root traces (Mullett et al. 1988; Mullett 1989; Taylor 1993). Invertebrate body fossils include small bivalved animals that could be bivalves, conchostracans (clam shrimp), or ostracodes (seed shrimp) (Moore and Nealey 1993), bivalves, gastropods (Gregory 1950), and ostracodes (Brouwers et al. 2000; Eaton et al. 2011). Shelled invertebrates may be found as molds or casts (Sable and Hereford 2004). Invertebrate trace fossils include ant nests, wasp and bee cocoons and brood cells with possible parasite traces, dung beetle nests (Bown et al. 1997), and crayfish burrows (Hasiotis and Bown 1997). The insect traces have been found in the clastic rocks of the lower member (Bown et al. 1997). Vertebrates are represented by fish bones and a fragmentary rodent jaw (Eaton et al. 2011).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary deposits of CEBR include a variety of alluvial, bog, colluvial, eolian, landslide, and talus deposits of widely-ranging thickness and grain size (Moore and Nealey 1993; Moore et al. 2004; Hatfield et al. 2010; Biek et al. 2011; Rowley et al. in prep.). Quaternary deposits of the area are up to about 17 m (55 ft) thick (Moore et al. 2004). Deposits are found scattered throughout the monument (Hatfield et al. 2010). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Summaries of Quaternary fossil sites in southern Utah and the vicinity include Mulvey et al. (1984), Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (1999, 2000), Madsen et al. (2002), and Coats et al. (2008).

The Quaternary deposits of CEBR are locally fossiliferous. CEBR Quaternary fossils to date are mostly from the Holocene and fall into the paleoecological category. The best-known fossils are from Alpine Pond near the eastern border of the monument. Its fossil record goes back approximately 3,000 years and includes pollen, spores, conifer needles, and spruce wood (Anderson et al. 1999). Similar fossils are known from the more thoroughly described Lowder Creek Bog, 2.8 km (1.7 miles) northeast of CEBR (Mulvey et al. 1984; Anderson et al. 1999; Madsen et al. 2002a, 2002b). Lowder Creek Bog is associated with a small former alpine glacier on Lowder Creek (Mulvey et al. 1984; Anderson et al. 1999). Its fossil record goes back approximately 17,000 years (Madsen et al. 2002a). Fossils from Lowder Creek Bog include peat, plant debris, wood fragments (Madsen et al. 2002a), pollen, spores, conifer needles and other plant macrofossils, zooplankton, and hard parts of aquatic arthropods (chironomid midges, beetles, and mites) (Anderson et al. 1999). Another well-studied paleoecological site is the drained Red Valley bog, 10.1 km (6.3 miles) east of CEBR. It too has a fossil record extending back to the latest Pleistocene, approximately 11,795 ± 90 radiocarbon years before present at the base of the deposit (13,840–13,420 calendar years before present per Calib 6.0 [http://intcal.qub.ac.uk/calib/calib.html]; note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years). The record here is not continuous to the present, though, as there appears to be a gap between the oldest levels and younger mid-Holocene material. Fossil material from this site ranges from peat and charcoal, to pollen and spores, to plant macrofossils such as seeds, cone scales, needles, and branches, to remains of 33 taxa of insects and arachnids. Most of the arthropods are beetles, but ants, shield bugs, midges, caddisflies, moths, and mites are also present (Madsen et al. 2002b).

Aside from Alpine Pond, CEBR also has a nearly continuous record of Holocene mollusks (Sharpe 1993), and Pleistocene and Holocene bog deposits with peat (Hatfield et al. 2010), indicating further possibilities to find sites like Alpine Pond and Lowder Creek Bog. Additionally, limestones of the basal Claron Formation are known to dissolve into shallow caves (Hatfield et al. 2010), which are potential locations for packrat middens. However, Madsen et al. (2002b) considered the rock type present at CEBR to be too friable to preserve middens for long periods of time. They surveyed a number of potential midden sites on the Markagunt Plateau, but found few middens, and those they did find they estimated to be generally hundreds of years old, at most a few thousand years old. In general, they found the region unsuitable for midden preservation.

Comparable paleoecological fossils are also known from ZION (see the ZION summary in this report for more information). At ZION, sediments deposited in Pleistocene and Holocene lakes preserve a variety of fossils. Among them are diatoms, pollen, spores, stumps, impressions of plants, other plant fossils, bivalves, gastropods, invertebrate trace fossils, a few fish and mammal bones, and tracks of a large bird and a mammal (Hamilton 1979; Hevly 1979).

Quaternary mammal fossils are not yet known from CEBR. Fossils of large Quaternary mammals on the Colorado Plateau tend to be found in the vicinity of rivers, as shown by the pattern of localities in Agenbroad (1997). The most commonly found large mammals are sloths, mammoths, horses, tapirs, camels, bison, oxen, and mountain goats (Agenbroad 1997). The nearest find of an extinct Pleistocene mammal consists of mammoth bones found 67 km (42 miles) west of CEBR, near Newcastle (Larson 1999).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within CEBR. However, they are known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from one or more of them.

Straight Cliffs Formation (Upper Cretaceous)

The Straight Cliffs Formation is a heterogeneous unit deposited in multiple paleoenvironmental settings. New mapping by Biek et al. (2011) shows that it outcrops only in the lowest elevations of CEBR along Ashdown Creek and its tributaries, and thus has a considerably more limited extent in the park than envisioned by Thornberry-Ehrlich (2006) and Hatfield et al. (2010). Rocks from this formation weather into the Gray Cliffs of southern Utah (Davis and Pollock 2010). The Straight Cliffs Formation is divided into four members, based on exposures from the Kaiparowits Plateau. From oldest to youngest, these are the Tibbet Canyon Member, the Smoky Hollow Member, the John Henry Member, and the Drip Tank Member (Peterson 1969a).

These four members are present along the west flank of the Markagunt Plateau, but only the Drip Tank Member and the upper few meters (tens of feet) of the John Henry Member are present in CEBR (Biek et al 2011). As originally inferred by Moore and Straub (2001), Lawton et al. (2003), and Eaton (2006), mapping by Biek et al. (2011) showed that the Drip Tank Member of the Straight Cliffs Formation and the lower conglomerate member of the Grand Castle Formation are the same unit. The following information is mostly specific to the latter two members. For details on all four members, treated individually, see the Straight Cliffs Formation section of the BRCA summary in this report.

The Straight Cliffs Formation of CEBR is regarded as having been deposited by fluvial processes (Moore and Straub 2001; Moore et al. 2004; Hatfield et al. 2010). The Drip Tank Member and equivalent rocks in the Grand Castle Formation were deposited by braided streams that tapped the Sevier highlands to

the west (Goldstrand and Mullett 1997). The age of the Straight Cliffs Formation at CEBR is constrained by a date in the upper part of the formation in Cedar Canyon of 86.72 ± 0.58 Ma (Eaton et al. 2001), and by bivalves found in the lower Straight Cliffs Formation east of Cedar City (Cobban et al. 1996) that are part of a faunal zone (*Mytiloides puebloensis* zone) dating to sometime between roughly 93.32 and 93.19 Ma (Cobban et al. 2006).

Regionally, the fossil assemblage of the upper members of the Straight Cliffs Formation includes coal (Peterson 1969a), palynomorphs (McCord 1997), sparse wood fragments and plant impressions (Sable and Hereford 2004), petrified logs (Peterson 1969a), bivalves, cephalopods (Peterson 1969a), gastropods (Gallin et al. 2010), crabs (Eaton and Cifelli 1988), invertebrate burrows (Peterson 1969a), sharks, rays, ray-finned fish including bowfin and gars, the unusual amphibian *Albanerpeton*, frogs, turtles, lizards, several types of crocodilians and crocodile relatives, dromaeosaurids ("raptor" dinosaurs), armored dinosaurs, hadrosaurids ("duckbill" dinosaurs) (Eaton et al. 1999a), mammals (including multituberculates, symmetrodonts, and marsupials) (Eaton 2006a), and dinosaur tracks (Titus et al. 2005). Fossils are much better represented in the John Henry Member than the overlying Drip Tank Member.

The upper Straight Cliffs Formation of the Markagunt Plateau is fossiliferous, but fossils are not yet documented from the formation within CEBR. A variety of palynomorphs have been recovered from these rocks on the plateau, including freshwater algal cysts, moss and fern spores, and gymnosperm and angiosperm pollen (Nichols 1997). Other fossils in the Straight Cliffs Formation in the vicinity of CEBR include plant fragments, leaf impressions, petrified wood, bivalves, gastropods (Moore and Nealey 1993; Moore et al. 2004), and a variety of vertebrates (Eaton et al. 1999b; Eaton 2006b). Vertebrates have been described from two sites in the Straight Cliffs Formation of Cedar Canyon, just south and west of CEBR. Vertebrates found at these sites include sharks, guitarfish, bony fish, turtles, lizards, crocodilians or crocodile relatives, dinosaurs, multituberculates, marsupials (Eaton et al. 1999b), and possible eutherian mammals (Eaton 2006b). The rocks may correlate to the John Henry Member (Eaton et al. 1999b; Eaton 2006b). Additionally, a partial skeleton of a theropod (bipedal carnivorous) dinosaur was collected from the formation south of CEBR, by David Archibald (San Diego University) (Santucci and Kirkland 2010).

Wahweap Formation (Upper Cretaceous)

The Wahweap Formation is a heterogeneous formation deposited on alluvial plains by fluvial and floodplain processes (Sable and Hereford 2004). In the past, it was mapped with rocks assigned to the Grand Castle Formation at CEBR (Thornberry-Ehrlich 2006; Hatfield et al. 2010), but new mapping clarifies the relationship of these two units (Biek et al. 2011). The Wahweap Formation of CEBR consists of soft variegated mudstone interbedded with lesser quantities of variegated sandstone and siltstone, becoming more sandstone-rich higher in the section. The Wahweap Formation is about 245 m (800 feet) thick at CEBR; Moore and Straub (2001) measured 230 m (760 feet) of strata that Biek et al. (2011) assigned to the Wahweap Formation. The formation thins dramatically northward into Parowan Canyon, where just a few tens of feet of Wahweap-like strata are present at the base of the middle sandstone member of the Grand Castle Formation (Biek et al. 2011). Rocks identified previously as the Kaiparowits Formation on the Markagunt Plateau are actually the Wahweap Formation (Roberts 2007; Biek et al. 2011).

On the Kaiparowits Plateau to the east, where the Wahweap Formation was first described and is better exposed, it is divided into four informal members: the lower, middle, upper, and capping sandstone members. The three members below the capping sandstone are generally composed of sandstone and mudstone, and are interpreted as having been deposited by meandering streams flowing to the north or northeast. The capping sandstone also includes conglomeratic and siltstone beds, and is thought to represent braided stream deposition, with flow to the east and southeast (Lawton et al. 2003; Titus et al. 2005). The capping sandstone member of the Wahweap Formation is equivalent to the middle sandstone member of the Grand Castle Formation, described below (Biek et al. 2011). Fossils are abundant in all four members (Titus et al. 2005). Unlike underlying formations, the Wahweap Formation lacks coal (McCord 1997).

The Wahweap Formation is regarded as entirely nonmarine in deposition (Eaton 1991; McCord 1997), the Western Interior Seaway having previously retreated out of south-central Utah. The climate appears to have become more arid toward the end of deposition, with possibly semi-arid to arid conditions (Simpson et al. 2008, 2010a). Sediment deposition was probably not continuous, with the capping sandstone member in particular differing from the rest of the formation. This member has disconformable contacts at its base and top (Kirkland and DeBlieux 2010). Deposition occurred after 81 Ma and before 76 Ma (Jinnah et al. 2009; Getty et al. 2010; Simpson et al. 2010b).

The Wahweap Formation is not known to be fossiliferous within CEBR, but it is a good candidate to eventually yield fossils within the monument. It has yielded carbonized wood and leaf impressions in the area (Moore and Nealey 1993; Hatfield et al. 2010), and vertebrates are known in these rocks from Cedar Canyon to the west (Eaton et al. 1999b, Eaton et al. 2001; Eaton 2006b), including cartilaginous fish, bony fish, turtles, lizards, dinosaurs (tyrannosaurids, dromaeosaurids, armored dinosaurs, possible horned dinosaurs, and hadrosaurs), and mammals (multituberculates, symmetrodonts, marsupials, and other early forms) (Eaton et al. 1999b).

Overall, the paleontological fossil resources of the Wahweap Formation are diverse. Fossils are best known in the formation

in the BRCA and Grand Staircase–Escalante National Monument areas to the east. Microvertebrate remains are well-known from this formation. Bacteria, algae, and plants are represented by charophytes (Peterson 1969b), cryptobiotic soils (Simpson et al. 2010a), palynomorphs (Lawton et al. 2003), leaf impressions (Moore and Nealey 1993), compressed plant fossils (Kirkland et al. 2005), petrified wood (Titus et al. 2005), carbonized wood (Moore and Nealey 1993), and root traces (Simpson et al. 2008). Possible cycads, conifers, and angiosperms are among the plants that have been found (Erwin 1997). Petrified wood is common in channel-fill deposits of the various members, with specimens the size of large logs in some cases (Titus et al. 2005). Additionally, reworked palynomorphs from older formations have been found in the capping sandstone (Lawton et al. 2003). Invertebrates are represented by bivalves, gastropods, ostracodes (Peterson 1969b), crab claws (Eaton 1991), and invertebrate wood borings, which have been found in the upper and capping members (Moran et al. 2010). Mollusks sometimes form shell beds, which are most abundant in the middle mudstone-rich member (Kirkland et al. 2005).

Vertebrates of the Wahweap Formation are represented by freshwater sharks and rays, gars, bowfins, and other bony fish, salamanders, frogs, turtles, lizards, crocodilians and crocodile relatives, a variety of dinosaurs, diverse early mammals (Eaton et al. 1999a, 1999b; Titus et al. 2005; Kirkland and DeBlieux 2010), probable lungfish burrows (Simpson et al. 2008), dinosaur eggshell (Eaton 2004), crocodilian or crocodile relative tracks (Simpson et al. 2010c), dinosaur tracks (Simpson et al. 2010d), and theropod digging traces associated with mammal burrows (Simpson et al. 2010b). Many of the trace fossil occurrences are known from the capping sandstone (Simpson et al. 2010b, 2010c, 2010d). The most abundant dinosaurs are horned dinosaurs and hadrosaurs (Titus et al. 2005), with tyrannosaurids, dromaeosaurids, troodontids, armored dinosaurs, pachycephalosaurs, and hypsilophodonts as well (Eaton et al. 1999a; Weishampel et al. 2004; Titus et al. 2005). Dinosaur bonebeds are present but rare (Kirkland et al. 2005). The first unique dinosaur taxon from the Wahweap Formation was named in 2010: *Diabloceratops eatoni*, an early horned dinosaur (Kirkland and DeBlieux 2010). Mammals have been extensively described, and include multituberculates, symmetrodonts, marsupials, insectivores, and other early forms (Eaton and Cifelli 1988; Cifelli 1990; Eaton et al. 1999a, 1999b; Titus et al. 2005; Eaton 2006b; Kirkland and DeBlieux 2010). Generally, microvertebrate remains are more common in sandstone channel lags, but there are significant microvertebrate sites in mudstone beds. The lack of articulated remains is attributed to the rarity of overbank deposits (Kirkland et al. 2005).

Grand Castle Formation (Upper Cretaceous–Paleocene?)

As noted above, the Grand Castle Formation was named from rocks that had previously been assigned to the lower Claron Formation (Goldstrand and Mullett 1997; Goldstrand 1994). The formation consists of a distinctive, slope-forming, white quartz-rich sandstone member bounded above and below by cliff-forming conglomerate members. It is another fluvial unit (Goldstrand and Mullett 1997). The sandstone member was deposited by sandy braided streams, and the lower and upper conglomerates by gravelly braided rivers that flowed to the southeast (Goldstrand 1994). Sediment came from Sevier thrust sheets in southwestern Utah (Goldstrand 1992, 1994; Goldstrand and Mullett 1997; Lawton et al 2003).

Previously, the age of the formation was poorly constrained as Paleocene (Goldstrand and Mullett 1997). More recent discoveries indicate a Late Cretaceous age for a substantial part of the formation. A dinosaur footprint has been reported from the middle sandstone member, confirming a Late Cretaceous age for at least the lower two members (Hunt et al. 2011). Furthermore, recent mapping demonstrated the equivalence of the capping sandstone member of the Wahweap Formation (known to be Late Cretaceous in age) and the middle sandstone member of the Grand Castle Formation, as originally inferred by Goldstrand and Mullett (1997) and Lawton et al. (2003) and supported by preliminary detrital zircon analyses of Johnson et al. (2011). The age of the upper conglomerate member of the Grand Castle Formation remains uncertain; palynomorph evidence suggests that it may be Late Cretaceous (Nichols 1997; Biek et al. 2011).

The Grand Castle Formation is not known to be fossiliferous within CEBR, and is not extensively fossiliferous. Its primary fossil assemblage, in the middle sandstone member, includes pollen (Nichols 1997; Biek et al. 2011), plant fragments, petrified wood, log casts, leaf impressions, root traces, burrows (Goldstrand and Mullett 1997), and a dinosaur footprint (Hunt et al. 2011). The conglomeratic members of the formation also includes reworked Paleozoic rocks with fusulinid foraminifera (a large extinct type of protist that secreted a "shell"), sponges, rugose corals, bryozoans (moss animals), brachiopods (lamp shells), and crinoids (sea lilies) (Goldstrand 1992).

Brian Head Formation (Upper Eocene–Lower Oligocene)

The Brian Head Formation is a light-gray volcaniclastic sedimentary unit, about 150 m (500 feet) thick at its type section at Brian Head peak immediately north of CEBR. At CEBR, the Brian Head Formation is preserved only in the northeast park of the park (Biek et al. 2011) and is thus more restricted than shown by Thornberry-Ehrlich (2006) and Hatfield et al. (2010), who followed Sable and Maldonado's (1997) now outdated definition of the formation. At CEBR, Brian Head Formation strata are restricted to the middle unit of Sable and Maldonado (1997), which is predominantly grayish volcaniclastic sandstone and mudstone, with some limestone and airfall tuff. The upper unit of Sable and Maldonado (1997), composed of volcanic lava flows and mudflows, is not present at CEBR (Hatfield et al. 2010).

The Brian Head Formation is late Eocene to early Oligocene in age based on several isotopic dates (Sable and Maldonado 1997; Biek et al. 2011). Mammal fossils from rocks in the lower part of the formation on the Sevier Plateau to the northeast are of the Duchesnean North American Land Mammal Age (NALMA) (Eaton et al. 1999c); the Duchesnean NALMA dates to between approximately 40 and 37 Ma (Robinson et al. 2004). The Brian Head Formation is a fluvial and lacustrine unit, with significant volcanic input. The volcanics incorporated into the formation mark the first major Cenozoic volcanism in the area (Sable and Maldonado 1997).

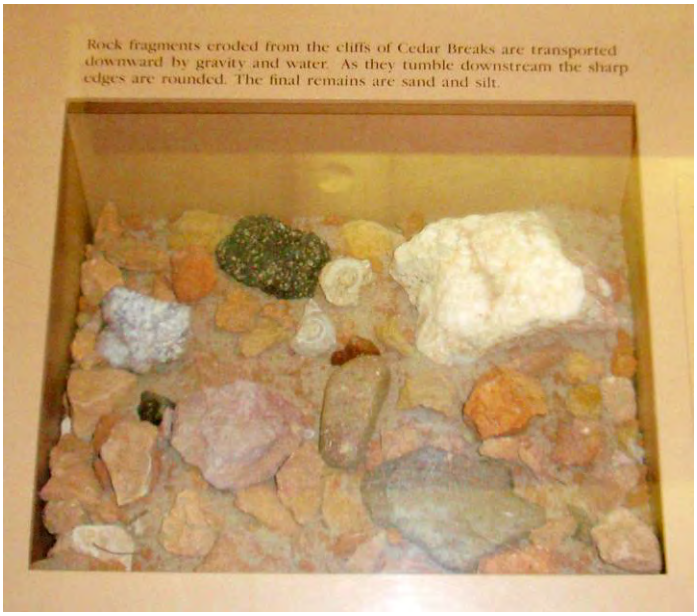
The Brian Head Formation is not widely fossiliferous within CEBR, although there are some reports of tubular sedimentary structures that may be root casts or burrows (Bowers 1972). On the Markagunt Plateau, the fossil assemblage for the Brian Head Formation include general bioturbation of sediment (disruption of sedimentary features by the actions of organisms), root casts (Sable and Maldonado 1997), bivalves, gastropods (Gregory 1949; Sable and Maldonado 1997), and possible worm burrows (Sable and Maldonado 1997). A greater fossil assemblage is known from the Brian Head Formation on the Sevier Plateau. There, rocks of the variegated unit (the lower part of Sable and Maldonado's [1997] middle unit) have yielded charophytes, gastropods, ostracodes, rays, bowfins, gars, catfish, other ray-finned fish, turtles, alligators, birds, marsupials, rodents, carnivorans, brontotheres, possible tapir relatives, and anthracotheres. These probably represent an assemblage from a lake setting (Eaton et al. 1999c). Higher rocks in the section have yielded bioturbation (Golder and Wizevich 2009), charophytes (Eaton et al. 1999c), root traces (Golder and Wizevich 2009), bivalves, gastropods (Feist et al. 1997), multiple types of invertebrate traces including crayfish burrows and insect or larval traces (Golder and Wizevich 2009), pellets (Golder et al. 2009), and vertebrates (Eaton et al. 1999c).

Park Collections

CEBR has a small collection of clasts eroded from Cedar Breaks on display, which includes two fossil gastropods (CEBR Figure 5). The work on pollen and other microfossils from Alpine Pond was performed at Northern Arizona University in Flagstaff (Anderson et al. 1999).

Cultural Resource Contexts

Humans have been present in south-central Utah for at least 11,000 years (Agenbroad 1997), so it would not be surprising to find paleontological resources at CEBR associated with or as cultural resources, such as fossils taken for cultural purposes or used as tools. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.



CEBR Figure 5. Among this collection of eroded rocks are two gastropods (upper center) (VLS).

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for CEBR during July 1999 (NPS 1999), and produced a geologic resource evaluation report for the monument in 2006 (Thornberry-Ehrlich 2006). The participants of the scoping meetings developed a series of recommendations for future work, including: performing a comprehensive inventory of CEBR paleontological resources; inventorying all paleontological specimens collected in the monument; determining and assessing fossil localities for GPRA goal Ia9; interpreting CEBR's fossil resources for visitors; developing paleontological resource monitoring strategies; and investigating caves, sinkholes, and springs in the Claron Formation. A Geologic Resource Monitoring Manual by the

Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- There should be a coordinated inventory of paleontological resources in the Claron and surrounding formations in CEBR to supplement the work done in adjacent areas and extrapolated to the monument.

- Should CEBR be expanded to the west, as has been discussed, it would incorporate greatly-expanded exposures of the Grand Castle, Wahweap and Straight Cliffs formations. With the extensive fossil resources of these formations, paleontological resources should be given greater emphasis in monument management and scientific investigations.

- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Data Sets

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Cedar Breaks National Monument Collections
1933–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by CEBR staff; status: Active.

Cedar Breaks National Monument Digital Geologic Map
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Northern Arizona University
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Colorado National Monument

Colorado National Monument (COLM), Colorado, was established for its outstanding scenic value (COLM Figure 1). Its deep rocky canyonlands host diverse paleontological and archeological resources as well. It was proclaimed May 24, 1911. COLM has undergone several boundary changes, on March 3, 1933; August 5, 1959; October 21, 1976; and November 10, 1978. It currently encompasses 8,309.79 ha (20,533.93 acres), all of which are under federal administration (COLM Figure 2).

Geologic Background

COLM is located in northwestern Mesa County, western Colorado, near the midpoint of the border with Utah. It is on the northeast side of the Uncompahgre Plateau, near the northern tip of this structure. The Colorado River is joined by the Gunnison River a short distance east of COLM, and the Colorado-Gunnison drainage system roughly parallels the monument's eastern and northeastern boundaries. The land along the northeastern border is known as the Redlands, for the color of the rocks. A series of canyons and mesas makes up most of COLM, formed by ephemeral streams draining into the river. Among these canyons, Monument Canyon occupies much of northern COLM, and the entire southeastern section includes much of No Thoroughfare Canyon. COLM is adjacent to the cities of Grand Junction on the east, Fruita on the north, and Glade Park on the southwest. Grand Junction and Fruita are centered on the opposite side of the Colorado River. McInnis Canyons National Conservation Area, administered by the Bureau of Land Management, adjoins the monument along most of its western boundary.



COLM Figure 1. Erosion of sedimentary rock formations gives COLM a variety of unusual landforms (VLS).

There are numerous publications that refer to the geology and paleontology of COLM. Many of them only briefly mention the monument, or deal primarily with faulting and deformation (for example Stearns and Jamison [1977] and Jamison and Stearns [1982], or a recent series of Geological Society of America abstracts, the most recent being Schoolmeesters et al. [2007]).

Documents about the general geology of COLM include: Lohman (1960, 1963, 1965a, 1965b, 1981); Miller and Coale (1969); Cole et al. (1999); Scott et al. (2001a, 2002); KellerLynn (2006); and Lucas et al. (2006a). Documents about the monument's paleontology include: Holt (1942); Callison (1977); Hasiotis (1997); Engelmann (1999); Engelmann and Hasiotis (1999); Engelmann and Fiorillo (2000); Anderson et al. (2000); Scott et al. (1999); Scott et al. (2001b); Koch and Santucci (2002); King et al. (2004); Trujillo et al. (2004, 2005); Bilbey et al. (2005); Imhof and Trujillo (2005); Kirkland et al. (2005); Foster and Lockley (2006); Lockley and Foster (2006); and Lucas et al. (2006b). The Geologic Resources Division of the National Park Service (NPS) coordinated a geologic resource inventory scoping session for COLM during June 1998 and produced a geologic resource evaluation report for the monument in 2006 (KellerLynn 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002). A paleontological survey was performed for the monument and documented by Trujillo et al. (2004).

The oldest rocks exposed at COLM are metamorphosed sedimentary rocks that are at least as old as the Paleoproterozoic (see the appendix for a geologic time scale). A date of 1,741 ± 11 Ma (million years) has been obtained for these rocks, but it is likely that this is the age of the metamorphism, not the original rocks. Now-metamorphosed igneous rocks were emplaced in the metasedimentary rocks around 1,721 ± 14 Ma, and are exposed in No Thoroughfare Canyon. The ancient rocks are records of a time before North America was assembled as a continent. COLM was near the proto-North American continental margin 1.74 billion years ago, when mountain building metamorphosed the local sedimentary rocks. Later, around 1,400 Ma, igneous dikes cut through the host rocks. Following the dikes, there is a gap in time of more than a billion years until the deposition of the Chinle Formation began, around 220 Ma. In the intervening time, rocks were formed and eroded, resulting in what is called the Great Unconformity. Near the end of the gap, western Colorado underwent uplift, with the Ancestral Uncompahgre Plateau and Ancestral Rocky Mountains appearing and then being worn down between about 320 and 245 Ma (Scott et al. 2001a). Eventually, the topography was once again suitable to preserve deposition, and rocks of the Chinle Formation were deposited in the Grand Junction area. The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time

(Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003).

Chinle Formation rocks show that the area gradually became drier over time, as the supercontinent Pangaea moved north. What had been a setting of rivers, lakes, and deltas turned arid (Tanner 2003). This is represented at COLM by the first of several giant ergs (dune seas), the Wingate Sandstone of the Glen Canyon Group (Scott et al. 2001a). The Wingate Sandstone erg of the Late Triassic and Early Jurassic was ended by a more humid period recorded in the Kayenta Formation. The Kayenta Formation represents a broad, gentle flood plain with a sea to the west (Blakey 1994; Peterson 1994).

About 190 Ma, this setting was gradually replaced by the vast erg that became the Navajo Sandstone (Harshbarger et al. 1957; Lucas et al. 2005), but the deposits of this sand sea are not present at COLM. Erosion destroyed the record of this period and the following period of marine incursion; a fuller record of the Middle Jurassic can be found at Arches National Park (ARCH), over 72 km (45 miles) southwest of COLM (see the ARCH summary in this document for further details). The Entrada Sandstone erg that replaced the seaway is the second and last large erg preserved at COLM. Another seaway was present in northern Colorado following deposition of the Entrada Sandstone, receding at the onset of Morrison Formation deposition (O'Sullivan 1992). Aridity decreased through the Late Jurassic (Lockley and Wright 2003) as Pangaea continued moving north; southern Utah reached 22° N during this time (Loope 2008). The more humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the last geologic formation of the Jurassic in COLM, the well-known Morrison Formation. By the time the Morrison sediments were deposited, the area had moved out of the low subtropical latitudes (Lockley and Mickelson 1997). The Morrison Formation is thought to have been a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004). During deposition of the upper portion of the formation (the Brushy Basin Member), a vast basin is thought to have extended from near Grand Junction to Albuquerque, New Mexico. This basin was periodically occupied by a shallow saline lake which has been named Lake T'oo'dichi'; the setting is described as analogous to the modern Lake Eyre basin in Australia (Turner and Fishman 1991). Another fluvial and floodplain unit, the Burro Canyon Formation, followed the Morrison Formation during the Early Cretaceous (Scott et al. 2001a).

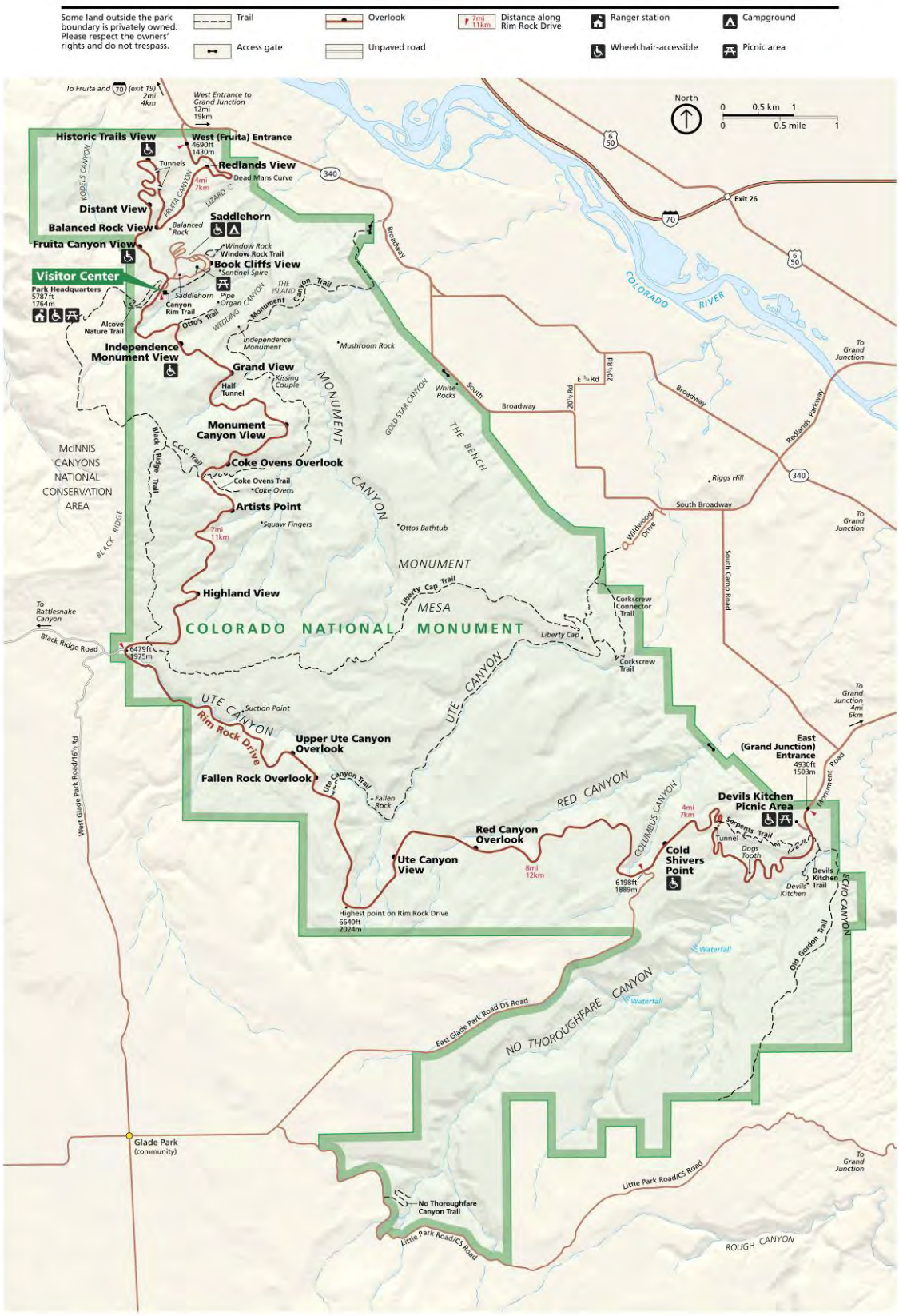
Terrestrial processes continued until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Formation signals the onset of several cycles when the Western Interior Seaway transgressed and regressed over the land (Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). None of the marine units of the Seaway are found within COLM, but the marine Mancos Shale is known from the Grand Junction–Fruita area (Lohman 1965a; Scott et al. 2001a).

Although deposition is not recorded at COLM for practically all of the past 100 million years, events occurring during that time span affected it significantly. Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of a major mountain-building event known as the Laramide Orogeny (Cather 2003). This event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). Laramide activity is responsible for much of the basic structure of COLM; the Redlands fault system that extends along most of the eastern boundary of COLM is traced by a large monocline. The lower "step" is northeast of the fault's hinge, and the higher "step", which makes up most of COLM, is southwest of it. The two are connected by an eroded hinge; these features are well-illustrated by the cross-sections of Scott et al. (2001a). Extensive downcutting and canyon formation followed episodes of uplift (Scott et al. 2001a).

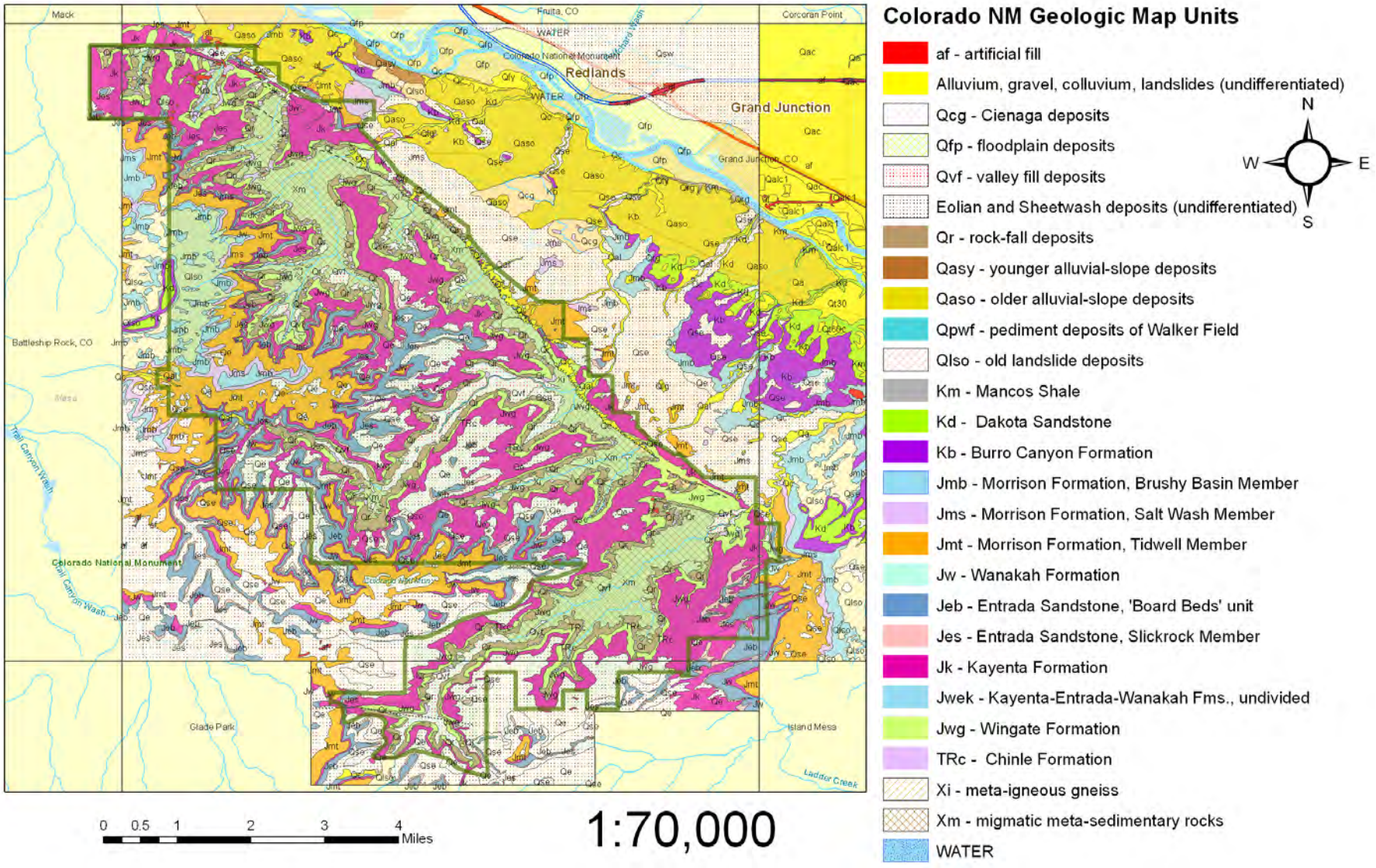
Glaciation occurred in the southern Rocky Mountains during the Pleistocene, which began approximately 2.59 Ma. There is evidence for glaciation on Grand Mesa, east of Grand Junction (Henderson 1923). People were present in the neighboring Piceance Creek Basin at least 8,000 to 7,000 years ago, perhaps as far back as 9,000 years ago (Hester 1974). Numerous archeological sites in COLM show the presence of humans for at least several thousand years (Lohman 1981).

Rock units exposed at COLM include, from oldest to youngest: metamorphosed sedimentary and igneous rocks (Paleoproterozoic); younger igneous dikes (Mesoproterozoic); the Chinle Formation (Upper Triassic); the Wingate Sandstone and Kayenta Formation of the Glen Canyon Group (Upper Triassic–Lower Jurassic); the Entrada Sandstone and Wanakah Formation of the San Rafael Group (Middle Jurassic); the Morrison Formation (Upper Jurassic); the Burro Canyon Formation (Lower Cretaceous); and the Dakota Formation (Lower–Upper Cretaceous). Surficial units include a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene). In general, outcrops in COLM are youngest in the west and oldest in canyon bottoms in the east (see COLM Figure 3 for a geological map) (Lohman 1965a; Scott et al. 2001a, 2002; KellerLynn 2006). All of the formations are definitely fossiliferous at COLM, with the exception of the Proterozoic units and the Dakota Formation (COLM Table 1). The Dakota

Formation is included under “Sources of Paleontological Resources” rather than “Potential Sources of Paleontological Resources” based on the descriptions of the unit in Scott et al. (2001a), who attributed common fossil types such as burrows to the formations for the area including COLM and its immediate vicinity. Most of the geologic nomenclature for the COLM area has been stable for decades. The exception is the stratigraphic interval including the Wanakah Formation and the lower Morrison Formation. Disputes here involve the placement of unit contacts and the names used. This document will use the nomenclature of KellerLynn (2006), established from the maps of Scott et al. (2001a, 2002); alternative interpretations will be noted in the text.



COLM Figure 2. Geography of COLM (NPS).



COLM Figure 3. Schematic geological map of COLM. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1047632.

Formation	Age	Fossils within COLM	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Plant fossils, a mammoth or mastodon tooth, bones of nine other mammal and bird taxa, and packrat middens	Alluvial, eolian, fluvial, and landslide deposits
Dakota Formation	Early–Late Cretaceous	Possibly plant fossils, root traces, and invertebrate traces	Terrestrial (especially fluvial) becoming shallow marine over time
Burro Canyon Formation	Early Cretaceous	Petrified wood, root traces, invertebrate traces, and dinosaur bones	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	Bivalves, gastropods, horseshoe crab traces, a lungfish tooth plate, bones of turtles, crocodile relatives, and dinosaurs, and a pterosaur footprint	Fluvial, floodplain, and lacustrine settings
Wanakah Formation	Middle Jurassic	Invertebrate traces and possible pterosaur feeding traces	Mud flats and/or shallow lakes
Entrada Sandstone	Middle Jurassic	Bioturbation	Coastal dunes and sand flats
Kayenta Formation	Early Jurassic	Local bioturbation and two bones	Primarily fluvial settings
Wingate Sandstone	Late Triassic–Early Jurassic	Bioturbation from roots and burrows, and tracks of dinosaurs, other reptiles, and mammal relatives	Desert with large eolian sand dunes
Chinle Formation	Late Triassic	Root traces, invertebrate traces, and rare bones	Fluvial, floodplain, and lacustrine settings, becoming drier over time
Igneous dikes	Mesoproterozoic	Unfossiliferous (intrusive igneous rocks)	Not applicable (igneous rocks that solidified at depth)
Metamorphosed sedimentary and igneous rocks	At least Paleoproterozoic	None to date; fossils are extremely unlikely because of lithology and metamorphism	Not applicable (igneous rocks that solidified at depth and strongly metamorphosed sedimentary rocks)

COLM Table 1. Concise tabulation of COLM stratigraphy, fossils, and depositional setting; details and references can be found in the text. Igneous and metamorphic rocks are included here for completeness, but are not discussed in detail in the text because of the improbability of fossils.

Sources of Paleontological Resources

The fossils of COLM present opportunities for education, interpretation, and continued or future scientific research in the monument. Seventy-five localities were reported by the 2004 paleontological survey (Trujillo et al. 2004, 2005), with an additional handful discovered since then, including a site at the mouth of Ute Canyon, and turtle and lizard track sites (J. Foster, Museum of Western Colorado curator of paleontology, pers. comm., December 2010). Fossils have been known from the Grand Junction area since at least 1885 (Chenoweth 1987). COLM fossil resources have historically been overshadowed by Morrison Formation sites just outside of monument boundaries: several localities excavated by Elmer Riggs and crew from the Field Columbian Museum (now the Field Museum) of Chicago in 1900 and 1901 in the Redlands east of the monument; and a cluster of sites united under the Bureau of Land Management's Fruita Paleontological Research Natural Area (FPA), just northwest of COLM in the McInnis Canyons National Conservation Area.

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location. At COLM, only one member is present, the informally-named red siltstone member (Stewart et al. 1972). The Chinle Formation of COLM is mostly composed of lime-rich mudstone, generally reddish in color, with a variety of

compositions ranging from silty claystone to sandy siltstone. Less than 10% of the formation is composed of reddish sandstone; lesser quantities of sandy conglomerate and limestone are also present. The formation produces a distinctive red slope beneath the cliffs of the Wingate Sandstone in the monument's canyon system (Scott et al. 2001a). The Chinle Formation is not especially thick at COLM; Stewart et al. (1972) found thicknesses of 29 m (95 ft) and 27 m (88 ft) at two locations bracketing the monument from north to south. The upper contact with the Wingate Sandstone is generally sharp in the area but appears to be conformable and locally gradational (Lohman 1965a). The red siltstone member is probably equivalent to the Dolores Formation of southwestern Colorado and the uppermost Chinle Formation elsewhere, including such members as the Church Rock Member in southeastern Utah and the Rock Point Member of Arizona (Stewart et al. 1972).

The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) because this division is known to be in the overlying Wingate Sandstone (Lucas et al. 1997a, 2006b). The red siltstone member probably dates to the latest stage of the Triassic (the Rhaetian) based on comparisons with other upper Chinle Formation members (Martz and Parker 2010). The depositional environment of the COLM-area Chinle Formation is interpreted as a vegetated floodplain or mud flat, with shallow ponds and streams (Scott et al. 2001a). The climate is thought to have been tropical monsoonal in this area (Dubiel 1992).

The red siltstone member is not known for its fossils, but it is fossiliferous at COLM. Scott et al. (2001a) reported root traces, possible crayfish burrows, and general bioturbation (biological disruption of sedimentary structures, usually from roots or burrows) to be common in Chinle Formation mudstone of the COLM area, and bioturbation and possible microbial laminations in the limestone. Scott et al. (2001b) reported a track site near the east entrance of the monument with footprints similar to the theropod footprint genus *Grallator* (these tracks are in a fallen block and may have come from the Wingate Sandstone instead; J. Foster, pers. comm., November 2011) and more generally the presence of root traces and invertebrate trace fossils including *Camborygma* (crayfish burrows), *Koupichnium* (horseshoe crab walking traces) and *Scoyenia* (possible beetle larval burrows). It is not clear if all of these trace fossil varieties are from COLM, as the original source, Hasiotis (1997) also discussed Canyonlands National Park and Petrified Forest National Park, and did not specify the trace fossil assemblages of the specific parks. Following the 2004 paleontological survey, Trujillo et al. (2004) reported the presence of trace fossils including root traces, possible invertebrate burrows, and feeding traces. They also reported that metoposaur amphibian material had been found within the current boundaries of COLM approximately 50 years before the survey. Purportedly, the material was donated to the Museum of Western Colorado (MWC; Grand Junction, CO), but could not be found in 2004 (Trujillo et al. 2004; J. Foster is skeptical that the material was given to the museum, because the museum didn't exist until approximately ten years after the collection would have been made, and didn't have a paleontologist on staff for almost ten more years [pers. comm., November 2011]). In an abstract, Trujillo et al. (2005) described the material as teeth and scutes of a metoposaur (which would seem to be an error, because metoposaurs did not have scutes).

There are also some fossil reports from the vicinity of the monument. There are finds including root traces, metoposaurs, therapsids (popularly known as "mammal-like reptiles"), aetosaurs (armored herbivorous reptiles), shuvosaurids (toothless reptiles resembling ostrich-mimic dinosaurs), "rauisuchians" (reptiles related to crocodiles but resembling large carnivorous dinosaurs), sphenosuchians (small terrestrial crocodile relatives), and theropod dinosaurs (bipedal carnivores) in the Eagle Basin east of COLM (Small and Sedlmayr 1995; Sedlmayr and Small 1997; Small 2001), and reptile tracks at Dinosaur National Monument (DINO; see the DINO summary in this document for more information). Numerous and diverse tetrapod tracks are known from roughly equivalent rocks at Gateway, a little over 40 km (25 miles) southwest of COLM (Lockley et al. 1996a, 2005), as well as pith casts of the horsetail *Neocalamites*, invertebrate traces, and lungfish burrows (Lockley and Jennings 1987). Roughly equivalent rocks near Bedrock, 75 km (47 miles) south-southwest of COLM have yielded fossils including plant casts and a coelacanth skull (Elliott 1987).

The various members of the Chinle Formation have similar types of fossils, but differ in the genera and species present and which are most commonly found. Plants are represented by charophyte algae (Lucas and Keitzke 1993), diverse palynomorphs (Gottesfeld 1972; Fisher and Dunay 1984; Parker 2004), and leaves, wood, and other remains of lycopods, horsetails, ferns and fern-like plants, cycads, bennettitales, ginkgoes, conifers, and taxa of uncertain classifications (Ash 1972, 1989, 2003, 2005a; Stewart et al. 1972; Parker and Ash 2004). Fungal remains are also present (Ash 1972; Stewart et al. 1972). Invertebrates are represented by tentaculites (enigmatic possible mollusks, misidentified as worm fossils [Ash 2005b]), bivalves, gastropods, ostracodes (seed shrimp), conchostracans (clam shrimp) (Breed 1972; Stewart et al. 1972), crabs (Rinehart et al. 2003), crayfish (Miller and Ash 1988), insects (Breed 1972; Stewart et al. 1972), and various trace fossils (Caster 1944; Breed 1972; Gillette et al. 2003; Demko et al. 2005), including borings in wood (Walker 1938; Tapanila and Roberts 2012) and a variety of feeding traces on plants (Ash 1996).

The Chinle Formation's vertebrate assemblage has been summarized in a number of documents, including Stewart et al. (1972), Colbert (1972), Parrish (1989), Long and Murry (1995), Heckert et al. (2005a, 2005b), Irmis (2005), Parker (2005, 2009), Lucas et al. (2007), Loughney et al. (2011), and Parker and Martz (2011). Vertebrates include: freshwater sharks; ray-

finned fish; lungfish; coelacanths; several types of enigmatic tetrapods known from isolated bones and teeth; metoposaur amphibians; procolophonids (stocky lizard-like herbivorous and insectivorous reptiles); sphenodontians (relatives of the modern tuatara); trilophosaurids (reptiles that resembled large herbivorous lizards); tanystropheids (long-necked piscivorous reptiles); the unusual aquatic reptile *Vancleavea*; drepanosaurids (beaked, possibly aquatic chameleon-like reptiles); several forms of crocodile-like phytosaurs, known under a variety of names; several forms of herbivorous armored aetosaurs, also known under a variety of names; the small unusual armored *Revueltosaurus*; "rauisuchians" including poposaurids, rauisuchids, and shuvosaurids; "sphenosuchians"; lagerpetonids (small bipedal dinosaur relatives); silesaurids (quadrupedal dinosaur relatives); several early dinosaur taxa; and dicynodonts (a type of non-mammalian therapsid, or "mammal-like reptile"). Lungfish, metoposaurs, aetosaurs, and phytosaurs are the most common forms. Vertebrate trace fossils include coprolites (fossil dung) (Hunt et al. 1998), swim traces, and footprints possibly attributable to sphenodontians, tanystropheids, phytosaurs, aetosaurs, carnivorous and herbivorous dinosaurs, and mammal relatives (Hunt and Lucas 2007). Bones accumulated in floodplains, bogs, small ponds, and in or near channels (Parrish 1989).

Glen Canyon Group: Wingate Sandstone (Upper Triassic–Lower Jurassic)

The Wingate Sandstone is an eolian sandstone unit that spans the Triassic–Jurassic boundary (Lucas et al. 2006b). Limestone lenses are also present locally (Harshbarger et al. 1957). At COLM, this formation is responsible for much of the monument's striking scenery, forming 100 m (330 ft)-tall cliffs. The great majority is orange-tinted stratified lime-rich sandstone, with brown to reddish brown siltstone intervals making up the rest. The siltstones represent interdune deposits among the great eolian (wind-deposited) sand dunes, which are thought to have reached heights of 30 to 45 m (98 to 148 ft) during arid periods (Scott et al. 2001a). Winds blew material to the southeast (Lohman 1965a). The Wingate Sandstone of the COLM area appears to have formed in a series of dune complexes over time (Scott et al. 2001a). It ranges from 95 to 110 m (310 to 360 ft) thick in the area, and in the monument a number of faults formed in it during the Laramide Orogeny (Jamison and Stearns 1982). The upper contact with the Kayenta Formation is sharp (Scott et al. 2001a) to gradational (Lohman 1965a). After much debate concerning the age of the Wingate Sandstone, it was thought that the Wingate Sandstone was entirely Jurassic in age, but the discovery of a skull of the phytosaur *Redondasaurus* at the base of the formation in Lisbon Valley in Utah shows that at least some of the unit is Late Triassic in age (Lucas et al. 1997a). Trace fossils suggest that the Triassic–Jurassic boundary is in the middle to upper 75 m (245 ft) of the formation (Lucas et al. 2006b).

The Wingate Sandstone has had a complicated history (Colbert and Mook 1951). It was originally defined to be much the same rocks as what was later named the Entrada Sandstone, but the Entrada Sandstone eventually took over most of these rocks, including the original type section for the Wingate Sandstone (at Fort Wingate, New Mexico). Furthermore, the Wingate Sandstone was later divided into the lower Rock Point Member and the upper Lukachukai Member, but where the Rock Point Member has been assigned to the Chinle Formation, the Lukachukai Member is abandoned as superfluous (Dubiel 1989).

The Wingate Sandstone of COLM is fossiliferous, with a number of trace fossil discoveries (COLM Figure 4, 5). According to John Foster (pers. comm., November 2011):

> The first tracks found in the Wingate at COLM were the ones above Devils Kitchen, found by Josh Smith and Bonnie [Carter?] in 2000. This is 6758 in Lucas's report [Lucas et al. 2006b]. The second site found was a large fall block in Ute Canyon that Vaia Barkas and I found in 2002. This became Lucas's 6761. We showed Ryan King these sites and he went out and found more in Ute Canyon in 2003.

These are the sites described by King et al. (2004). This publication reported five localities with 18 trackways and 79 tracks, all assigned to the theropod ichnogenus *Grallator*. They were found in interdune deposits (very thin mud drapes between dune sets; J. Foster, pers. comm., November 2011); unusually, one was found in the middle of the formation. Lucas et al. (2006b) provided descriptions of several of the track localities. New Mexico Museum of Natural History and Science (NMMNH) locality 6758 near Devils Kitchen has *in situ* *Grallator* and *Eosauropus* (from early sauropods) tracks 25 m (82 ft) above the base of the formation. The MWC collected one of the *Grallator* tracks from this site for COLM's collections (J. Foster, pers. comm., November 2011). NMMNH localities 6759 through 6763 are in Ute Canyon. Only NMMNH 6762 is *in situ*, with numerous "chirothere" (tracks of reptiles related to crocodiles such as aetosaurs and rauisuchians) tracks about 3 m (10 ft) above the base; there may be *Grallator* tracks at this site as well (J. Foster, pers. comm., November 2011). The other sites are fallen blocks, with synapsid (derived "mammal-like reptiles" and true mammals) and *Grallator* tracks. Trujillo et al. (2005) reported the presence of several places in the monument where numerous theropod tracks had been found on fallen blocks. John Foster (pers. comm., November 2011) reported finding a track site in an overhang just outside Ute Canyon, near the top of the formation, with Josh Smith around 2007. Scott et al. (2001a) reported both the sandstone and siltstone intervals of this formation to show evidence of bioturbation by roots and burrows.





COLM Figure 4. Wingate Sandstone theropod dinosaur tracks from Ute Canyon (photo courtesy John Foster, Museum of Western Colorado).
COLM Figure 5. An individual theropod footprint (photo courtesy John Foster, Museum of Western Colorado).

In general, fossils are uncommon in the Wingate Sandstone, with the phytosaur skull mentioned above being the only body fossil of note, but trace fossils are known from multiple levels in Utah, Arizona, and Colorado. Trace producers include invertebrates (Odier et al. 2004), mammal-like animals, nondinosaurian reptiles, and theropod, prosauropod, and possible sauropod ("brontosaur") dinosaurs (Odier et al. 2004; Lucas et al. 2006b, 2006c). Mammal-like tracks are often found nearly alone and are confined to dunes (Odier et al. 2004). There appears to be distinct Late Triassic and Early Jurassic assemblages, with mammal-like tracks limited to the Late Triassic and prosauropod tracks appearing in the Early Jurassic (Lucas et al. 2006b, 2006c). Gateway (Lockley and Jennings 1987; Lockley et al. 1996a, 2005) and Bedrock (Dubiel et al. 1989) also have track sites.

Glen Canyon Group: Kayenta Formation (Lower Jurassic)

The Kayenta Formation is mostly made up of fine- to coarse-grained sandstone with some small amounts of interbedded shale and siltstone, and rare limestone and conglomerate (Vandiver 1936). At COLM, the Kayenta Formation forms ledges and cliffs (Scott et al. 2001a), often making a bench between the cliffs of the Wingate Sandstone and Entrada Sandstone. The Kayenta Formation of the monument is composed of red and purple siltstone, shale, sandstone, and conglomerate (Lohman 1981). It ranges from up to 24 m (80 ft) thick in the northern part of the monument to 14 m (47 ft) thick in the southeast, disappearing outside of the monument. It is more conglomeratic in the western part of the monument. The upper contact with the Entrada Sandstone is a disconformity (Lohman 1965a). The age of the Kayenta Formation is now thought to be middle Early Jurassic, between approximately 197 and 190 Ma (Lucas et al. 2005). The Kayenta Formation was formed by shifting freshwater braided and meandering streams (Colbert 1974); floodplain deposits are also known (Lucas et al. 2005). The primary depositional environment of the Kayenta Formation in the COLM area is interpreted as fluvial (Lohman 1965a), with many channels present (Stearns and Jamison 1977).

The Kayenta Formation is sparsely fossiliferous at COLM. To date, two bones have been reported from this formation at the monument, both from a mudstone-pebble conglomerate near the top of the formation. Rodney Scheetz found a dinosaur bone fragment during a 2002 survey of COLM trails, and the 2004 survey collected another fragment, identified in 2004 as possibly from a theropod limb bone (Trujillo et al. 2004), and in 2005 as a reptilian rib fragment (Trujillo et al. 2005). The 2004 bone is in COLM collections (Trujillo et al. 2004). Siltstone may be also locally bioturbated (Scott et al. 2001a). In neighboring Utah, the formation is known for its dinosaur tracks (Santucci and Kirkland 2010). Vertebrate body fossils of the Kayenta Formation have been described in many publications, but most are from a silty facies found to the southwest in Arizona (see Lucas et al. [2005] or the Navajo National Monument summary of the Southern Colorado Plateau Network report for a partial list). Vertebrate body fossils of the sandy facies of the Kayenta Formation, which would be more typical of COLM, include fish scales, crocodylomorphs (crocodiles and their relatives), and tritylodontid synapsids (early mammal relatives) such as *Kayentatherium* (Clark and Fastovsky 1986). Kayenta Formation vertebrate trace fossils include coprolites (Luttrell and Morales 1993), therapsid tracks, lizard-like tracks (Hamblin and Foster 2000), and dinosaur tracks such as *Grallator* and *Eubrontes* (Lucas and Tanner 2007). The relevant ichnofossil taxonomy can be complicated. A recent summary for the Chinle Formation and Glen Canyon Group can be found in Lucas et al. (2006c). Kayenta Formation fossils that are not vertebrates or

vertebrate traces include microbial limestone (Luttrell and Morales 1993), petrified wood (Jenkins and Shubin 1998), plant impressions (Rigby et al. 1971) freshwater bivalves and gastropods (Harshbarger et al. 1957), ostracodes (Lucas and Tanner 2007), and invertebrate trails and burrows (Luttrell and Morales 1993).

San Rafael Group: Entrada Sandstone (Middle Jurassic)

The Entrada Sandstone is the most prominent cliff-former at COLM after the Wingate Sandstone; in some places, the two form a combined cliff with the Kayenta Formation (Lohman 1965a). The Entrada Sandstone is dominated by red sandstone, with some silt (Harshbarger et al. 1957). It is divided into multiple members depending on location. Only part of the formation is present at COLM. Historically, the Entrada Sandstone of COLM is divided into the Slick Rock Member and the Moab Member (or Tongue) (Lohman 1965a), but recently the Moab Member has been dropped for the informal "board beds member" (Cole et al. 1999; Scott et al. 2001a, 2002; Potter and Hayden 2006). The Moab Member was regarded by Kirkland et al. (2005) as pertaining to the Curtis Formation. The Slick Rock Member is around 34 m (110 ft) thick at COLM, and the "board beds member" is about 13 m (43 ft) thick (Scott et al. 2001a). The Slick Rock Member, almost entirely sandstone, is expressed as a pale orange thin cliff or bench (Scott et al. 2001a), while the "board beds member" is a series of white sandstone benches (Lohman 1965a).

The upper contact with the Wanakah Formation is sharp (Lucas et al. 2006c). The Entrada Sandstone lacks age-diagnostic fossils and is regarded as late Middle Jurassic in age (Lucas et al. 1985). The climate was semi-arid to arid (Ekdale and Picard 1985) and hot, with sparse vegetation. The Entrada dunes made up the last major erg of the area, and formed when southern Utah was about 22° N (Loope 2008). The Slick Rock Member is interpreted as a coastal dune field with dunes 10 to 20 m (33 to 67 ft) tall and wet interdunes. The "board beds member" is interpreted as representing a wet coastal sand flat. The Entrada Sandstone system is perhaps analogous to the Baja California coast at Guerrero Negro (Scott et al. 2001a).

There is one specific report of fossils at COLM in the Entrada Sandstone: Scheetz found a bioturbated zone in the "board beds" in 2002 (Trujillo et al. 2004). Additionally, Scott et al. (2001a) reported bioturbation in both members in the COLM area, and Potter and Hayden (2006) mentioned the presence of bioturbation and nonmarine trace fossils in the "board beds member" at unspecified sites among a set that includes COLM. Fossils are generally rare in the Entrada Sandstone. Many reports of fossils from the Entrada Sandstone are from the Moab Member, which is now regarded as part of the Curtis Formation (Doelling 2010). These include the well-known megatracksite of the Moab area (Lockley 1991; Lockley et al. 2007) and the crocodylomorph *Entradasuchus* (Hunt and Lockley 1995). Trace fossils have been reported from the Escalante Member in south-central Utah, including rare root traces, insect burrows (Loope 2008), vertebrate burrows, traces of sand-swimming vertebrates (Graverson et al. 2007), and theropod (large *Megalosauripus* and smaller *Therangospodus*) and sauropod (cf. *Brontopodus*) dinosaur tracks (Foster et al. 2000). Hoggan (1970) reported bivalves and crinoids in marine facies of the Entrada Sandstone in the Vernal area.

San Rafael Group: Wanakah Formation (Middle Jurassic)

The Wanakah Formation and overlying Tidwell Member of the Morrison Formation represent the most problematic stratigraphic interval of COLM. Generally, older references use the Summerville Formation for the rocks between the Entrada Sandstone and Salt Wash Member of the Morrison Formation at COLM (Lohman 1960, 1965a; Jamison and Stearns 1982). Scott et al. (2001a, 2002) used the Wanakah Formation and Tidwell Member instead, incorporating part of the Salt Wash Member into the Tidwell Member. As noted above, for consistency with KellerLynn (2006), the Wanakah-Tidwell terminology is used here. Lucas et al. (2006a) produced a different interpretation, in which the Summerville Formation name is reintroduced, the Tidwell Member of the Morrison Formation becomes the Tidwell Member of the Summerville Formation, and the contact of the Tidwell and Salt Wash members is returned to its previously-recognized level. This is further complicated by Lucas et al. (2006a)'s contention that the Summerville and Wanakah formations are synonymous; other workers, including Condon (1992), O'Sullivan (1992, 2004) and O'Sullivan et al. (2006), have found the Wanakah Formation to be older than the Summerville Formation. The Wanakah Formation is divisible into multiple members at some locations (Goldman and Spencer 1941; Steele 1985; O'Sullivan 1992, 2004), but only the uppermost part would be present as far north as COLM (O'Sullivan 2004).

Scott et al. (2001a) described the Wanakah Formation of COLM as a red and green slope-forming mudstone unit with minor sandstone and limestone, about 9 m (30 ft) thick, with a disconformable upper contact with the Morrison Formation. They interpreted its depositional environment as terrestrial mud flats and/or shallow lakes (Scott et al. 2001a), although elsewhere there are beds deposited under marine incursions (O'Sullivan et al. 2006). The formation dates to the end of the Middle Jurassic (Callovian stage) (Scott et al. 2001a), about 165 to 160 Ma. The Summerville Formation is interpreted as hypersaline shallow marine and tidal flat deposits (Morris et al. 2010).

The Wanakah Formation is fossiliferous at COLM, having yielded a few invertebrate trace fossils and possible pterosaur feeding traces. These fossils were found by Scheetz in 2002; the possible pterosaur traces could not be relocated by the 2004 survey (Trujillo et al. 2004). Scott et al. (2001a) noted the presence of bioturbation and root traces. Bilbey et al. (2005) mentioned undescribed sauropod bones from COLM discovered

from a level at or near the base of the Morrison Formation, which would indicate either the Wanakah Formation or Tidwell Member. The Wanakah Formation is not known for its fossils; reports from areas outside of COLM of fossils in the upper section include microbial limestone (Hansen 1968), possible sponge spicules, bivalves, gastropods, ostracodes, indeterminate fossils, and possible worm tubes and other trace fossils (O'Sullivan et al. 2006).

If the stratigraphy proposed by Lucas et al. (2006a) is correct, and the interval is the Summerville Formation with an upper Tidwell Member, then horseshoe crab traces (Scott et al. 2001b), a lungfish tooth plate (Imhof and Trujillo 2005) and a pterosaur footprint (*Pteraichnus*) (Lucas et al. 2006a) from the Tidwell Member have also been found in the formation at COLM.

For reference, Summerville Formation fossils include plant debris (Petersen and Pack 1982), a single sauropod tail vertebra (Lucas and Heckert 2000), invertebrate traces (Lockley and Wright 2003), and a variety of vertebrate traces including pterosaur tracks (Lockley et al. 1995; Lockley and Wright 2003; Mickelson et al. 2004), theropod tracks (Lockley et al. 1996b), and sauropod tracks from near the top of the Summerville Formation at Moab (Lockley 1991). If the Tidwell Member is included in the Summerville Formation, as suggested by Lucas et al. (2006a), the fossil assemblage is broadened to include root traces (Kjemperud et al. 2008), algal and microbial laminations, charophytes (green algae), gastropods, ostracodes (Scott et al. 2001a), burrows (Kjemperud et al. 2008), horseshoe crab traces (Scott et al. 2001b), a lungfish tooth plate (Imhof and Trujillo 2005), indeterminate reptile bones (the putative pterosaur *Utahdactylus*; Bennett 2007), the poorly-known sauropod *Dystrophaeus* (Gillette 1996 Lucas and Anderson 1997), and pterosaur and dinosaur tracks (Lockley et al. 1998).

Morrison Formation (Upper Jurassic)

The Morrison Formation is a heterogeneous and variegated unit dated to the middle Late Jurassic, between about 155 and 148 Ma (Foster 2007). The formation is divisible into several members. From oldest to youngest, the Tidwell, Salt Wash, and Brushy Basin members are reported from COLM (although see above for discussion of alternative interpretations of the Tidwell Member). Extensive outcrops of all three members are found in the western part of the monument (Scott et al. 2001a). Overall, the formation is about 160 m (525 ft) thick at COLM, but this ranges widely (Scott et al. 2001a). Scott et al. (2001a) reported thicknesses of 38 m (125 ft), 31 m (102 ft), and 95 m (312 ft) for the Tidwell, Salt Wash, and Brushy Basin members, respectively. The Tidwell Member of COLM is a slope-former mostly made of mudstone, claystone, and siltstone, but with significant amounts of sandstone and limestone; all lithologies are present in widely-ranging proportions. It is a fluvial and lacustrine unit. The Salt Wash Member is a cliff-former composed of sandstone, claystone, siltstone, and mudstone, deposited in fluvial, floodplain, and lake settings. It has been compared to the Gulf Coast of Texas. The Brushy Basin Member is a slope-former composed largely of mudstones, with minor sandstone and limestone. It is interpreted as representing mud flat to shallow lacustrine deposition at COLM (Scott et al. 2001a), befitting a location near the margin of the Lake T'oo'dichi' basin. The climate is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004). The upper contact with the Burro Canyon Formation is arbitrarily defined (Scott et al. 2001a). Lohman (1965a) reported having seen both potential intertonguing of the Brushy Basin Member and the Burro Canyon Formation, and apparent disconformable relationships at other sites in the COLM area.

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Brachiosaurus*, *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks (Turner and Peterson 1999; Foster 2007). The Morrison Formation is the most fossiliferous unit at COLM, and each member of the formation is fossiliferous in the monument (Trujillo et al. 2005). A survey was carried out by George Callison and field assistants in 1977 on the monument's Morrison Formation outcrops. They found fourteen fossil sites and reported fossils of bivalves, gastropods, turtles, crocodylomorphs (relatively close relatives to crocodiles), and dinosaurs (Callison 1977), including a pelvic bone of a dryosaur, a small herbivore (Scott et al. 2001b). Fossils were most abundant near Black Ridge (Callison 1977). George Engelmann resurveyed the sites in 1985, finding additional bivalves, gastropods, and a tail vertebra from a sauropod. Engelmann and Fiorillo performed another survey in 1995, finding several new sites and discovering that one of Callison's localities (adjacent to the Black Ridge Trail) had been vandalized (Scott et al. 2001b).

The Tidwell Member of COLM has yielded algal and microbial laminations, charophytes, gastropods, ostracodes (Scott et al. 2001a), horseshoe crab traces (Scott et al. 2001b), a lungfish tooth plate (COLM Figure 6) (Imhof and Trujillo 2005; Trujillo et al. 2005), a dinosaur locality (Trujillo et al. 2004), and a pterosaur footprint (*Pteraichnus*) (Lucas et al. 2006a). The Salt Wash Member has yielded charophytes (Scott et al. 2001a), petrified wood (LeRoy 1960), ostracodes (Scott et al. 2001a), theropod and sauropod bones (Trujillo et al. 2005), and tracks of turtles and theropod, sauropod, and ornithopod dinosaurs (Lockley and Foster 2006; Foster and Lockley 2006). Engelmann and Hasiotis (1999) noted the presence of deep sauropod tracks from within or adjacent to COLM, in the middle shaly beds of the Salt Wash Member. Lockley and Foster (2006) described two track horizons at a site they ascribed to the Salt Wash Member, but which was mapped as the Tidwell Member by Scott et al. (2001a; their Tidwell–Salt Wash contact may be too high [J. Foster, pers. comm., November 2011]). Both track horizons were found as natural casts on the underside of

sandstone overhangs. One of the horizons had two clusters of bipedal dinosaur tracks, and the other had 50 or more turtle traces (COLM Figure 7) (*Chelonipus* isp. per J. Foster, pers. comm., November 2011). The site was found by Marilyn Sokolosky in 2005 (Lockley and Foster 2006). Natural casts of sauropod tracks, associated with mudstones, have been found east of the turtle track site at the head of Ute Canyon (J. Foster, pers. comm., November 2011).





COLM Figure 6. Lungfish tooth plate from the Tidwell Member (photo by Margaret Imhof).
COLM Figure 7. Turtle tracks from the underside of a slab in the Salt Wash Member (photo by John Foster, Museum of Western Colorado).

Foster and Lockley (2006), in a wider survey of track sites, reported on several localities from the monument; all were from the Salt Wash Member when the member could be determined. They include: a sauropod hind foot print from Black Ridge Trail; a loose slab with two short ornithopod (beaked bipedal herbivorous dinosaurs) trackways from elsewhere on the trail, now in monument collections as COLM 00203; natural casts of single sauropod and bipedal dinosaur prints at the Frost Locality near the east entrance of the monument; a sauropod hind foot print from above and northwest of the Liberty Cap Trail in western Monument Mesa; a sauropod hind foot print from upper Ute Canyon; and the sites described by Lockley and Foster (2006), approximately 100 m (330 ft) west of the Ute Canyon site.

The Brushy Basin Member of COLM has yielded turtle shell material and dinosaur bones (Trujillo et al. 2005). All three members are also bioturbated (Scott et al. 2001a). Holt (1942) reported collecting specimens of the unionid bivalve *Vetulonaea faberi* from Rim Rock Drive approximately 5 km (3 miles) south of COLM headquarters, on a road cut. The member was unspecified. Engelmann (1999) and Engelmann and Fiorillo (2000) drew on information from COLM's Morrison Formation fossil sites. Kirkland et al. (2005) noted the presence of termite nests, soil bug burrows, and beetle burrows in the Morrison Formation of the monument. During 2010, COLM staff found a tridactyl (three-toed) dinosaur track south of Artists Point in the Tidwell Member. A party from the MWC investigated the site and found possible lizard tracks on a slab (COLM Figure 8). These tracks are currently under study (J. Foster, pers. comm., November 2011). Additional tracks were found at Artists Point in the Tidwell Member during 2011, and some were collected (J. Foster, pers. comm., November 2011).



COLM Figure 8. Lizard track (Photo by John Foster, Museum of Western Colorado).

The Grand Junction area is of historic and ongoing importance for the paleontology of the Morrison Formation. Turner and Peterson (1999) and Foster (2003) provided site-level inventories for vertebrate finds in the entire Morrison Formation, including a number of sites in the Grand Junction-Fruita area, and Armstrong and McReynolds (1987a) provided an overview of Grand Junction-Fruita sites and their geographic and stratigraphic positions. Of particular interest are the sites excavated by Elmer Riggs and his crew in 1900 and 1901 (recent summaries in Chenoweth [1987, 1991]), and the FPA (described in detail in Kirkland et al. 2005 and Kirkland 2006). The Riggs sites are known for providing the first described specimen of the giant sauropod *Brachiosaurus* and most of a skeleton of *Apatosaurus* (Chenoweth 1987). The FPA is

particularly noted for its record of small vertebrates, including sphenodonts, lizards, tiny dinosaurs, and mammals (Callison 1987; Kirkland 2006). Among the diminutive fossils are articulated specimens of lizards and sphenodontians (Kirkland 2006), the robust sphenodontian *Eilenodon* (Rasmussen and Callison 1981a), rare examples of unusual terrestrial crocodylomorphs (*Fruitachampsa* [Clark 1985, 2011]; *Macelognathus* [Göhlich et al. 2005]), the heterodontosaurid *Fruitadens* (tiny fanged dinosaur) (Callison and Quimby 1984; Callison 1987; Galton 2007; Butler et al. 2010, 2012), and early mammals (Rasmussen and Callison 1981b; Rougier et al. 1996), including the burrowing *Fruitafossor* (Luo and Wible 2005). Another FPA site provided a skull and skeleton of a new species of the horned theropod *Ceratosaurus* (Madsen and Welles 2000).

The Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites, fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans, ostracodes, and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods (lamp shells), gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006); however, some of the traces, such as those attributed to anthozoans, brachiopods, horseshoe crabs, and echinoderms are only known from marine-influenced units that are not found at COLM (Hasiotis 2004). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians, lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods, diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

Burro Canyon Formation (Lower Cretaceous)

The Burro Canyon Formation of the COLM area is heterogeneous, with lithologies including shale, siltstone, sandstone, conglomerate, and limestone (Lohman 1965a). It includes a lower sandstone-rich portion and upper siltstone/claystone-rich portion. The sandstone-rich portion is white to yellowish gray in color and occupies the lower third to two-thirds of the formation and forms cliffs, while the finer-grained upper portion forms pale red or greenish slopes. The formation is only found on Black Ridge within the monument (Scott et al. 2001a). On Black Ridge, the finer-grained portion is thicker, while east of the monument, near the mouth of No Thoroughfare Canyon, sandstone becomes more prominent. Overall, the lithology of the formation is quite variable (Lohman 1965a). It is nearly 30 m (98 ft) thick on Black Ridge (Scott et al. 2001a), but is at least 37 m (120 ft) thick to the east (Lohman 1965a). It is interpreted as mostly fluvial in origin in the area, with minor lacustrine deposition (Lohman 1965a). The sandstone-rich section represents braided streams with narrow floodplains, and the overlying finer-grained section is interpreted as floodplain and lake deposits (Scott et al. 2001a). Sediments were deposited under a semi-arid climate, probably reworking material from the Morrison Formation (Walters et al. 1987). The predominant paleoflow was to the north (Craig 1981).

The age of the Burro Canyon Formation is uncertain. Scott et al. (2001a) proposed an Aptian-Albian age (approximately 125 to 100 Ma) for the Burro Canyon Formation at COLM. It is usually thought of as having been deposited between the Barremian stage of the Early Cretaceous (130 to 125 Ma) and the onset of Dakota Formation sedimentation around 100 Ma. However, the base of the formation has been reported to intertongue with the Brushy Basin Member of the Morrison Formation in places, which would imply a Late Jurassic age for part of the formation (Ekren and Houser 1959; Dickinson and Gehrels 2008). It should be noted that the Burro Canyon Formation and Brushy Basin Member can be difficult to distinguish. Mudstones of the Burro Canyon Formation weather as fragments, while Brushy Basin Member mudstones forms a hard frothy crust (Ekren and Houser 1959). The upper contact with the Dakota Formation is disconformable (Lohman 1965a), but the Dakota Formation and the Burro Canyon Formation can also be difficult to distinguish; one difference is the relative

abundance of plant fossils, which are common in the Dakota Formation and rare in the Burro Canyon Formation (Simmons 1957). The Burro Canyon Formation and Cedar Mountain Formation of Utah are lateral equivalents. Generally, the Burro Canyon Formation is present east of the Colorado River, and the Cedar Mountain Formation is present west of it (Dickinson and Gehrels 2008).

The Burro Canyon Formation is fossiliferous within COLM. Trujillo et al. (2005) reported that large amounts of dinosaur material were present in fallen Burro Canyon Formation blocks in the monument. Because blocks of the Burro Canyon Formation and the overlying Dakota Formation are often displaced by landslides in the COLM area originating in the Brushy Basin Member of the Morrison Formation (Scott et al. 2001a), fossils from the two formations could have a significantly wider distribution than their limited outcrops in the monument. Two localities with fallen dinosaur bones were originally reported by Engelmann as Morrison Formation localities (Trujillo et al. 2004). Scott et al. (2001a) found that petrified wood is locally abundant in the sandstone beds, and root traces and burrows can be found throughout the formation. There are also significant plant fossils known from just outside of COLM. A newspaper article (Fishell 1981) reported the discovery of what were then the oldest known flowers, found in the lower Burro Canyon Formation on the west side of Black Ridge just west of the COLM boundary. The flowers were mentioned briefly by Armstrong and McReynolds (1987b). Young (1987) reported that fossils of the tree fern *Tempskya minor* and the angiosperm *Platanus* sp. (sycamore) had been found at Black Ridge. Poorly preserved plants and shells are known from the Grand Junction area (Lohman 1965a).

In general, the fossils of the Burro Canyon Formation are not well known at this time. The current fossil assemblage from the formation over its entire area of outcrops includes organic debris, lake-dwelling algae, palynomorphs including conifer and angiosperm pollen (Tschudy et al. 1984), fern leaves (Simmons 1957), *Tempskya* (Dayvault and Hatch 2005), cycadeoids (Tidwell et al. 1976), conifers (Simmons 1957), early flowers (Fishell 1981; Armstrong and McReynolds 1987b), plant fragments (Stokes 1952), wood fragments (Tschudy et al. 1984), freshwater bivalves (Reeside 1957), gastropods, freshwater ostracodes, ganoid fish scales (Simmons 1957), and dinosaur bones (Trujillo et al. 2005), including a theropod femur and claw (Armstrong 2010). Additionally, a site in Utah near the border with Colorado (the Hotel Mesa Quarry) has yielded fossils of sharks, bony fish, lungfish, various crocodile relatives, theropods, the sauropod *Brontomerus*, and ornithopods (beaked bipedal dinosaurs) (Taylor et al. 2011). It has been attributed to both the Burro Canyon Formation (Kirkland and Madsen 2007) and the Ruby Ranch Member of the Cedar Mountain Formation (Taylor et al. 2011); by geography, it is within the Burro Canyon Formation (Taylor et al. 2011).

Dakota Formation (Lower–Upper Cretaceous)

The Dakota Formation (also Sandstone or Group) is a heterogeneous rock unit with outcrops as distant as Minnesota; because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in areas on the western side of the seaway may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). The Dakota Formation is limited to the cap of Black Ridge within COLM (Scott et al. 2001a), where it is approximately 31 m (102 ft) thick; however, part of the upper section of the formation has been eroded (Scott et al. 2001a). In the COLM area, the Dakota Formation includes a basal conglomerate, followed by a grayish coaly mudstone interval, then a light-colored sandstone interval, and finally an interval of interbedded shale and sandstone that transitions into the overlying Mancos Shale (Scott et al. 2001a). With the exception of the upper interbedded interval, the conglomerate-coaly shale-sandstone division is common for the formation throughout the Colorado Plateau. These divisions reflect environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma (see Titus et al. [2005] for discussion of the age).

The fossils of the Dakota Formation are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. The Dakota Formation of COLM is not specifically known to be fossiliferous, but may be. Scott et al. (2001a) found bioturbation from plant roots and burrows to be common throughout the formation in the COLM area, and specifically reported the presence of coal seams and stringers, plant fragments, and bioturbation in the coaly mudstone interval, and bioturbation in the sandstone interval. Dinosaur tracks have been found in this formation in the Redlands east of COLM (Scott et al. 2001a). Marine bivalves and a gastropod have been found in the upper Dakota Formation near Grand Junction (Lohman 1965a). A salient issue for finding fossils in the Dakota Formation at the monument is the very restricted extent of outcrops (Trujillo et al. 2004).

The best-known terrestrial fossils from the Dakota Formation are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Formation (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Formation of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Formation for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms

have also been recognized (Baxter 1954). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Formation rocks in Utah (including from near Bryce Canyon National Park) reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Formation marine fossils include algae, foraminifera (amoeba-like organisms that form "shells") (Tibert et al. 2003), dinoflagellate cysts (a type of plankton) (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Formation may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera (large foraminifera from the late Paleozoic useful for correlating and dating beds), corals, bryozoans, brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary deposits of COLM are a heterogeneous group of deposits left by water, wind, and gravity movements, scattered throughout the monument (Scott et al. 2001a). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). COLM has records of both types. In 1965 a mastodon (or possibly mammoth; J. Foster, pers. comm., November 2011) tooth was reportedly found in No Thoroughfare Canyon by Jack Roadifer, but has been lost (Scott et al. 2001b). Bone and plant specimens were collected from a fissure exposed by a rockfall in January 2000. Nine vertebrate taxa were tentatively identified, including eagle, wolf, and bighorn sheep (Trujillo et al. 2004). COLM has packrat middens, but they are almost completely unpublished. Middens were observed and in some cases collected from the Rimrock Drive area during 1994 by Kirk Anderson of Northern Arizona University (Flagstaff, AZ) (Trujillo et al. 2004). Anderson et al. (2000) mentioned the presence of an unspecified number of middens in the monument, from elevations ranging from 1,890 to 2,139 m (6,200 to 7,018 ft), and ranging in age from 8,500 calendar years before present to modern. Scott et al. (1999) noted the presence of numerous packrat middens under overhangs at COLM, and the 2004 survey observed middens in "upper Monument Canyon, the Squaw Fingers, and upper and lower Ute Canyon;" these examples may be significantly older than Anderson's Holocene middens (Trujillo et al. 2004). Charcoal in valley fill within the monument has been dated to a range that includes most of the Holocene, from 9,190 ± 50 radiocarbon years before present (10,440–10,240 calendar years before present) to 1,280 ± 50 radiocarbon years before present (1,290–1,120 calendar years before present) (Scott et al. 1999, 2001a). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate.

Quaternary fossils are not common within about 100 km (60 miles) of COLM. The best-known site is Bison Alcove within ARCH. Bison Alcove is an alcove site that has abundant packrat middens (Sharpe 1993), packrat droppings (Smith and Betancourt 1991), and remains of other mammals, including *Ovis* (bighorn sheep) and the namesake bison (Mead et al. 1991; Sharpe 1993). Most of the midden dates are from the latest Pleistocene, particularly from 22,000 to 14,000 radiocarbon years before present and 14,000 to 11,000 radiocarbon years before present, while the megafaunal remains are much younger, from 405 ± 60 radiocarbon years before present (530–320 calendar years before present) and 355 ± 60 radiocarbon years before present (510–300 calendar years before present) (Sharpe 1993). Humans have also been in the alcove, leaving behind an artifact composed of a hornsheath and twig joined by resin (Mead et al. 1991). The alcove has also been host to raptor roosts (Sharpe 1993). Of other fossils, there are isolated reports of mammoths from several locations: Arches; Bartlett Wash, about 92 km (57 miles) southwest of COLM in Utah; Fruita; and Professor Valley, about 61 km (38 miles) southwest of COLM in Utah (Agenbroad and Mead 1989). Isolated mammal bones have been found deep in the old river gravels of the Grand Junction area, including a horse metatarsal from Fruita and a bison radius from Grand Junction (J. Foster, pers. comm., November 2011). Peat records from as far back as the Early Holocene are known from fens on Grand Mesa, but they are from high elevations (>3,100 m [10,170 ft]) (Sullivan and Gilmore 2006), much higher than COLM. Finally, at least one other ancient charcoal record is known from the vicinity, in Sieber Canyon 19 km (12 miles) west of COLM; its record goes back to about the Pleistocene–Holocene boundary, 11,160 to 10,750 radiocarbon years before present (Aslan et al. 2004).

Park Collections

There are 443 catalog numbers for paleontological specimens in COLM collections (omitting 9 educational models). Numerically, the bulk of the collection (325 specimens) come from Quaternary sediments. A number of formations not present within COLM are also represented. Stratigraphically, aside from the Quaternary specimens, there are 38 catalog numbers for fossils from the Upper Cretaceous-age Mancos Shale, 26 for the Eocene Wasatch/DeBeque formations, 16 for the Morrison Formation, 15 for the Eocene Green River Formation, 7 for the Pennsylvanian Round Valley Limestone, 7 for unknown provenance, 3 for the Upper Cretaceous Mesaverde Group, 2 each for the Hunter Canyon Formation of the Mesaverde Group and the Wingate Sandstone, and one each for the Middle Jurassic Sundance Formation and Eocene Uinta Formation. Taxonomically, there are 13 catalog numbers for bulk sediment samples from middens, 2 for ferns, 10 for gnetophytes, 66 for conifers, 42 for monocotyledonous plants, 116 for dicotyledonous plants, 44 for unidentified plants, 7 for brachiopods, 28 for bivalves, 19 for gastropods, 11 for cephalopods, 19 for insects, 1 for plants with insects, 1 for sharks, 1 for bony fish, 5 for non-dinosaurian reptiles, 17 for dinosaurs, 28 for mammals, and 1 for unidentified animal bones. Several specimens are casts of bones, including a series of *Brachiosaurus* casts (K. Saba, assistant archivist, Intermountain Region Museum Services Program, pers. comm., December 2010). Two additional trace fossil specimens are on display at the monument and are not currently accessioned (J. Anzelmo, COLM superintendent, pers. comm., November 2010). Casts of the dinosaur and turtle traces described by Lockley and Foster (2006) are in the University of Colorado–Denver Dinosaur Tracks Museum and the Museum of Western Colorado (Lockley and Foster 2006).

Cultural Resource Contexts

People were present in the neighboring Piceance Creek basin at least 8,000 to 7,000 years ago, perhaps as far back as 9,000 years ago (Hester 1974), and COLM has evidence of human presence for several thousand years; a 1963 survey found 75 archeological sites in the monument. The survey found a variety of objects that, depending on age, could be examples of paleontological resources associated with or used as cultural resources, such as wood awls, corncobs, corn kernels, bone fragments, and yucca fiber strands. Two cave localities in No Thoroughfare Canyon have yielded similar material (Lohman 1981). Stone tools found in the monument are also potentially examples of such crossover resources, as stone tools are often made from fossils or fossiliferous materials like petrified wood and fossiliferous chert. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for COLM during June 1998 and produced a geologic resource evaluation report for the monument in 2006 (KellerLynn 2006). The participants of the scoping meetings developed a series of recommendations for future work. Monitoring recommendations at the site level were provided by Trujillo et al. (2004), as well as directions for future paleontological work, including: formation- or member-specific studies of the Wingate Sandstone, Kayenta Formation, Tidwell Member, Burro Canyon Formation, and Quaternary deposits; and further exploration of the Liberty Cap site and two microvertebrate sites.

- COLM reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 75 localities, 72 of which were reported to be in good condition in 2010. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.

- Fossil theft and vandalism are of concern at COLM; for example, one of Callison (1977)'s sites within COLM has been vandalized (Scott et al. 2001b; KellerLynn

2006). This issue was also discussed by Trujillo et al. (2004). Parks such as Petrified Forest National Park have investigated casual collecting of fossils and the social science driving the behavior. Contact the Geologic Resources Division for further information or assistance.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Within the NCPN, the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and increase understanding of this important fossil-bearing unit. A similar USGS–NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Data Sets

Colorado National Monument Paleontological Archives
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Colorado National Monument Collections
1911–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by COLM staff; status: Active.

Colorado National Monument Digital Geologic Map
2006–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=1047632.

Morrison Extinct Ecosystem Project
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Museum of Western Colorado
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New Mexico Museum of Natural History and Science
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Northern Arizona University
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University of Colorado–Boulder Dinosaur Tracks Museum collection (formerly at University of Colorado–Denver)
2000s–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by museum staff; status: Active.

Curecanti National Recreation Area

Curecanti National Recreation Area (CURE), Colorado, protects three reservoirs (Blue Mesa, Morrow Point, and Crystal) and associated areas on the Gunnison River, including the upper Black Canyon of the Gunnison (CURE Figure 1). These three reservoirs are popular locations for outdoor recreation. CURE was established February 11, 1965 under a cooperative agreement with the Bureau of Reclamation. It encompasses 16,985.64 ha (41,972.42 acres), all of which are under federal administration (CURE Figure 2). It is administered jointly with Black Canyon of the Gunnison National Park (BLCA), which adjoins CURE at its downstream boundary.

Geologic Background

CURE spans a 64-km (40-mile)-long stretch of the Gunnison River in eastern Montrose and western Gunnison counties in western Colorado. The recreation area encompasses three reservoirs on the river. From east to west, these are the Blue Mesa, Morrow Point, and Crystal reservoirs, separated by individual dams. The Black Canyon of the Gunnison begins below Blue Mesa Dam, which separates Morrow Point Reservoir from Blue Mesa Reservoir. Immediately downstream of CURE is BLCA. To the north are the West Elk Mountains, which include the Gunnison National Forest. Physiographically, the Black Canyon crosses the Rocky Mountain and Colorado Plateau provinces, with CURE within the Rocky Mountain province and BLCA within the Colorado Plateau province (Hansen 1981). The city of Gunnison is about 7 km (5 miles) east of the eastern terminus of CURE, and the city of Montrose is about 20 km (12 miles) west of the western terminus of CURE.



CURE Figure 1. A view across Blue Mesa Reservoir, over eroding Morrison Formation rocks (JPK).

Documents concerning the general geology of CURE include Hansen (1965, 1967, 1971, 1981, 1987a, 1987b), Hedlund and Olson (1973), Hedlund (1974), Heise (1999), and Thornberry-Ehrlich (2005). Documents including information about the paleontology of CURE include Fiorillo and May (1995, 1996), Fiorillo (1996, 1999), Fiorillo and Hasiotis (1996), Fiorillo and McCarty (1996), Fiorillo et al. (1996), Bailey and Wixom (1997), Turner and Peterson (1999), Fiorillo and Harris (2000), Trujillo (2000, 2001), Scott et al. (2001a), Koch and Santucci (2002), Engard (2003), Foster (2003), Koch (2005), Kenworthy (2006), Koch and Zichterman (2006), Koch et al. (2006), Odier (2006), Santucci et al. (2009), and Olstad (2011). The Geologic Resources Division of the National Park Service (NPS) coordinated a geologic resource inventory scoping session for CURE and BLCA (Heise 1999) and produced a geologic resource evaluation report for the two units in 2005 (Thornberry-Ehrlich 2005). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The oldest rocks exposed at CURE are metamorphosed sedimentary rocks that are at least as old as the Paleoproterozoic (see the appendix for a geologic time scale), predating 1,700 Ma (million years ago) (Hansen 1981). During this time, the area was covered by a thick sequence of sediments (Hansen and Peterman 1968), possibly as part of an arc or back-arc basin (Jessup et al. 2006). The geographic area consisted of a series of island arcs similar to the present-day Banda Sea in Indonesia, colliding with the growing proto-North American continent around 1,700 Ma (Jessup et al. 2005). During this process, igneous rocks intruded the crust at depth, and additional crustal fragments accreted onto the continent, deforming the original rocks of the Black Canyon region. Tectonic activity paused, then resumed around 1,430 Ma and lasted for about 30 million years, producing another episode of deformation (Jessup et al. 2006). Diabase dikes intruded around 510 Ma (Hansen 1981), but otherwise the Neoproterozoic and Paleozoic are not represented by rocks in the Black Canyon. It is known that the Black Canyon area was part of the ancestral Uncompahgre uplift around 300 Ma (Kellogg 2004), which would have led to erosion of overlying rocks and precluded deposition until the uplift in turn was eroded.

There follows a hiatus in the rock record at CURE until the Middle Jurassic. The dune sandstones of the Entrada Sandstone are the oldest such rocks in the Black Canyon area (Hansen 1965; Kellogg 2004), but the formation pinches out near the BLCA–CURE boundary (Hansen 1987a). The overlying Wanakah Formation is the oldest Mesozoic formation at CURE, representing a shallow marine setting that was replaced by a short-lived dune field (Kellogg 2004). Aridity decreased through the Late Jurassic (Lockley and Wright 2003). The more

humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the last geologic formation of the Jurassic in CURE, the well-known Morrison Formation. The Morrison Formation is interpreted as a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004); however, a narrow marine embayment (the Black Canyon Embayment) persisted into early Morrison time in the Black Canyon area, permitting marine influence (O'Sullivan 1992). During deposition of the upper portion of the formation (the Brushy Basin Member), a vast basin is thought to have extended from near Grand Junction, Colorado to Albuquerque, New Mexico. This basin was periodically occupied by a shallow saline lake which has been named Lake T'oo'dichi'; the setting is described as analogous to the modern Lake Eyre basin in Australia (Turner and Fishman 1991). Another fluvial and floodplain unit, the Burro Canyon Formation, followed the Morrison Formation during the Early Cretaceous (Hansen 1965).

Terrestrial processes continued until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Sandstone signals the onset of several cycles when the Western Interior Seaway transgressed (advanced) and regressed (withdrew) over the land (Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Marine deposition at CURE from the Seaway is represented by the Mancos Shale (Hansen 1965). The Black Canyon area was under water for almost the entire Late Cretaceous, only becoming emergent again in the last few million years of the period (Lillegraven and Ostresh 1990; Franczyk et al. 1992). The sea was not very deep; depth during the most extensive marine transgression was on the order of 90 to 180 m (300 to 600 ft) in western Colorado and eastern Utah (Hill 1982). The Sevier Orogeny (mountain-building event) pushed up rocks on the west of the Colorado Plateau in central Utah, providing source rocks for the thick Late Cretaceous sedimentary formations of the Plateau (Dickinson and Gehrels 2008). At times, such as approximately 94 Ma, volcanoes were active to the west, leading to the deposition of ash beds (Bagshaw 1977).

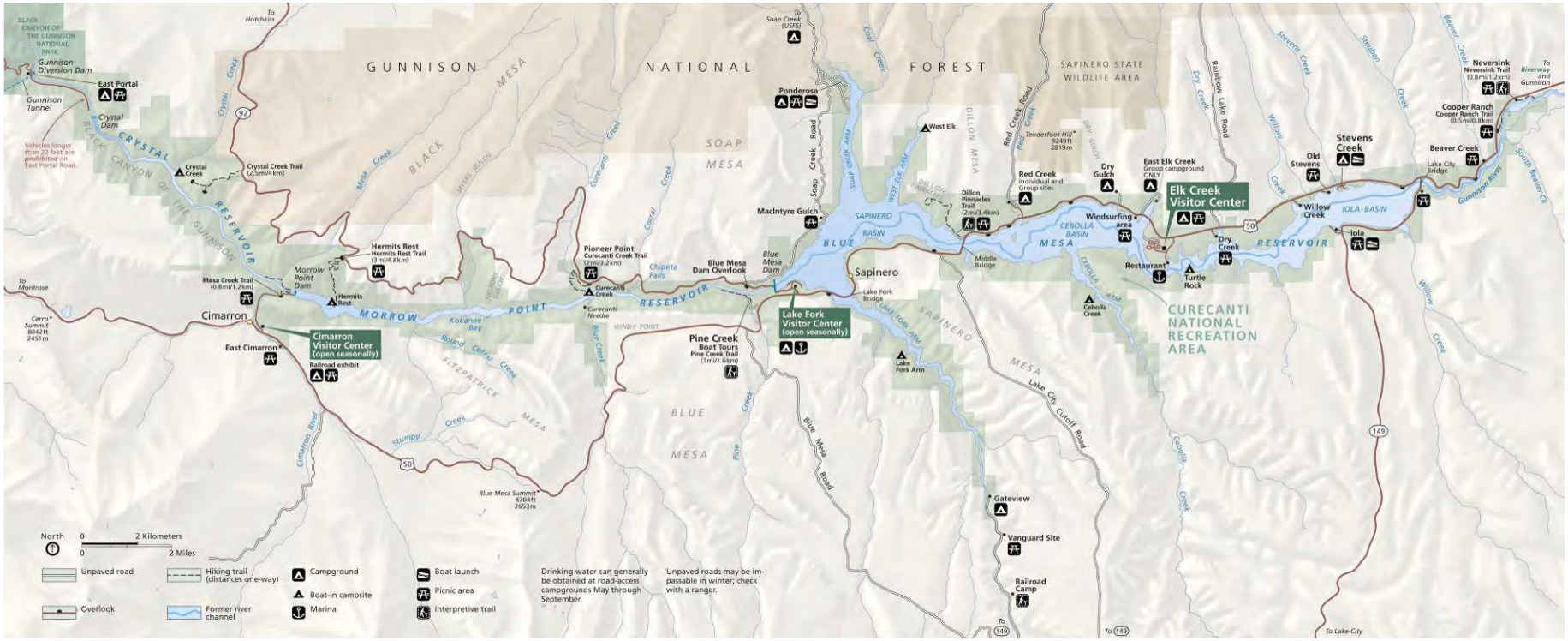
Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of a major mountain-building event known as the Laramide Orogeny (Cather 2003). This event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). The Gunnison uplift, which includes the Black Canyon, is a Laramide structure, although its movements were based on reactivated faults from the Proterozoic tectonic activity (Hansen 1981, 1987a). By the late Eocene, the Gunnison uplift had eroded back into its surroundings (Hansen 1987a).

In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent widespread igneous activity during the late Eocene, Oligocene, and Miocene. Several formations of volcanic ash, known as tuffs, are present within and near CURE, erupted from volcanic calderas to the south. They form the rims of the Black Canyon within CURE. The oldest volcanic rocks belong to the West Elk Breccia, derived from an eruption that took place around 32 Ma (Hansen 1981) in the West Elk Mountains to the north. A series of five tuffs followed from the San Juan Mountains to the south. From oldest to youngest, they are the Blue Mesa, Dillon Mesa, Sapinero Mesa, Fish Canyon, and Carpenter Ridge tuffs, all erupted between approximately 29 and 27 Ma (Olson et al. 1968; Steven and Lipman 1976; Hansen 1981). Volcanic activity dwindled after this time, persisting at a low level. For example, the Mesa Falls Ash, which erupted about 1.2 Ma, and the Lava Creek B Ash, which erupted about 640,000 years ago, are present in the Black Canyon (Sandoval and Karlstrom 2006).

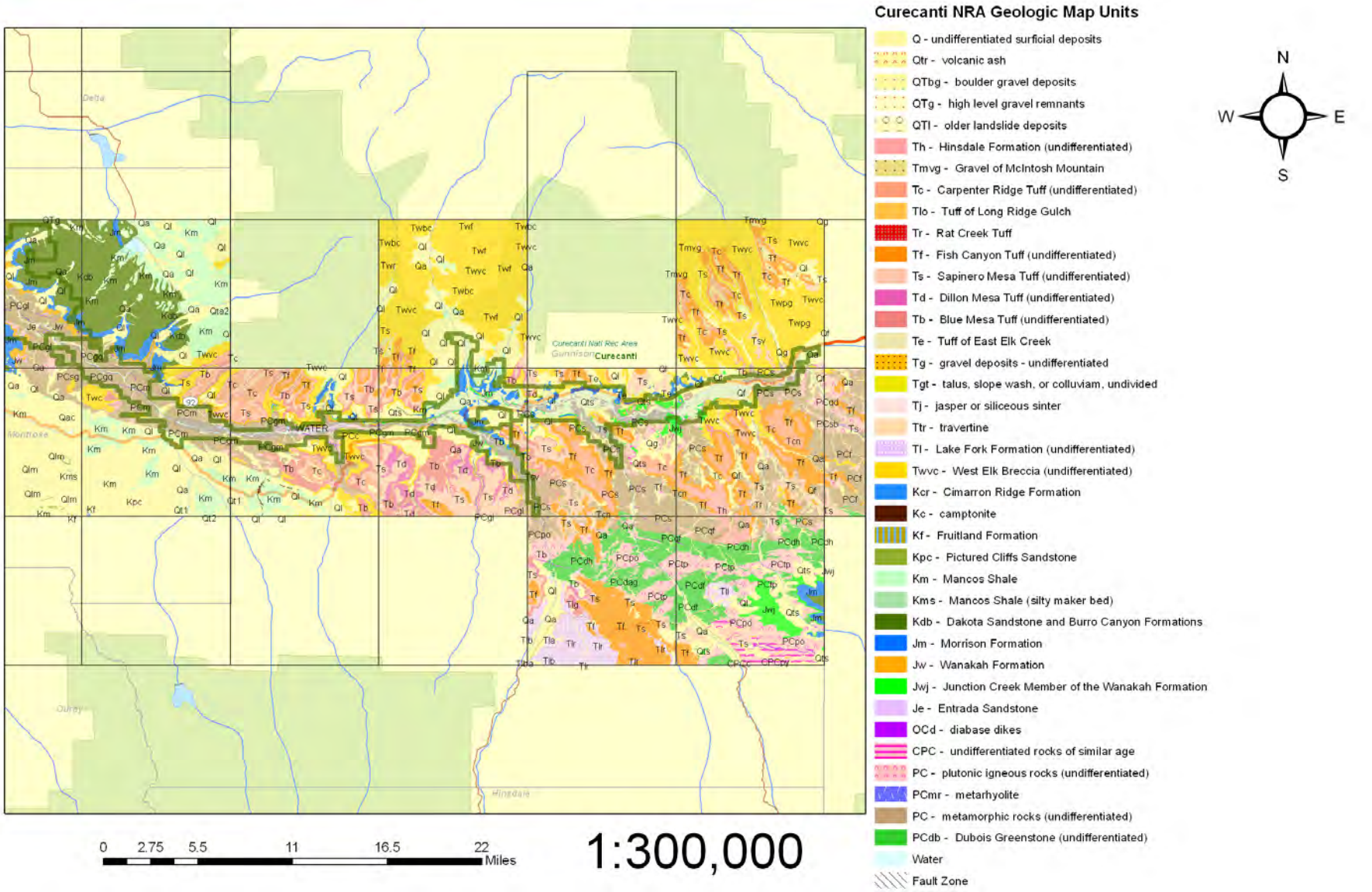
Glaciation occurred in the southern Rocky Mountains during the Pleistocene, which began approximately 2.59 Ma. There is evidence for glaciation on Grand Mesa to the northwest (Henderson 1923). The most recent local episode of widespread glaciation, the Pinedale Glaciation, ended around 14,000 years ago. Glaciation resulted in the formation of terraces on the Gunnison River because of increased sediment load from glaciers eroding mountains in the river highlands (Kellogg 2004). The river began downcutting in the area about 2.5 Ma (Kellogg 2004). Its incongruous presence in the hard crystalline Proterozoic rocks of the Black Canyon is due to being trapped there: it became confined between the thick deposits of volcanic rocks from the West Elk and San Juan volcanic centers and was further restricted by a syncline (concave fold), the axis of which was oriented favorably to the drainage (Hansen 1987a). By the time it had eroded through the Tertiary and Mesozoic rocks and reached the hard crystalline basement, it could not move laterally and was limited to downcutting, which produced the steep walls of the Black Canyon (Kellogg 2004). People arrived in the nearby Uncompahgre River valley by at least between 11,000 and 10,000 years ago, based on the presence of Clovis and Folsom points (Kellogg et al. 2004a).

Rock units exposed at CURE include, from oldest to youngest: a variety of metamorphic and igneous rocks (Proterozoic); diabase dikes (Cambrian or Ordovician); the Wanakah Formation of the San Rafael Group (Middle Jurassic); the Morrison Formation (Upper Jurassic); the Burro Canyon Formation (Lower Cretaceous); the Dakota Sandstone (Lower–Upper Cretaceous); the Mancos Shale (Upper Cretaceous); and volcanic and sedimentary rocks of the West Elk Breccia, Blue Mesa Tuff,

Dillon Mesa Tuff, Sapinero Mesa Tuff, and Fish Canyon Tuff (Oligocene). Surficial units include a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene) (see CURE Figure 3 for a geological map). In general, the Mesozoic formations, which are of greatest interest for paleontological resources, have limited exposures. The Proterozoic rocks form the deepest part of the canyon and the upper reaches are capped by the Oligocene tuffs, leaving a narrow band dominated by the Morrison Formation. Much of this band is covered by Quaternary sediments or is now under water (Hansen 1971; Hedlund and Olson 1973; Hedlund 1974). The Morrison Formation, Dakota Sandstone, and Mancos Shale are known to be fossiliferous at CURE, and the Quaternary deposits are probably fossiliferous. The other Mesozoic units are potentially fossiliferous. The Proterozoic rocks are unlikely candidates for yielding fossils, being either igneous in origin or heavily metamorphosed, and dating from a time before the evolution of hard parts. Fossils have not been reported from any of the Oligocene volcanic units, and they are also unlikely candidates; however, there is some possibility of finding fossils in them, particularly in portions composed of fluvially-reworked tuff (Trujillo 2001) (CURE Table 1). The stratigraphic interval including the Wanakah Formation and the lower Morrison Formation has been of some dispute concerning the placement of unit contacts and the names used. This document will use the nomenclature of Thornberry-Ehrlich (2005), established from the maps cited above; alternative interpretations will be noted in the text.



CURE Figure 2. Geography of CURE (NPS).



CURE Figure 3. Schematic geological map of CURE. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1038963.

Formation	Age	Fossils Within CURE	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Packrat middens are known from several sites either just within or just outside of the recreation area	Alluvial, eolian, fluvial, landslide, and volcanic ash deposits
Oligocene volcanic units	Oligocene	None to date; fossils are unlikely but not impossible	Volcanic ash deposits, some reworked by fluvial processes
Mancos Shale	Late Cretaceous	Plant fragments, bivalves, invertebrate traces, and fragments of fish	Open marine
Dakota Sandstone	Early–Late Cretaceous	Plant fragments, wood, molds of plant and invertebrate fossils, and invertebrate burrows	Terrestrial (especially fluvial) becoming shallow marine over time
Burro Canyon Formation	Early Cretaceous	None to date; fossils known elsewhere include plant fossils, freshwater invertebrates, and freshwater and terrestrial vertebrates	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	Conchostracans, invertebrate burrows and other trace fossils, bones of turtles, crocodile relatives, mammals, and several dinosaur taxa (including a partial skeleton of *Apatosaurus*), and possible mammal burrows	Fluvial, floodplain, and lacustrine settings
Wanakah Formation	Middle Jurassic	None to date; microbial ("algal") limestone (which is common in the Black Canyon of the Gunnison), invertebrates, and invertebrate trace fossils are most likely	Closed marine to eolian sand dunes over time
Diabase dikes	Cambrian or Ordovician	Unfossiliferous (intrusive igneous rocks)	Not applicable (igneous rocks that solidified at depth)
Igneous and metamorphic rocks	At least Paleoroterozoic; Mesoproterozoic for the igneous rocks at Curecanti Needle	None to date; fossils are extremely unlikely because of lithology and metamorphism	Not applicable (igneous rocks that solidified at depth and strongly metamorphosed sedimentary rocks)

CURE Table 1. Concise tabulation of CURE stratigraphy, fossils, and depositional setting; details and references can be found in the text. Igneous and metamorphic rocks are included here for completeness, but are not discussed in detail in the text because of the improbability of fossils.

Sources of Paleontological Resources

The fossils of CURE present opportunities for education, interpretation, and continued or future scientific research in the recreation area. Fossil research in CURE began in the 1990s with Anthony Fiorillo and Cathleen May, who found trace fossils and dinosaur bones. There have been several paleontological surveys of CURE and BLCA since then, including in 2000, 2001, and 2002 by Kelli Trujillo, and in 2005 and 2011 by the National Park Service. The 2005 survey reported 22 fossil sites in CURE (Koch et al. 2006) and the 2011 survey reported an additional six. The Morrison Formation is so far the best unit for fossil discoveries at CURE, yielding 18 sites (Kenworthy 2006; Olstad 2011).

Of interest is the fact that the primary geologic map of the BLCA–CURE area, Hansen (1971), predates the filling of the Blue Mesa Reservoir. This graphically demonstrates a major paleontological resource management concern for CURE: the modern water level falls on outcrops of the Morrison Formation for significant stretches of shoreline. The most fossiliferous unit within the recreation area is directly exposed to the weathering and erosion that come with being on a shore. Furthermore, the

reservoir is also subject to ice and its attendant freeze-thaw cycles. These environmental factors create both a great capacity for destroying fossils and for exposing new material. The other published maps that document most of the rest of CURE, Hedlund and Olson (1973) and Hedlund (1974), also show the pre-reservoir geology, and include indications of the high-water marks. Kenworthy (2006) and Santucci et al. (2009) provide additional information on fossil management at the reservoir sites.

Morrison Formation (Upper Jurassic)

The Morrison Formation is a heterogeneous and variegated unit dated to the middle Late Jurassic, between about 155 and 148 Ma (Foster 2007). The formation is divisible into several members. At CURE, these include from oldest to youngest at least the Salt Wash and Brushy Basin members (Hansen 1965, 1981). As discussed below, the Tidwell Member is also reported as present in recent reports on the geology of the Black Canyon (O'Sullivan 1992; Kellogg 2004), due to a different interpretation of the underlying upper Wanakah Formation. The Salt Wash Member of the area is mostly composed of gray cliff-forming sandstone and red, gray, and green shale, between 35 and 52 m (120 and 170 ft) thick (Kellogg 2004). It often has three sandstone ledges, and becomes thicker down the canyon (Hansen 1965). The Brushy Basin Member is mostly composed of variegated shale and yellow sandstone (Kellogg 2004), up to 105 m (350 ft) thick (Hansen 1981). The Salt Wash Member is interpreted as primarily fluvial in origin, while the Brushy Basin Member derives from the shallow Lake T'oo'dichi' basin and associated drainage (Foster 2003). The climate is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004). The upper contact with the Burro Canyon Formation is a disconformity in the Black Canyon area (Kellogg 2004).

The lower Morrison Formation and underlying Wanakah Formation are a problematic stratigraphic interval on the Black Canyon. The issue revolves around what should be included in the two formations. Historically, authors identified Jurassic eolian (wind-deposited) sandstone below the Morrison Formation as the Junction Creek Sandstone, sometimes as a member of the Wanakah Formation; see for example the geologic maps of Hansen (1971), Hedlund and Olson (1973), and Hedlund (1974). Research published in the 1990s instead included the "Junction Creek Sandstone" of the Black Canyon and the overlying beds of the Wanakah Formation in the Morrison Formation, as the Tidwell Member (O'Sullivan 1992). This has been followed by other workers in the area (Kellogg 2004; Kellogg et al. 2004b). As noted earlier, for consistency with Thornberry-Ehrlich (2005), the older terminology is used here. The situation is further complicated by Lucas et al. (2006a)'s contention that the Summerville and Wanakah formations are synonymous; other workers, including Condon (1992), O'Sullivan (1992, 2004) and O'Sullivan et al. (2006), have found the Wanakah Formation to be older than the Summerville Formation.

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Brachiosaurus*, *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks (Turner and Peterson 1999; Foster 2007). The Morrison Formation of CURE has yielded petrified wood (Kenworthy 2006), rhizoliths (root traces), conchostracans (clam shrimp), unionid clam burrows, crayfish burrows, termite nests, other invertebrate burrows (Fiorillo 1999; Olstad 2011), isolated bones of turtle and crocodile relatives, teeth, bones, and bone molds of at least six dinosaur genera (*Allosaurus*, *Apatosaurus, Camarasaurus*, *Ceratosaurus*, *Stegosaurus*, and *Diplodocus* or *Barosaurus*), possible mammal bones, and large vertebrate burrows (Koch et al. 2006). The Morrison Formation finds in CURE and BLCA are of added interest because of the paucity of Morrison Formation fossils in this area of Colorado (Turner and Peterson 1999; Foster 2003). The only other significant find from the vicinity is another partial *Apatosaurus* skeleton, in this case from low in the formation about 10 km (6 miles) east of Gunnison (Bartleson and Jensen 1988). Both CURE and BLCA were part of the Morrison Formation Extinct Ecosystems Project, a cooperative synoptic survey of this well-known formation (Turner and Peterson 1998; Koch et al. 2006).

Dinosaur bones were first found in the Morrison Formation of the Black Canyon by uranium prospectors in the early 1950s (Jenkins 2004). Fossils from the CURE–BLCA area were not studied in depth until the 1990s, when sites were located in CURE. The work by Fiorillo and May uncovered ichnofossils and the Dino Cove vertebrate locality (Koch et al. 2006). CURE invertebrate trace fossils are best known from sandstones of the lower Brushy Basin Member on Red Creek, and help illustrate how the Morrison Formation includes a mosaic of depositional environments (Fiorillo 1999). Most of the known traces are on large sandstone blocks on the shores of Blue Mesa Reservoir. They have been documented through the use of latex peels (Fiorillo and Harris 2000). Vertebrates are best known from three localities on Blue Mesa Reservoir: Dino Cove, Dinosaur Beach, and Northern Dinosaur Beach (Koch and Zichterman 2006; Koch et al. 2006). Stratigraphically, all three appear to be in the Brushy Basin Member, with the two beach sites slightly lower in section than Dino Cove (Koch et al. 2006). Both beach sites derive from a distinctive conglomeratic layer that includes red and green mudstone fragments, inspiring the informal name "Christmas Conglomerate" (Kenworthy 2006). Dino Cove was discovered in the middle 1990s. It is best known for yielding a partial associated skeleton of *Apatosaurus*, apparently deposited as a floating carcass by a flood and later scavenged (Fiorillo and May 1995; Fiorillo et al. 1996). Other material recovered from this site includes conchostracans, teeth of crocodile relatives and *Allosaurus*, *Stegosaurus* vertebrae, and a possible mammal bone. The specimens are at the Museum of Western Colorado in Grand Junction (Koch et al. 2006). Dino Cove was completely

excavated in the 1990s, with no exposed material remaining (Koch and Zichterman 2006). Dinosaur Beach has yielded generally fragmentary remains from conglomeratic sandstone that is being undercut by the lake (CURE Figure 4) (Koch et al. 2006). Some of the Morrison Formation blocks may be *in situ* (Kenworthy 2006). Dinosaur Beach has yielded remains of turtles, *Allosaurus*, and *Camarasaurus*, as well as a complete ulna (a lower arm bone) from a goniopholidid (closely related to the ancestry of modern crocodilians) (Koch et al. 2006) and recently a possible *Stegosaurus* tail vertebra (J. Foster, Museum of Western Colorado curator of paleontology, pers. comm., November 2011). Large vertebrate burrows have been found just to the west (Koch et al. 2006). These burrows were briefly described by Odier (2006), who assigned them to mammals. Dinosaur Beach holds the potential for further discoveries (Koch et al. 2006).



CURE Figure 4. A bone exposed at Dinosaur Beach (JPK).

Northern Dinosaur Beach is another undercut conglomeratic sandstone. Termite burrows, other invertebrate trace fossils, and teeth and bones from *Allosaurus*, *Apatosaurus, Camarasaurus*, *Ceratosaurus*, *Diplodocus* or *Barosaurus*, and possibly *Stegosaurus* have been found here. Most specimens are fragments, with some teeth, vertebrae, and larger limb bones. Fossils from this site have also been collected, due to the risks from vandalism and intense erosion (Koch et al. 2006). CURE staff and volunteers from the Museum of Western Colorado collected 303 dinosaur bone fragments from this site during October 2005; additional material remains at the site. All of the specimens appear to be in or from float blocks (Kenworthy 2006). Fossils from the two beach sites are in CURE collections (Koch et al. 2006).

The Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites, fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans, ostracodes, and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods (lamp shells), gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006); however, some of the traces, such as those attributed to anthozoans, brachiopods, horseshoe crabs, and echinoderms are only known from marine-influenced units that are not found at CURE (Hasiotis 2004). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians (tuatara), lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods, diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of

herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

Dakota Sandstone (Lower–Upper Cretaceous)

The Dakota Sandstone (also Formation or Group) is a heterogeneous rock unit with outcrops as distant as Minnesota; because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in areas on the western side of the seaway may be inappropriate (Young 1959; Witzke and Ludvigson 1994; Titus et al. 2005). In the Black Canyon, it is composed of sandstone and carbonaceous shale, with minor conglomeratic sandstone, and coal beds near the base (Hansen 1971). It can be 20 to 34 m (65 to 110 ft) thick in the area, and interfingers with the overlying Mancos Shale (Kellogg 2004). More resistant than the Mancos Shale, it caps hills, forms parts of the canyon rims, and produces hogback ridges (Hansen 1965).

On the Colorado Plateau it is common for the Dakota Sandstone to have conglomerates followed by coaly shale and then sandstone at the top. These divisions reflect environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma (see Titus et al. 2005 for discussion of the age). Hansen (1981) gave an age of approximately 105 Ma for the Dakota Sandstone of the Black Canyon.

The fossils of the Dakota Sandstone are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. Koch (2005) reported that four fossiliferous sites had been found in Dakota Sandstone rocks at CURE, yielding plant fragments, wood, a plant mold, shell molds, and invertebrate burrows. Olstad (2011) reported an additional two sites, with fossils including wood molds and the mold of a conifer cone (CURE Figure 5). Similar fossils have been found in the vicinity. Sites along the North Rim at BLCA have yielded fossil wood, other plant fragments, leaf impressions, root traces, and abundant invertebrate burrows (Koch 2005; Kenworthy 2006; Olstad 2011). Coal beds, leaf and twig impressions, and invertebrate trace fossils are present in exposures near the lower Black Canyon (Kellogg 2004). Petrified wood may be found as well (Jenkins 2004). John Foster (pers. comm., November 2011) reported observing petrified logs up to 5 to 6 m (15 to 20 ft) long in this formation on a ranch outside the North Rim. Plant fossils are also known from the Ouray area 50 km (30 miles) to the south (Burbank 1930), and ammonites have been found near Delta, 45 km (30 miles) to the northwest (Young 1959).



CURE Figure 5. A conifer cone from the Dakota Sandstone west of Dry Gulch (CURE locality5GN5841) (photo courtesy Forest Frost, BLCA/CURE).

The best-known terrestrial fossils from the Dakota Sandstone are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Sandstone (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Sandstone of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Sandstone for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (horsetail) (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Sandstone rocks in Utah (including from near Bryce Canyon National Park) reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled

freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Sandstone marine fossils include algae, foraminifera (another type of "shell"-forming protist; Tibert et al. 2003), dinoflagellate cysts (a type of plankton) (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils (such as marks left by the shells of dead ammonites striking the seafloor; Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Sandstone may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera, corals, bryozoans (moss animals), brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Mancos Shale (Upper Cretaceous)

The Mancos Shale is a marine unit largely made up of shale with some sandy, silty, or carbonate-rich layers. It is composed of dark gray silty clay-rich shale with sandstone lenses and siltstone concretions in the Black Canyon. Although the upper portion is not preserved, it is still quite thick in the area, up to perhaps 670 m (2,200 ft) (Hansen 1971). This maximum thickness is found at Cathedral Peak (Hansen 1971), with erosion having destroyed much of that farther east. In eastern CURE, it is less than a tenth of the Cathedral Peak figure where present (Hedlund and Olsen 1973). Most of the exposures in CURE are covered by vegetation (Koch 2005). Easily eroded, the Mancos Shale forms drab badlands of rounded hills in the area (Kellogg 2004). If the Black Canyon section was complete, the Mancos Shale would span on the order of 25 to 30 million years, given the length of time the area was beneath the Western Interior Seaway (Franczyk et al. 1992). Fossils from BLCA and Gunnison Gorge National Conservation Area show that the section only includes the period from about 100 to 85 Ma (Merewether et al. 2006; Ball et al. 2010). Deposition occurred as mud on the floor of the shallow Western Interior Seaway (Hansen 1965).

The Mancos Shale is a widespread formation in the Colorado Plateau and its internal subdivisions differ by location. It is not usually divided on the Black Canyon (Hansen 1971, 1981; Hedlund and Olson 1973; Hedlund 1974; Franczyk et al. 1992; Kellogg 2004; F. Frost [BLCA/CURE archeologist] and A. Mims [BLCA/CURE visual information specialist], pers. comm., February 2011), although a few recent documents have applied the terminology originated in the Mesa Verde area (Merewether et al. 2006; Ball et al. 2010). The members these authors use, from oldest to youngest, are the Graneros Member, the Bridge Creek Limestone Member, the Fairpoint Member, the Blue Hill Member, the Juana Lopez Member, the Montezuma Valley Member, and the Niobrara Member. Additional details on these members can be found in Leckie et al. (1997).

The Mancos Shale is fossiliferous within CURE. Four fossil sites have been found below Dillon Pinnacles and above Cimarron, producing fossils including plant fragments, bivalve shells (CURE Figure 6), invertebrate burrows, and various fish remains including scales, fins, and gill rakers (Koch 2005; Olstad 2011).



*CURE Figure 6. A bivalve (*Ostrea*?) shell from the Mancos Shale at Cimarron (CURE locality 5MN9724) (photo courtesy Forest Frost, BLCA/CURE).*

Diverse marine fossils are known from the Mancos Shale across the Colorado Plateau, with the most abundant being ammonites and bivalves. Marine microorganisms and invertebrates are represented by acritarchs (a "wastebasket" classification for organic microfossils that cannot otherwise be classified), dinoflagellates (Crowley et al. 1994), nannoplankton (Cohen 1959), radiolarians (Olesen 1991), foraminifera (Bagshaw 1977; Leckie et al. 1997), corals (Ball et al. 2010), brachiopods (Leckie et al. 1997), bryozoans (Hill 1982), ammonites, bivalves (Merewether et al. 2006; Ball et al. 2010), gastropods (Leckie et al. 1997), crustaceans (Kues 1980; Scott et al. 2001a), ostracodes (Bagshaw 1977), crinoids (sea lilies) (Leckie et al. 1997), echinoids (sea urchins) (Hill 1982), and worm tubes and other invertebrate trace fossils (Leckie et al. 1997). Marine vertebrates are represented by sharks and other cartilaginous fish (Ekren and Houser 1965; Leckie et al. 1997), teeth, scales, and bones of bony fish (Leckie et al. 1997), turtles (Irby 1995), plesiosaurs (Carter 1991; Spielmann and Lucas 2006), mosasaurs (Kass 1999; Spielmann and Lucas 2006), and sea birds (Lucas and Sullivan 1982). Terrestrial fossils are present near the western margin of the continental seaway, and include root traces, coal, leaves (Hill 1982), ferns, gymnosperms, angiosperms (Young 1987), terrestrial microfossils (spores, and pollen from gymnosperms and angiosperms) (Crowley et al. 1994), crocodilian bones (Irby 1995), dinosaur bones (Hill 1982) and at least one partial dinosaur skeleton (Lucas et al. 2006b). Hundreds of palynomorph (organic microfossil) taxa are known. Thompson (1972) reported 220 taxa of acritarchs, pollen, spores, and dinoflagellates and other algae from the

Mancos Shale in the Mesa Verde area and Cushman (1994, 1996) found 118 taxa of terrestrial plants and 146 taxa of marine palynomorphs in western Colorado.

Fossils are common in the Mancos Shale of western Colorado. Sites are known within and just outside of BLCA, yielding marine invertebrates (Cobban 1973; Scott et al. 2001a; Merewether et al. 2006). Plant remains, a coral, ammonites, bivalves, burrows, fecal pellets, a shark tooth, bones, and fish scales have been recovered from a core taken from western Gunnison Gorge National Conservation Area near Olathe, about 29 km (18 miles) northwest of CURE (Ball et al. 2010). Trujillo (2001) reported observing diverse fossils from the Mancos Shale in the Juana Lopez Member of the Gunnison area, including ammonites, bivalves, and isolated shark teeth and tetrapod limb bones. Lime-rich beds and concretions are fossiliferous in the area (Hansen 1965). A species of mosasaur (*Prognathodon stadtmani*) was named from a partial skeleton discovered near Cedaredge, about 50 km (30 miles) northwest of western CURE (Kass 1999). Mollusks have been found at a number of sites within a similar radius (Cobban 1973; Weimer 1982; Merewether et al. 2006).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary sediments of CURE are a heterogeneous group of deposits left by water, wind, and gravity movements, scattered throughout the recreation area (Hansen 1971; Hedlund and Olson 1973; Hedlund 1974). These deposits are confirmed as fossiliferous at CURE, but there are several sites that may be just within or just outside of the recreation area. Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Quaternary fossils are uncommon within about 100 km (60 miles) of CURE.

Paleoecological materials are the most abundant Quaternary fossils in the area. Packrat middens have been found in the Black Canyon; Anderson et al. (2000) reported middens dated between 5,500 and 6,000 calendar years before present. More detailed information was not provided, but the U.S. Geological Survey/National Oceanic and Atmospheric Administration packrat midden database (http://esp.cr.usgs.gov/data/midden/) has a record of one midden cited to Anderson et al. (2000), with locality information that places it within CURE; it is not clear where this locality information came from. Alison Mims and Forest Frost (pers. comm., February 2011) do not know of any middens from the recreation area. Emslie et al. (2005) reported on 17 middens collected in the upper Gunnison Basin in 1996, with four midden sites (Dillon Pinnacles Site 1, Red Creek Site 2, Two Ravens Cave 2, and West of Haystack Cave) from lands adjacent to CURE. The Gunnison Basin middens are all younger than 3,500 radiocarbon years before present, which may be due to preservation factors (Emslie et al. 2005). Approximate coordinates for the four sites mentioned above (S. Emslie, University of North Carolina, pers. comm., February 2011) place Red Creek Site 2 and Two Ravens Cave 2 just inside CURE boundaries and Dillon Pinnacles Site 1 and West of Haystack Cave just outside. Ancient pollen has been reported in lake sites in the mountains to the north (Fall 1988, 1997; Thompson et al. 2000) and south (Maher 1961; Carrara et al. 1984; Thompson et al. 2000), and Holocene-age peat records have been recovered from fens on Grand Mesa (Sullivan and Gilmore 2006), but such sites are unlikely to be found in CURE given its geography. Packrat middens, which can be found in rockshelters and alcoves, are more likely Quaternary fossils, as well as isolated resistant materials such as wood or teeth in surficial deposits.

The best-known local Quaternary fossil site is Haystack Cave, a cavity in the Sapinero Mesa Tuff about 1 km (less than a mile) north of Blue Mesa Reservoir (Emslie 1986). Sediments in the cave represent both the late Pleistocene and the Holocene, with human artifacts in the upper deposits. Bones from the cave have been dated to between 14,935 ± 610 radiocarbon years before present (19,365–16,880 calendar years before present) and 12,154 ± 1,700 years before present (18,640–10,135 calendar years before present) (Emslie 1986). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate. A wide variety of animal remains have been reported, including bones of frogs, lizards, raptors, ground birds, perching birds, rodents, pikas, rabbits, the extinct American cheetah *Miracinonyx*, mountain lions, coyotes, foxes, badgers, skunks, weasels, bears, horses, deer, pronghorns, bighorn sheep, bison, the extinct muskox *Bootherium*, and the extinct shrub ox *Euceratherium* (Emslie 1986; Scott et al. 2001a). The accumulation probably comes from a combination of denning carnivores, rodent (particularly packrat) nesting, and birds roosting. The site has been damaged by vandals (Emslie 1986).

Recently, a second extensive cave assemblage has been found near Crested Butte, 39 km (24 miles) north of CURE. This site, Crescent Creek Cave, includes the only known high-elevation fauna that spans the last glacial maximum, with material dated from 43,330 to 1,120 radiocarbon years before present. Bones of frogs, snakes, ducks, ground birds, perching birds, bats, rodents, pikas, rodents, dogs, weasels, skunks, bears, horses, bison, bighorn sheep, and *Euceratherium* have been found here. Holocene-age shrews and voles have also been found at the nearby Signature Cave (Emslie 2011), and Pleistocene shrews are reported from Cement Creek Cave, another Crested Butte-area cave (Emslie 2002; Emslie et al. 2005). Another recent discovery, albeit somewhat farther away, is the Snowmastodon site, approximately 75 km (47 miles) north of CURE outside of Snowmass Village. This site, discovered in 2010 and currently within Ziegler Reservoir, yielded several thousand bones deposited in a lake setting between 150,000 and about 45,000 years ago. More than 30 vertebrate taxa are represented, including fish, frogs, salamanders, snakes, lizards, birds, ground

sloths, bats, rodents, rabbits, weasels or minks, otters, Columbian mammoths, mastodons, horses, deer, bison, camels, and others (Stucky et al. 2011a, 2011b). Non-vertebrates are represented by pollen, seeds, cones, leaves, wood (which sometimes bears beaver chewing marks), gastropods, ostracodes, and beetles and other insects (Anonymous 2011; Stucky et al. 2011b).

Another source of Quaternary fossils in the Black Canyon area is from archeological sites. Several such sites are known. An example is the Kezar Basin site just south of CURE, which was used periodically between 8,800 and 5,800 years before present. It appears to have been a game processing site, and pelvic and shoulder girdle elements from deer and pronghorns have been recovered from it. Other local sites are younger (Stiger 2001).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within CURE. However, they are known to preserve fossils elsewhere, and future field investigations within the recreation area may recover fossils from one or more of them.

San Rafael Group: Wanakah Formation (Middle Jurassic)

As noted in the discussion of the Morrison Formation, the components of the upper Wanakah Formation and lower Morrison Formation have been a matter of some disagreement, with newer reports transferring some of the Wanakah Formation to the Morrison Formation, including the Junction Creek Sandstone Member.

The Wanakah Formation is divisible into multiple members at some locations (Goldman and Spencer 1941; Steele 1985; O'Sullivan 1992, 2004). Aside from the Junction Creek Sandstone Member, an underlying Pony Express Limestone Member is often identified in the Black Canyon (Hansen 1965, 1971, 1981; O'Sullivan 1992), but it doesn't persist east of Dead Horse Mesa (Hansen 1971). Using the older definition, the Wanakah Formation of the Black Canyon includes up to four divisions. Removing from consideration the Pony Express Limestone, the lowest part of the formation is composed of gypsum and gypsum-rich sandstone and mudstone, up to 44 m (145 ft) thick in BLCA but becoming thinner and less gypsum-rich to the east in CURE. Above this is the Junction Creek Sandstone Member, a discontinuous gray, yellow, or pink dune sandstone up to 27 m (90 ft) thick. Finally, there is an upper part composed of gray silty mudstone and microbial chert-rich limestone, between 8 and 18 m (25 and 60 ft) thick (Hansen 1971). Under the more recent definition, the Junction Creek Sandstone and the overlying beds are the Tidwell Member of the Morrison Formation, and the Junction Creek Sandstone is renamed the informal "beds at Smith Fork" (Kellogg 2004). The lower Wanakah Formation is interpreted as deposition from a shallow closed seaway. The hot climate promoted evaporation and the formation of gypsum beds. The overlying sandstone beds are dune sands deposited on the floor of the basin following the seaway drying up (Kellogg 2004). The formation dates to the end of the Middle Jurassic (Callovian stage) (Scott et al. 2001b), about 165 to 160 Ma.

The Wanakah Formation is not known for its fossils, and it is not known to be fossiliferous within CURE (Trujillo 2001). However, invertebrate burrows are known from the formation at BLCA (Koch 2005; Olstad 2011). Additionally, the microbial limestone of the upper section is a commonly-noted feature of this formation on the Black Canyon (Hansen 1965, 1968; Kellogg 2004). The limestone can be identified by wavy microbial structures (Kellogg 2004). The presence of microbial limestone has not been confirmed within CURE, and this section of the formation is among the beds that have commonly been assigned to the lower Morrison Formation in recent years. Hansen (1965) also reported bivalves and gastropods in the microbial limestone. Reports from other areas of fossils in the upper section include root traces (Scott et al. 2001b), possible sponge spicules, bivalves, gastropods, ostracodes, indeterminate fossils, possible worm tubes and other invertebrate trace fossils (O'Sullivan et al. 2006), and possible pterosaur (flying reptile) feeding traces (Trujillo et al. 2004).

For reference, Summerville Formation fossils include plant debris (Petersen and Pack 1982), a single sauropod tail vertebra (Lucas and Heckert 2000), invertebrate traces (Lockley and Wright 2003), and a variety of vertebrate traces including pterosaur tracks (Lockley et al. 1995; Lockley and Wright 2003; Mickelson et al. 2004), theropod tracks (Lockley et al. 1996), and sauropod tracks from near the top of the Summerville Formation at Moab (Lockley 1991). If the Tidwell Member is included in the Summerville Formation, as suggested by Lucas et al. (2006a), its fossil assemblage is broadened to include root traces (Kjemperud et al. 2008), algal and microbial laminations, charophytes, gastropods, ostracodes (Scott et al. 2001b), burrows (Kjemperud et al. 2008), horseshoe crab traces (Scott et al. 2001a), a lungfish tooth plate (Imhof and Trujillo 2005), indeterminate reptile bones (the putative pterosaur *Utahdactylus*; Bennett 2007), the poorly-known sauropod *Dystrophaeus* (Gillette 1996; Lucas and Anderson 1997), and pterosaur and dinosaur tracks (Lockley et al. 1998).

Burro Canyon Formation (Lower Cretaceous)

The Burro Canyon Formation of CURE and its surroundings is poorly exposed and difficult to distinguish from the overlying Dakota Sandstone (Hansen 1965), with which it is often mapped. One area it can be seen in CURE is on the north shore of Blue Mesa Reservoir (Jenkins 2004). It is mostly composed

of cliff-forming light gray conglomeratic sandstone and pebble conglomerate; some shale beds are also present (Hansen 1971). Chert pebbles are common, especially near the top of the unit (Kellogg 2004). It is less than 35 m (115 ft) thick (Hansen 1981). In this area, it is interpreted as the deposits of meandering streams flowing across broad alluvial plains (Hansen 1965). Sediments were deposited under a semi-arid climate, probably reworking material from the Morrison Formation (Walters et al. 1987). The predominant paleoflow was to the north (Craig 1981). The upper contact with the Dakota Sandstone is conformable (Harris et al. 2004).

The age of the Burro Canyon Formation is uncertain. Hansen (1981) gave an age of about 110 Ma for the Burro Canyon Formation in the Black Canyon. It is usually thought of as having been deposited between the Barremian stage of the Early Cretaceous (130 to 125 Ma) and the onset of Dakota Sandstone sedimentation around 100 Ma. However, the base of the formation has been reported to intertongue with the Brushy Basin Member of the Morrison Formation in places, which would imply a Late Jurassic age for part of the formation (Ekren and Houser 1959; Dickinson and Gehrels 2008). It should be noted that the Burro Canyon Formation and Brushy Basin Member can be difficult to distinguish. Mudstones of the Burro Canyon Formation weather as fragments, while Brushy Basin Member mudstones form a hard frothy crust (Ekren and Houser 1959). The Burro Canyon Formation and Cedar Mountain Formation of Utah are lateral equivalents. Generally, the Burro Canyon Formation is present east of the Colorado River, and the Cedar Mountain Formation is present west of it (Dickinson and Gehrels 2008).

The Burro Canyon Formation is not known to be fossiliferous within CURE, but has some potential (Trujillo 2001). In general, the fossils of the Burro Canyon Formation are not well known at this time. The current fossil assemblage from the formation over its entire area of outcrops includes organic debris, lake-dwelling algae, palynomorphs including conifer and angiosperm pollen (Tschudy et al. 1984), fern leaves (Simmons 1957), *Tempskya* (Dayvault and Hatch 2005), cycadeoids (Tidwell et al. 1976), conifers (Simmons 1957), early flowers (Fishell 1981; Armstrong and McReynolds 1987), plant fragments (Stokes 1952), wood fragments (Tschudy et al. 1984), freshwater bivalves (Reeside 1957), gastropods, freshwater ostracodes, ganoid fish scales (Simmons 1957), and dinosaur bones (Trujillo et al. 2004), including a theropod femur and claw (Armstrong 2010). Additionally, a site in Utah near the border with Colorado (the Hotel Mesa Quarry) has yielded fossils of sharks, bony fish, lungfish, various crocodile relatives, theropods, the sauropod *Brontomerus*, and ornithopods (beaked bipedal dinosaurs) (Taylor et al. 2011). It has been attributed to both the Burro Canyon Formation (Kirkland and Madsen 2007) and the Ruby Ranch Member of the Cedar Mountain Formation; by geography, it is within the Burro Canyon Formation (Taylor et al. 2011).

Park Collections

CURE has a number of fossils in its collections, all housed at a storage facility at Cimarron, Colorado. Eleven unidentified bone fragments collected in 2002, with collection numbers CURE 17865–17869, are catalogued under the accession number 171. Several hundred specimens collected during the 2005 field season, with collection numbers CURE 17870–17975, as well as CURE 17976 (collected by Tyra Olstad from locality 5MN9724 in July 2011), are catalogued under the accession number 172. Some of the specimens, from the two beach sites, were previously known under field numbers. They include: partial vertebrae from diplodocid sauropods (CURE 17882, 17885, 17893, and 17906, formerly known as CURE 001 and 002); *Camarasaurus* tooth crowns (CURE 17881, 17898, and 17969, formerly CURE 003, 011, and 012); a *Ceratosaurus* tooth (CURE 17919, formerly CURE 004); a diplodocid tooth (CURE 17875, formerly CURE 005); an ulna (lower arm bone) from a crocodile relative (CURE 17874, formerly CURE 006); and *Allosaurus* hand and vertebral bones (CURE 17870, 17871, 17879, and 17884, formerly CURE 007, 008, 009, and 010) (Koch et al. 2006; Olstad 2011). Dino Cove bones are kept at the Museum of Western Colorado in Grand Junction, with specimens including MWC 5140, the *Apatosaurus* partial skeleton, and MWC 5525, two *Stegosaurus* tail vertebrae (Koch et al. 2006). A small amount of Dino Cove material was sent to the Dinosaur Depot in Cañon City, Colorado for preparation and exhibition (Engard 2003; Kenworthy 2006), but was returned to the Museum of Western Colorado in 2007 (P. Monaco, Dinosaur Depot, pers. comm., April 2011). Additionally, the Academy of Natural Sciences in Philadelphia, Pennsylvania was involved in the 1996 excavation at Dino Cove. It temporarily curated specimens from the site and may have briefly exhibited some (Kenworthy 2006).

Cultural Resource Contexts

There are not yet any records of fossils found in cultural resource contexts at CURE, but there is the potential for such finds. As noted earlier, people have been present in the area for millennia, having arrived in the Uncompahgre River valley between 11,000 and 10,000 years ago at least (Kellogg et al. 2004a). A concentration of Paleo-Indian sites (12,000 to 7,500

years ago) may exist along the Gunnison River east of Blue Mesa Dam, and the present area of the Blue Mesa Reservoir was "a high-use base camp area for aboriginal hunting and its associated activities (tool production and food processing)" (Moeller et al. 1993). Animal remains have been found at some of these sites, such as the Kezar Basin site mentioned above, and in some cases date to at least the Middle Holocene (Stiger 2001). Tools also represent a potential crossover between cultural and paleontological resources. It is not uncommon for people to use fossiliferous materials in tools. In the Black Canyon region, people used chert from the microbial limestone of the upper Wanakah Formation (or Tidwell Member of the Morrison Formation) for tools (Kellogg 2004). Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division produced a geologic resource evaluation report for CURE and BLCA in 2005, recommending that paleontological research projects be encouraged at the two units (Thornberry-Ehrlich 2005). This recommendation was echoed by Koch et al. (2006). Additional recommendations were provided by Koch (2005) and Kenworthy (2006).

- CURE reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 22 localities. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009), including some discussion of CURE sites on Blue Mesa Reservoir. One of the major monitoring issue at CURE is the presence of fossil sites on the reservoirs, which exposes fossils to fluctuating water levels and intense freeze and thaw cycles. Fossil theft and vandalism are potential threats as well; for example, nearby Haystack Cave has been vandalized (Emslie 1986).

- Recreation area staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, recreation area staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- The large vertebrate burrows from the Morrison Formation are worthy of further research.

- The collections should be fully cataloged, and plans for storage should be improved.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the recreation area and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Dinosaur National Monument

Dinosaur National Monument (DINO), Colorado and Utah, was established to protect Dinosaur Quarry (also known as Carnegie Quarry or Douglass Quarry), one of the richest known localities for dinosaur fossils in the world. It also contains a geologic record from the Precambrian to present, and has outstanding scenery and many archeological sites (DINO Figure 1). DINO was proclaimed an national monument October 4 1915. It has had several boundary changes: July 14, 1938; September 8, 1960; February 21, 1963; October 9, 1964; and November 10, 1978. DINO currently encompasses 85,096.31 ha (210.277.55 acres), all but 1,858.32 ha (4,592.01 acres) of which are under federal administration (DINO Figure 2).

Geologic Background

DINO encompasses an area near the northwestern corner of Colorado and the northeastern corner of Utah. Most of its land is in western Moffat County, Colorado; the rest of the monument, including Dinosaur Quarry, is in northeastern Uintah County, Utah. The eastern end of the Uinta Mountains passes through northern DINO, and the southern part of the monument is in the northern Uinta Basin. Geographically, DINO encompasses a three-pronged area, with river drainage and an associated canyon in each prong. The Green River enters through the northern prong and the Yampa River enters through the eastern prong. The two rivers join course in central DINO at Echo Park and exit through the western prong. The nearest population center is Vernal, Utah, which is about 16 km (10 miles) west of western DINO. Structurally, DINO includes several large-scale anticlines (convex folds), monoclines (step-like folds), synclines (concave folds), and faults, mostly with east-west axes (Gregson et al. 2010).



DINO Figure 1. DINO is marked by canyons cut by the Green–Yampa river system (TC).

DINO is documented extensively in the paleontological and geological literature, not only because of its dinosaur fossils, but also because it has excellent exposures of a number of rock formations, and a wide variety of geologic structures. A selection of DINO-specific geological publications includes: Untermann and Untermann (1949a, 1949b, 1950, 1954, 1969); Stokes (1949); Good et al. (1958); Bilbey et al. (1974); Sumison (1976); Lawton (1977); Fiorillo (1994); Chure (1987, 1989); Schaffer (1987); Chure and McIntosh (1990); Elder (1999); Graham (2006); Erickson (2007); and Gregson et al. (2010).

Publications that are not specifically about DINO geology or paleontology but devote attention to these topics include: Poole and Stewart (1964a, 1964b); Bissell (1964); Mayer (1964); Schell and Yochelson (1966); Stewart et al. (1972a); Fryberger (1979); Dodson et al. (1980); Herr and Picard (1981); Herr et al. (1982); Driese (1985); Dubiel (1992); Santucci et al. (1998); Turner and Peterson (1999); Scott et al. (2001a); Foster (2003); Hunt et al. (2006); Lucas et al. (2006a); and Santucci and Kirkland (2010).

The paleontology of DINO has been the subject of hundreds of publications. Considering only those publications that are mostly or wholly dedicated to DINO fossils, and which are not abstracts, a partial list includes the following: Holland (1915, 1919, 1924a, 1924b); Gilmore (1916, 1925a, 1925b, 1926, 1932, 1936); Ellinger (1950); White (1958); McIntosh and Berman (1975); Shepherd et al. (1977); Berman and McIntosh (1978, 1986); Gaffney (1979); Chure (1981, 1994, 2000a); Galton (1982); McGinnis (1982); Lockley et al. (1990, 1992, 1993); Sharpe (1991, 2002); Engelmann (1992, 2004); Hubert and Chure (1992); Hunt et al. (1993); Carpenter (1994); Chure et al. (1994, 2010); Madsen et al. (1995); Hubert et al. (1996); Chure and Evans (1998); Engelmann and Callison (1998, 1999); Evans and Chure (1998a, 1998b, 1999); Fraser and Wu (1998); Henrici (1998a, 1998b); Jones et al. (1998); Hasiotis et al. (1999); Koch and Santucci (2002); Evans et al. (2005); Lockley (2006); Carpenter and Wilson (2008); Whitlock et al. (2010); Anderson et al. (2011); and McDonald (2011).

In addition, the DINO area has had extensive geological mapping at the 1:24,000 scale. These geological maps include: Hansen (1977a, 1977b, 1977c); Rowley and Hansen (1979); Rowley et al. (1979a, 1979b, 1981); Hansen and Carrara (1980); Hansen and Rowley (1980a, 1980b); Hansen et al. (1980, 1982); and the summary map Hansen et al. (1983). Larger-scale maps include Rowley et al. (1985) and Sprinkel (2006, 2007). The Geologic Resources Division of the National Park Service (NPS) coordinated a geologic resource workshop for DINO during August 1998 (NPS 1998) and produced a geologic resource evaluation report for the monument in 2006 (Graham 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The geologic record exposed at DINO is one of the most extensive in the NPS system. It begins around 770 Ma (million years ago) and continues with relatively few breaks until the present. The major hiatuses are the end of the Precambrian, part of the early Paleozoic (Ordovician to Devonian), and the end of the Cretaceous and most of the Paleogene (see the appendix for a geologic time scale). Much of the latter hiatus is bridged by rocks just outside of the monument.

The Uinta Mountain Group, the oldest rocks at DINO, date to the Neoproterozoic, roughly 770 to 740 Ma (Dehler et al. 2007). At this time the craton Laurentia, the future continent of North America, was located near the equator as part of the supercontinent Rodinia (Condie et al. 2001). Near what is now western North America were the landmasses that became Antarctica and Australia (Dehler et al. 2010). A depositional basin existed where the Uinta Mountains exist today (Hansen 1986a), perhaps formed in response to the stresses that later caused the breakup of Rodinia between 720 and 670 Ma (Dehler et al. 2010), or as an artifact of the assembly of the craton (the area is near a zone where two ancient landmasses were joined; Condie et al. [2001]). A shallow sea was present to the (modern) west, between North America and the other landmasses (Condie et al. 2001), which transgressed over the land during this time (Dehler et al. 2010). The Uinta Mountain Group appears to be related to the similar Chuar Group of the Grand Canyon and Pahrump Group of Death Valley (the ChUMP hypothesis, first proposed by Dehler et al. [2001]).

A depositional hiatus on the order of 200 million years followed the Uinta Mountain Group at DINO. It ended during the Late Cambrian, when a marine transgression led to the deposition of the Lodore Formation (Herr and Picard 1981). Much of the rest of the first half of the Paleozoic (Ordovician, Silurian, and Devonian) is not represented at DINO, but there may be some Devonian rocks in the canyons of Lodore and Jones Hole, in the form of "irregularly bedded gray sandstone at the top of the Lodore Formation" (Hansen 1977a); these rocks may belong to the Chaffee Formation (Hansen 1977a; Herr and Picard 1981). Mississippian deposition at DINO began with a marine transgression from the west that brought a shallow, warm sea over the area. The sea then retreated about halfway through the Mississippian, but was followed by another transgression-regression cycle during the second half of the period. The first cycle produced the Madison Limestone, and the second produced the Humbug and Doughnut formations (Mayer 1964). A third cycle spans the Pennsylvanian at DINO. In this episode, the Round Valley Limestone, Morgan Formation, and Weber Sandstone record shifting deposition from shallow marine to eolian dunes (Driese and Dorr 1984; Broadhead and Driese 1994).

Several more marine transgressions are recorded in the Permian, Triassic, and Jurassic rocks of the Uinta Mountains region (Osmond 1965). During the latter part of the Paleozoic, northeastern Utah and northwestern Colorado were part of a structural trough, with uplifts to the east (the ancestral Front Range) and the south (the ancestral Uncomphagre uplift) (Johnson et al. 1988). The uplifts provided sediment for the next group of formations (Osmond 1965). The Permian Park City Formation represents a complex series of marine transgression-regression cycles of various scales, under a hot, dry climate (Peterson 1984).

The overlying Early Triassic-age Moenkopi Formation is a series of coastal redbeds (Rowley et al. 1979b). A shoreline running southwest-northeast was present in the area, with the sea on the west and land on the east (Van Deventer and Picard 1975). The top of the Moenkopi Formation is separated from the Chinle Formation by millions of years, with intervening erosion (Pipiringos and O'Sullivan 1978). The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time (Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003). Chinle Formation rocks show a gradually drying climate over time: during its deposition, the Colorado Plateau moved north from near the equator to arid-zone latitudes. What had been a setting of rivers, lakes, and deltas turned arid (Tanner 2003), as seen in the thick dune sandstones of the Nugget Sandstone.

The Early to early Late Jurassic of the DINO area is represented by four formations that alternated between expansive eolian dune fields (ergs) and coastal to shallow marine deposition. From oldest to youngest, they are the Nugget Sandstone (eolian), Carmel Formation (shallow marine), Entrada Sandstone (eolian), and Stump Formation (shallow marine) (Gregson et al. 2010). The shoreline of the marine body responsible for the Carmel Formation ran through the DINO area. Western DINO had predominantly marine to marginal marine deposition, while eastern DINO had marginal marine and terrestrial deposition (Rigby 1964; D. Sprinkel, Utah Geological Survey geologist, pers. comm. January 2012). The succeeding Entrada dunes made up the last major erg of the region (Loope 2008); in the DINO area, the coast remained nearby to the west (Crabaugh and Kocurek 1993), and occasionally transgressed into the dune field (Gregson et al. 2010). The overlying Stump interval represents the last major marine expansion during the Jurassic in the western United States (Eicher 1955; Hoggan 1970).

Following the deposition of the Chinle Formation, the Colorado Plateau had had a generally arid climate, but that changed with the gradual northward movement of the North American continent. Southern Utah reached 22° N during this time (Loope 2008). Aridity decreased through the Late Jurassic (Lockley and Wright 2003). The more humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the last geologic formation of the Jurassic in DINO, the well-known Morrison Formation. By the time it was

deposited, the area had moved out of the low subtropical latitudes (Lockley and Mickelson 1997). The Morrison Formation is interpreted as a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004). During deposition of the upper portion of the formation (the Brushy Basin Member), a vast basin is thought to have extended from near Grand Junction, Colorado, to Albuquerque, New Mexico. This basin was periodically occupied by a shallow saline lake which has been named Lake T'oo'dichi'; the setting is described as analogous to the modern Lake Eyre basin in Australia (Turner and Fishman 1991). At DINO, the early Morrison Formation included fluvial deposits southeast-flowing rivers derived from highlands in western Utah and northeastern Nevada, along with some eolian deposition (Hubert and Chure 1992). The later part of the Morrison Formation at the monument was deposited in a braided river system (Dodson et al. 1980). After another depositional hiatus, the rock record of DINO resumes in the Early Cretaceous with the Cedar Mountain Formation, which is very similar to the Morrison Formation in lithologies and depositional settings (Gregson et al. 2010). During the deposition of this formation, a transition began from internal drainage and low sedimentation, as seen in the upper Morrison Formation, to greater sedimentation in a foreland basin near a sea, as seen in the overlying Dakota Sandstone (Eberth et al. 2006).

Terrestrial processes continued until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Sandstone signals the onset of several cycles when the Western Interior Seaway transgressed and regressed over the land (Aubrey 1991; Ulicny 1999; Currie et al. 2008). Six major cycles occurred from the late Early Cretaceous until nearly the end of the Cretaceous in northeastern Utah and northwestern Colorado (Cole 1987), but not all of them are represented at DINO. The Seaway eventually extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Marine deposition at DINO from the Seaway is represented by the Mowry Shale, some of the Frontier Formation, and Mancos Shale (Gregson et al. 2010), although the record is not as complete as in nearby areas because the development of a submarine high in the Vernal area interrupted deposition between the Mowry and Frontier formations (Molenaar and Wilson 1990). During part of this absent time, the DINO area was probably an island or offshore shoal. A few million years later, by the time of the Frontier Formation, the offshore high had been connected to the mainland, leading to a coastal plain setting (Franczyk et al. 1992). The resulting cape is known as the Vernal Delta, a complex of smaller deltas that projected as far west into the Western Interior Seaway as the DINO area (Ryer and Lovekin 1986). Another marine transgression then submerged the area for much of the rest of the Late Cretaceous (Franczyk et al. 1992).

Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of a major mountain-building event known as the Laramide Orogeny (Cather 2003). This event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Uinta Mountains uplifted as a large, generally east-west anticlinal fold during the latest Cretaceous phase of the Laramide Orogeny (Hansen 1986a, 1986b; Franczyk et al. 1992), with separate domes for the eastern and western Uintas (Hansen 1984). The eastern Uinta Mountains had been taller than the western mountains, but collapsed beginning in the late Oligocene and continuing in the Miocene (see below) (Hansen 1984). The Laramide Orogeny ended during the early Oligocene for the eastern Uinta Mountains (Hansen 1984). The Uinta Basin formed south of the mountains during the early Paleogene (Osmond 1965) and also experienced uplift: its floor rose from 300 m (1,000 ft) above sea level during the Eocene to 1,200 m (3,900 ft) by the middle Oligocene (Hansen 1986b).

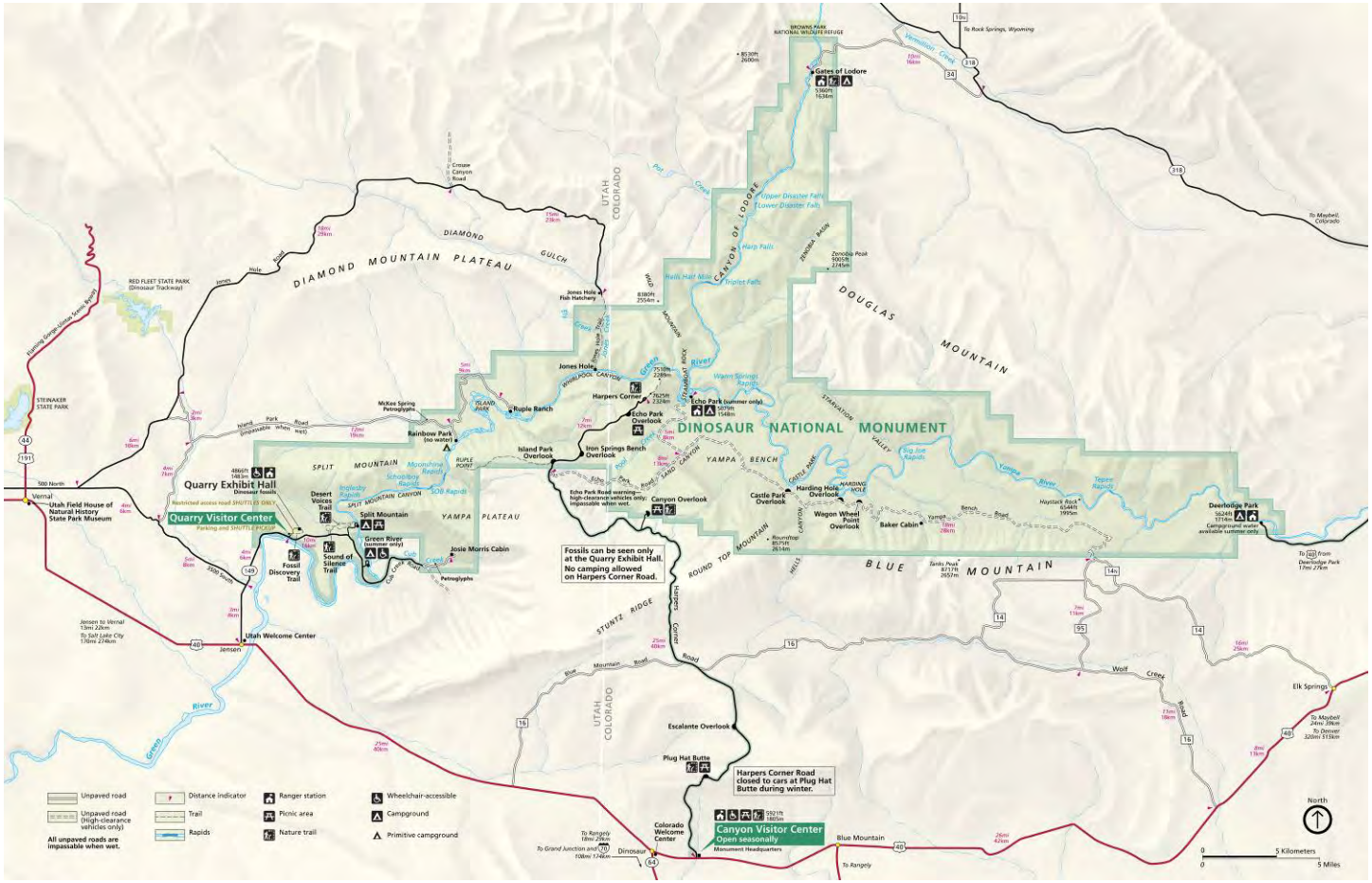
Cenozoic formations predating the Oligocene are not present at DINO, but are widely exposed just outside of the monument to the south and west in the Uinta Basin (Untermann and Untermann 1949a; Rowley et al. 198;5 Sprinkel 2006, 2007; Gregson et al. 2010), a location well-known for its Eocene lake deposits (Franczyk et al. 1992; Gregson et al. 2010). The effects of Paleogene tectonic activity are pervasive in the region. Tectonic activity persisted in the Uinta Mountains for much of the Cenozoic, with only brief stable periods. Episodes of tectonic stability during the Oligocene led to the formation of broad, gently inclined erosional surfaces called pediments around the Uinta Mountains. Two of these surfaces, the older Wild Mountain surface and the more recent Gilbert Peak surface of the Early Oligocene, can be seen within DINO. Alluvial fans formed on the Gilbert Peak surface and eventually coalesced to form a bajada, resulting in the thick Bishop Conglomerate that can be found in many areas of DINO. Resumption of tectonic activity ended deposition of the Bishop Conglomerate. The new activity was extensional, and caused the entire eastern half of the Uinta Mountains to collapse; the south flank tilted north, the north flank warped and tilted south, and the entire complex subsided. In turn, the drainage patterns of the region were completely altered. Before the eastern Uintas collapsed, flow was divided by the range. During the collapse, drainage began to flow down the former crestline, and streams that had been flowing away from the mountains switched direction, setting the conditions that allowed the Yampa River to expand northward and capture the Green River during the Pleistocene (Hansen 1986b). Several ash beds within the Bishop Conglomerate in and around DINO have been dated to 30 Ma and have been geochemically identified with volcanic eruptions from eastern Nevada and western Utah (Kowallis et al. 2005). The capture of the Green River probably occurred around 600,000 years ago. Previously, the Green River had flowed east, to the Mississippi (Hansen 1985).

Uplift has continued episodically since the Laramide Orogeny. The Elkhead Mountains northeast of DINO and White River Plateau southeast of DINO are two nearby areas with uplift of over 600 m (2,000 ft) over the past 23 million years (Reheis et al. 1991). In addition to periods of uplift, volcanic activity became widespread in the western United States during the middle Cenozoic. Volcanic activity began in the Great Basin by the late Eocene and persisted into the Miocene (Hansen 1986b). Eruptions from multiple volcanic centers contributed ash to the Oligocene and Miocene rocks of the DINO area (Luft 1985). Faulting, uplift, and volcanism have continued into the Quaternary (Hansen 1986b).

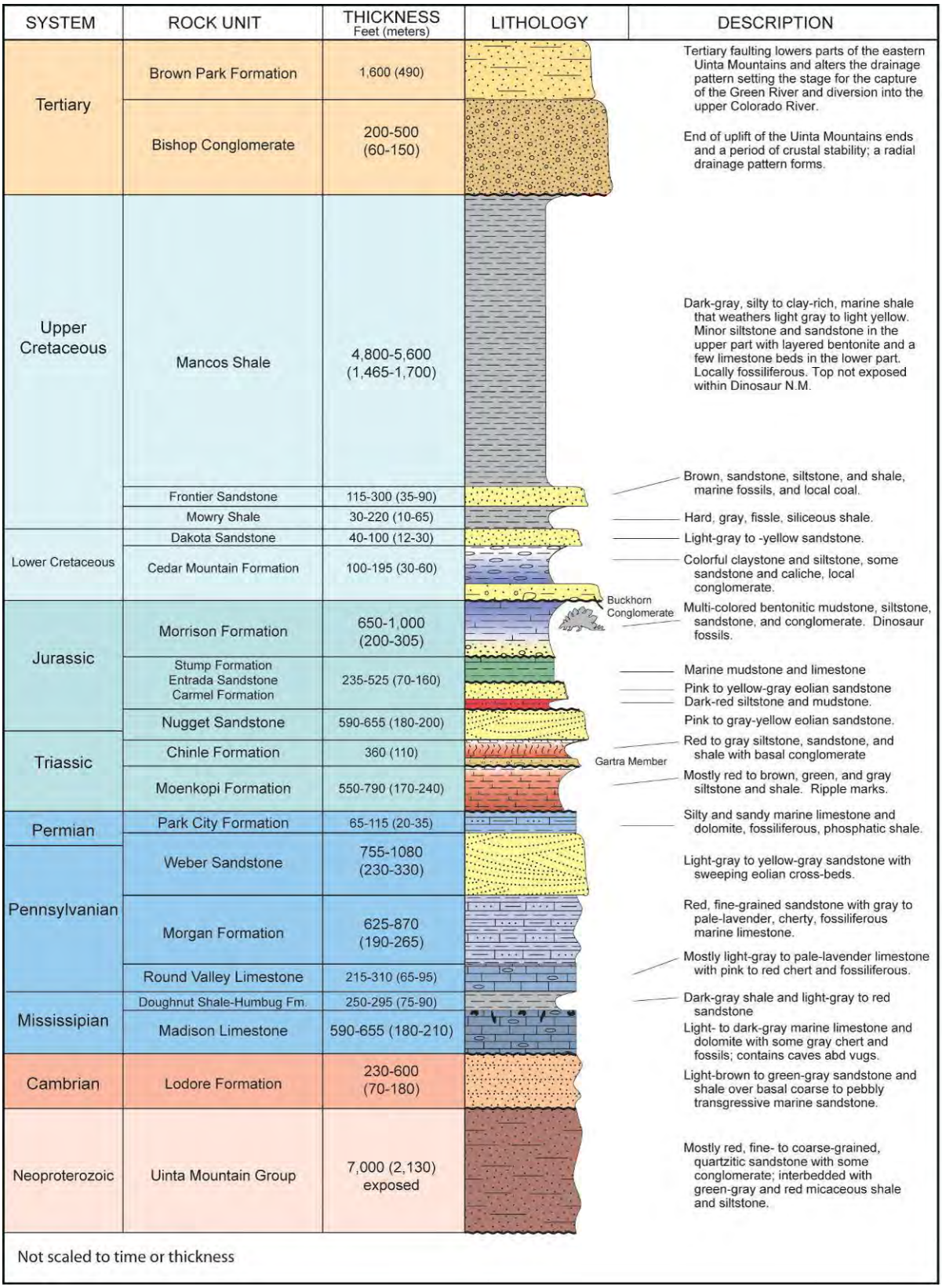
Glaciation affected the Uinta Mountains during the Pleistocene, but it did not reach as far east as DINO (Untermann and Untermann 1954; Munroe and Laabs 2009). Plant fossils from DINO just after the end of the most recent glacial episode show that area temperatures have increased and precipitation has decreased over the past 10,000 years. The modern pinyon-juniper woodland appeared in the past few thousand years (Sharpe 2002). People have been in the Uinta Mountains since the late Pleistocene (Bilbey 1992a), and in the DINO area for the past several thousand years (Breternitz 1970).

Although the geology of DINO has been studied extensively, there have been many variations in the nomenclature used for the units, due to causes both natural (location adjacent to a mountain range, which hinders correlation) and artificial (location near the borders of three states, which have their own nomenclatures). The names used by Graham (2006) are followed here for consistency between NPS products, with the exception of the Glen Canyon Sandstone, which is now designated the Nugget Sandstone by most researchers (Sprinkel 2006, 2007; Gregson et al. 2010; Sprinkel et al. 2011a). Other alternate names will be discussed in the appropriate locations. Rock units exposed within DINO include, from oldest to youngest: the Uinta Mountain Group (Neoproterozoic); the Lodore Formation (Upper Cambrian); an upper Cambrian or Lower Ordovician igneous dike; the Madison Limestone (Lower–Middle Mississippian); the Humbug Formation (Middle Mississippian); the Doughnut Shale (Middle–Upper Mississippian); the Round Valley Formation (Lower Pennsylvanian); the Morgan Formation (Middle Pennsylvanian); the Weber Sandstone (Middle Pennsylvanian–Lower Permian); the Park City Formation (Lower–Middle Permian); the Moenkopi Formation (Lower Triassic); the Chinle Formation (Upper Triassic); the Nugget Sandstone (Upper Triassic–Lower Jurassic); the Carmel Formation and Entrada Sandstone of the San Rafael Group (Middle Jurassic); the Stump Formation and Morrison Formation (Upper Jurassic); the Cedar Mountain Formation and Dakota Sandstone (Lower Cretaceous); the Mowry Shale, Frontier Formation, and Mancos Shale (Upper Cretaceous); the Bishop Conglomerate (Lower Oligocene); and the Browns Park Formation (Miocene). Surficial deposits (Pleistocene–Holocene) are found in many areas (see DINO Figure 3 for a stratigraphic column and DINO Figure 4 for a geological map) (Hansen et al. 1983; Graham 2006).

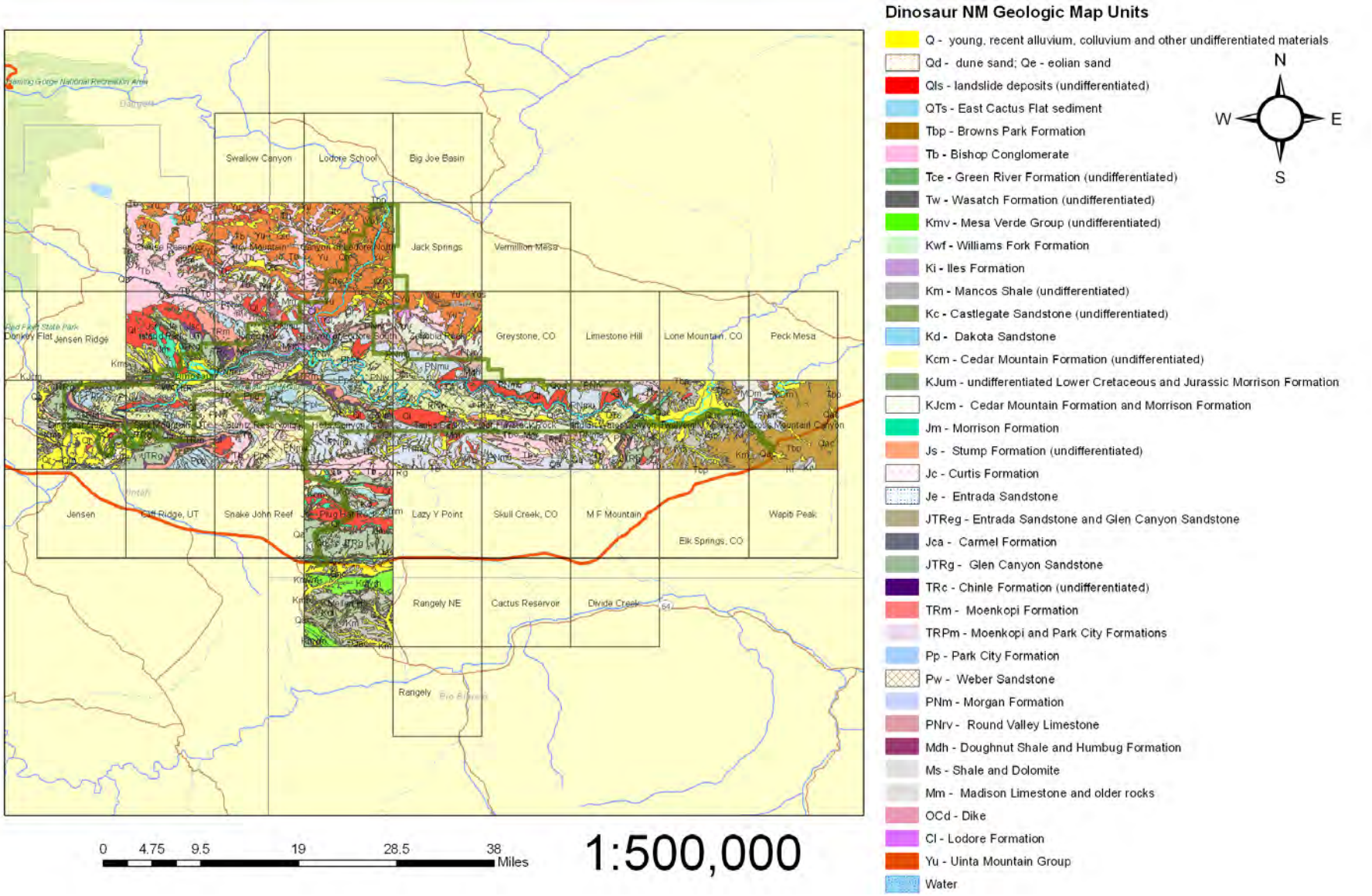
The majority of these units are fossiliferous within DINO, and the rest are potentially fossiliferous in the monument with the exception of the igneous dike. Fossils range from marine invertebrates (all of the Paleozoic units, the Carmel Formation, the Stump Formation, the Mowry Shale, the Frontier Formation and the Mancos Shale), to terrestrial trace fossils (the Chinle, Nugget, and Carmel formations), to abundant dinosaur bones (the Morrison and Cedar Mountain formations), to paleoecological materials from the past 10,000 years (the Quaternary surficial deposits) (DINO Table 1). The fossils of DINO are important parts of the research and exhibition collections of several institutions. DINO also has examples of fossils found in cultural contexts.



DINO Figure 2. Geography of DINO (NPS).



DINO Figure 3. Stratigraphic column for DINO (from Gregson et al. [2010], used with permission of the Utah Geological Association).



DINO Figure 4. Schematic geological map of DINO. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1038936.

Formation	Age	Fossils Within DINO	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Packrat middens	Alluvial, eolian, fluvial, and landslide deposits
Browns Park Formation	Miocene	None to date; root traces, arthropod traces, and mammal fossils are most likely	Eolian, fluvial, and lacustrine settings
Bishop Conglomerate	Early Oligocene	None to date; fossils are not currently known from this formation; as a conglomerate, reworked fossils from older units are possible	Alluvial fan complex
Mancos Shale	Late Cretaceous	Fossils have not been explicitly confirmed, but a variety of marine invertebrates (particularly bivalves and ammonites) are probably present	Open marine
Frontier Formation	Late Cretaceous	Coal, petrified wood, leaf impressions, bivalves, ammonites, gastropods, invertebrate burrows, and shark teeth	Nearshore marine to coastal terrestrial settings
Mowry Shale	Late Cretaceous	Angiosperm leaves, an ammonite, shark teeth, fish scales and bones, possible plesiosaurs, and possible crocodile relatives	Open marine
Dakota Sandstone	Early Cretaceous	None to date; petrified wood, twig casts, dinosaur bones, and bird tracks have been found in the immediate vicinity	Terrestrial (especially fluvial) becoming shallow marine over time
Cedar Mountain Formation	Early Cretaceous	Microbial mats, root casts, gastropods, ostracodes, invertebrate burrows, an undescribed dromaeosaurid dinosaur, and the sauropod dinosaur *Abydosaurus*	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	Microbial mounds, charophytes, spores, pollen, petrified wood, horsetails, ferns, cycads, ginkgoes, czekanowskiales, conifers, root casts, bivalves, gastropods, conchostracans, ostracodes, invertebrate trace fossils (including burrows and borings on wood and bone), ray-finned fish, lungfish, frogs, salamanders, turtles, sphenodontians, lizards, a possible snake, choristoderes, several taxa of crocodile relatives, over a dozen species of dinosaurs (often represented by partial skeletons or more), 16 or more species of early mammals, and dinosaur eggshell	Fluvial, floodplain, and lacustrine settings
Stump Formation	Late Jurassic	Fossils have not been explicitly confirmed, but a variety of marine invertebrates (particularly mollusks) are probably present	Nearshore marine (Curtis Member) becoming offshore (Redwater Member)
Entrada Sandstone	Middle Jurassic	None to date; fossils are very rare in this formation in general	Desert with large eolian sand dunes and occasional marine incursions
Carmel Formation	Middle Jurassic	Bivalves and reptile tracks	Marginal marine to coastal terrestrial settings

Formation	Age	Fossils Within DINO	Depositional Environment
Nugget Sandstone (formerly identified as the Glen Canyon Sandstone)	Late Triassic–Early Jurassic	Invertebrate burrows, fish scales, a partial phytosaur skeleton and other phytosaur fragments, and vertebrate tracks and coprolites (Bell Springs interval); invertebrate burrows, scorpion tracks, and tracks of dinosaurs, other reptiles, and mammal relatives (upper Nugget Sandstone)	Tidal flat or fluvial and lacustrine settings at the base (Bell Springs interval), transitioning to large eolian sand dunes for most of its history
Chinle Formation	Late Triassic	Petrified wood, root traces, crayfish burrows and other invertebrate trace fossils, and vertebrate tracks	Fluvial, floodplain, lacustrine, and eolian settings, becoming drier over time
Moenkopi Formation	Early–Middle Triassic	None to date; marine mollusks, invertebrate trace fossils, and tetrapod trace fossils have been found in the vicinity	Coastal plain with marine incursions
Park City Formation	Early–Middle Permian	Bryozoans, brachiopods, bivalves, ammonite and nautiloid cephalopods, and gastropods	Marine to supratidal settings deposited under multiple transgression-regression cycles of various scales
Weber Sandstone	Middle Pennsylvanian–Early Permian	Fusulinids, sand-sized grains that are fragments of bryozoans, crinoids, and possible algae, millimeter-scale burrows, and root traces or other invertebrate burrows	Coastal sand dunes
Morgan Formation	Middle Pennsylvanian	Fusulinid and other foraminifera, algae, sponges (including a *Chaetetes* reef), corals, bryozoans, brachiopods, ostracodes, crinoids, echinoids, tubular burrows, and conodonts	Shallow nearshore marine to beach settings
Round Valley Formation	Early Pennsylvanian	Foraminifera, solitary corals, bryozoans, brachiopods, and echinoids	Shallow marine
Doughnut Shale	Middle–Late Mississippian	Foraminifera, fossils of the plants *Calamites* and *Lepidodendron*, and ostracodes	Shallow marine and possibly coastal marsh
Humbug Formation	Middle Mississippian	Voids and molds left by the erosion of marine invertebrates, including corals and crinoids	Shallow marine
Madison Limestone	Early–Middle Mississippian	Solitary and colonial corals, bryozoans, brachiopods, gastropods, and crinoids	Shallow marine
Igneous dike	Late Cambrian or Early Ordovician	Unfossiliferous (intrusive igneous rocks)	Not applicable (igneous rocks that solidified at depth)
Lodore Formation	Late Cambrian	*Hyolithes*, an enigmatic ribbed conical invertebrate, brachiopods, trilobites, and invertebrate trace fossils	Marine near a coast
Uinta Mountain Group	Neoproterozoic	None to date; microfossils are most likely	Fluvial system with marine incursions

DINO Table 1. Concise tabulation of DINO stratigraphy, fossils, and depositional setting; details and references can be found in the text. Igneous rocks are included here for completeness, but are not discussed in detail in the text because they are unfossiliferous.

Sources of Paleontological Resources

The fossils of DINO present opportunities for education, interpretation, and continued or future scientific research in the monument.

Lodore Formation (Upper Cambrian)

The Lodore Formation is a sandstone and shale formation deposited near a shoreline during a marine transgression. It was named from an exposure in the Canyon of Lodore (Gregson et al. 2010). At DINO, it is a pink, tan, gray, and green unit 70 to 180 m (230 to 600 ft) thick, with three main divisions: a lower portion up to about 85 m (280 ft) thick composed of pebbly cliff-forming sandstone; a middle portion around 50 m (160 ft) thick composed of recessive shale and ledges of marine sandstone; and an upper portion up to 15 m (50 ft) thick composed of ledge-forming sandstone (Hansen et al. 1983; Gregson et al. 2010). The full thickness can be seen around Lodore and Whirlpool Canyons (Hansen et al. 1983). The contact with the overlying Madison Formation is a disconformity (Gregson et al. 2010). The Lodore Formation is typically dated to the Late Cambrian (Gregson et al. 2010), although it has been suggested that it is late Middle Cambrian to early Middle Cambrian in age (Robison 1964; Herr et al. 1982), or even as old as early Middle Cambrian (Robison 1964). The climate is interpreted as semi-arid to arid (Herr and Picard 1981).

The Lodore Formation is fossiliferous within DINO. The sparse fossil assemblage includes the enigmatic conical fossil *Hyolithes*, a ribbed conical form similar to the Devonian-age possible mollusk *Tentaculites* (these conical fossils may be responsible for early reports of gastropods; cf. Untermann and Untermann 1949a), a handful of brachiopod (lamp shell) taxa, trilobites, and invertebrate trace fossils (Untermann and Untermann 1949a, 1954; Herr et al. 1982). These fossils have been found near Jones Hole in the monument (Untermann and Untermann 1954). Body fossils are limited to the upper half of the formation (Herr et al. 1982), where they have been found in two silty shale layers (Untermann and Untermann 1949a).

Madison Limestone (Lower–Middle Mississippian)

The Madison Formation is a widespread, primarily Lower Mississippian marine unit. At DINO and in the vicinity, the Madison Limestone is composed of gray to tan cliff-forming limestone with chert nodules, zones of dolomite, and sparse fossils. It is 180 to 200 m (590 to 660 ft) thick (Hansen et al. 1983; Gregson et al. 2010). The lower part is sandy (Untermann and Untermann 1949a, 1954). Outcrops can be found in several areas of the monument, including a broad area near the Warm Springs anticline where the northern prong joins the main body of DINO, along the axis of Blue Mountain, and in several canyons, such as the Canyon of Lodore near Echo Park, Whirlpool Canyon, and Hells Canyon (Hansen et al. 1983). The upper contact with the Humbug Formation may be a disconformity (Hansen et al. 1983). The Madison Formation of the DINO area is interpreted as a shallow marine formation, and is dated by fossils to the first half of the Mississippian (Kinderhookian–Osagean and potentially slightly younger, roughly 359 to 340 Ma) (Katz et al. 2007). It was deposited during a transgression-regression cycle over a nearly flat surface by a warm and very shallow sea (Mayer 1964).

The strata assigned to the Madison Limestone at DINO were previously sometimes divided into the Madison and Deseret limestone formations. This division at DINO was based on correlations that are now regarded as incorrect because of the paleontology (Rowley and Hansen 1979; Rowley et al. 1979b). The two can be separated to the west where the Delle Phosphatic Member of the Deseret Limestone is present, but this unit cannot be found east of about Highway 191 (D. Sprinkel, pers. comm., January 2012). Eventually it may prove possible to distinguish both formations at DINO, but lithological evidence is lacking (Gregson et al. 2010). For reference, the Deseret Limestone is discussed in the Timpanogos Cave National Monument summary of this report.

The Madison Limestone is fossiliferous at DINO. In the Jones Hole area, the formation has yielded solitary and colonial corals, brachiopods, gastropods, and crinoids (sea lilies) (Madison and Deseret limestones of Untermann and Untermann [1954]). The bryozoan (moss animal) *Fenestella* was found in the middle part of the formation near Five Springs Draw (Mayer 1964). The Madison Limestone fossil assemblage of the DINO area includes solitary and colonial corals, bryozoans, brachiopods, gastropods, trilobites, and crinoids (Untermann and Untermann 1949a, 1954; Mayer 1964). Fossils tend to be found in specific layers rather than throughout (Untermann and Untermann 1949a, 1954). In an area including DINO in Moffat County, Colorado, Mayer (1964) reported two main fossiliferous zones, one 60 to 80 m (200 to 270 ft) up from the base, and another 30 m (100 ft) down from the top.

In Wyoming, various Madison Limestone units of similar age have yielded an assemblage including calcisphere microfossils (Chen and Webster 1994), foraminifera (amoeba-like protists that form "shells"), corals, bryozoans, brachiopods, bivalves, gastropods, ostracodes (seed shrimp), trilobites, crinoids, conodonts (extinct eel-like chordates) (Sando 1967; Chen and Webster 1994), fish (Sando 1967; Sandberg and Klapper 1967), and trace fossils including microbial laminations (Chen and Webster 1994), invertebrate traces (Sandberg and Klapper 1967; Chen and Webster 1994), and coprolites (fossil feces) (Sandberg and Klapper 1967).

Humbug Formation (Middle Mississippian)

The Humbug Formation is generally discussed with the overlying Doughnut Shale at DINO, and is usually mapped with it (Hansen et al. 1983; Gregson et al. 2010). It was not initially

differentiated from the Madison Limestone (Untermann and Untermann 1949a), but was in use at the monument by 1954 (Untermann and Untermann 1954). In the DINO area, it is composed of red, gray, and black beds of sandstone, shale, and limestone. The combined thickness of the Humbug and Doughnut formations is 75 to 90 m (250 to 290 ft). The two formations are found in the same general areas of DINO as the underlying Madison Limestone (Hansen et al. 1983; Gregson et al. 2010). These formations thin going from west to east (Rowley and Hansen 1979).

The Humbug Formation is a shallow marine unit (Gregson et al. 2010). Like the Madison Limestone, it was deposited during a marine advance, but unlike it there was some nearby source for clastics (sediments from eroded rocks) (Mayer 1964). To the west, in central Utah, the paleoenvironmental setting of the Humbug Formation is known in greater detail. There, deposition occurred under water depths thought to have been less than 30 m (100 ft) on a carbonate bank, and strong currents dominated, with quiet intervals. Clastic sediment arrived from the west. The water temperature was probably near 20 °C (68 °F), and salinity was probably close to normal marine values (Livingston 1955). In the Flaming Gorge area, about 50 km (30 miles) northwest of DINO at its nearest, it is thought to have been deposited in an intertidal setting (Sprinkel 2010; Sprinkel et al. 2010). Dating is imprecise due to sparse fossils, but it is thought to date to the Middle Mississippian (Meramecian) (Sandberg et al. 1979), between 340 and 333 Ma.

Fossils are uncommon in the Humbug Formation (Sandberg et al. 1979). It is probably fossiliferous at DINO, but this should be confirmed. In the Dinosaur Quarry area, Untermann and Untermann (1954) reported voids in the Humbug Formation they described as spaces left behind when fossils such as corals and crinoids weathered out, but found no identifiable fossils. The fossil assemblage of the formation over the rest of its depositional area includes foraminifera, possible microbial structures (Livingston 1955), rugose and tabulate corals (Gilluly 1932), bryozoans, brachiopods (Gilluly 1932; Calkins and Butler 1943; Livingston 1955), gastropods, trilobites (Calkins and Butler 1943; Livingston 1955), blastoids (sea buds), crinoids, echinoids (sea urchins) (Gilluly 1932), and possible invertebrate fecal pellets (Livingston 1955).

Doughnut Shale (Middle–Upper Mississippian)

As noted in the Humbug Formation section, the Doughnut Shale is usually mapped and described with the Humbug Formation at DINO. In the monument, it is mostly composed of dark gray clay-rich shale, becoming redder low in the formation (Hansen et al. 1983; Gregson et al. 2010). It also has some sandstone and limestone (Rowley and Hansen 1979). Historically, the Doughnut Shale is one of the troublesome units for DINO-area stratigraphic nomenclature. Its interval was recognized as distinct early on, but had no name (Untermann and Untermann 1949a, 1949b). It was briefly known informally as the "black shale" (Untermann and Untermann 1950, 1954; Sadlick 1954), and after that was correlated to the Manning Canyon Shale (Sadlick 1955). The present name began to be used during the 1970s (for example Sumison 1976) because of correlation issues with the Manning Canyon Shale, which was named from an area of western Utah on the opposite side of a major thrust fault (Rowley and Hansen 1979).

The Doughnut Shale of northeastern Utah was mostly deposited under shallow marine conditions near a coast. The presence of small amounts of coal indicate some nonmarine deposition as well (Gregson et al. 2010; Sprinkel 2010), possibly as brackish-water swamps (Mayer 1964). The marine waters, which had advanced during Humbug Formation time, were retreating to the west, and the climate was warm and humid (Mayer 1964). The formation dates to the Mississippian (Untermann and Untermann 1954). The most recent age assessment comes from Morgan et al. (2011), who date it to the Meramecian to late Chesterian stages (between 340 and 318 Ma) based on fossils.

The Doughnut Shale is fossiliferous within DINO. In the Island Park area, it has yielded foraminifera, fossils of the plants *Calamites* (a tree-like horsetail) and the "scale tree" *Lepidodendron*, and ostracodes (Untermann and Untermann 1954). Foraminifera and ostracodes have also been found in the Doughnut Shale on Whirlpool Canyon (Untermann and Untermann 1949b). Elsewhere, outside of DINO, it has yielded palynomorphs (organic microfossils such as pollen and spores) (Morgan et al. 2011), charophytes (a form of green alga) (Mayer 1964), coal, some additional types of plant fossils, corals, brachiopods, bivalves, ammonite and nautiloid cephalopods, and ostracodes (Sadlick 1954, 1955).

Round Valley Formation (Lower Pennsylvanian)

The Round Valley Limestone is an extensively fossiliferous marine unit. In the DINO area, it is 65 to 95 m (210 to 310 ft) thick and composed of gray to lavender limestone, with some red and gray shale, and nodules and layers of pink to red chert. Like the Paleozoic units below it, it is present in canyons, around Blue Mountain, and associated with the Warm Springs anticline (Hansen et al. 1983; Gregson et al. 2010). The Round Valley Limestone was first described as part of the overlying Morgan Formation. It is probably the same as the lower or Morrow member (in reference to its age) of Untermann and Untermann (1949b, 1954); they specifically note that Sadlick had suggested using the Round Valley name instead of keeping the rocks with the Morgan Formation (Untermann and Untermann 1954), and their description of the lithology is consistent with the modern usage of the Round Valley Formation. The Round Valley Formation is interpreted as having been deposited on a shallow marine carbonate shelf (Broadhead and Driese 1994), as part of another marine transgression (Gregson et al. 2010). It dates to the Early Pennsylvanian (Gregson et al. 2010); going by the old "Morrow member" terminology, it would date to the Morrowan stage, between approximately 318 and 312 Ma.

Fossils have been reported in the Round Valley Limestone from several locations in DINO. Untermann and Untermann (1949b) reported foraminifera, solitary corals, bryozoans, brachiopods, and echinoids in their "lower member of the Morgan Formation" in Whirlpool Canyon. Later, Untermann and Untermann (1954) reported that the "lower member" had foraminifera, solitary corals, bryozoans, and brachiopods in the Dinosaur Quarry area, and foraminifera, solitary corals, bryozoans, brachiopods, and echinoids in the Island Park area. Over its entire depositional area, the Round Valley Limestone fossil assemblage includes foraminifera, corals, bryozoans, brachiopods (Untermann and Untermann 1954; Sadlick 1955), bivalves, gastropods, ostracodes (Sadlick 1955), crinoids (Bilbey and Hamblin 1992), and echinoids (Untermann and Untermann 1954).

Morgan Formation (Middle Pennsylvanian)

The Morgan Formation is a heterogeneous and fossiliferous formation. At DINO, it includes an unnamed lower and upper member. The lower member ranges from 40 to 90 m (130 to 295 ft) in thickness, and includes beds of gray, red, and green shale and siltstone, gray sandstone, and gray to lavender limestone. The upper member ranges from 150 to 175 m (495 to 575 ft) in thickness, and is mostly composed of red sandstone and gray to lavender limestone. The two members are exposed in eastern, central, and western DINO (Hansen et al. 1983; Gregson et al. 2010). The lower unit thickens to the east, while the upper unit thins (Gregson et al. 2010).

The Morgan Formation was initially divided into three members (Untermann and Untermann 1949b, 1950, 1954). The lowest member is now at least in part the Round Valley Formation, as discussed above. It was briefly thought that Untermann and Untermann's middle member was a distinct formation, the Hells Canyon Formation of Thompson (1945), which was adopted by Sadlick (1955). Instead, the former middle member is now the lower member of the Morgan Formation (Rowley et al. 1979a; Rowley and Hansen 1979).

The Morgan Formation was deposited in beach and shallow nearshore marine settings (Gregson et al. 2010) as the middle part of the Round Valley Formation–Weber Sandstone marine transgression-regression cycle (Driese and Dorr 1984; Broadhead and Driese 1994). The marine influence decreased over time: the lower Morgan Formation includes shallow marine and shoreface rocks, while the upper part of the formation includes shoreface, subtidal, and eolian rocks (Broadhead and Driese 1994). Sandstones of the upper member were deposited as eolian dunes and in interdune areas, and were sometimes reworked by marine waters before lithification. Limestones represent marine deposition (Driese and Dorr 1984). Cycles of dune sandstone and shallow marine limestone can be recognized (Broadhead and Driese 1994). The formation dates to the Middle Pennsylvanian (Gregson et al. 2010). Under the Untermann and Untermann nomenclature, the middle (=lower) member dates to the Atokan stage, and the upper to the early part of the Desmoinesian (Untermann and Untermann 1949b), which would give an absolute age between 312 and 307 Ma.

The Morgan Formation is fossiliferous within DINO. Thompson (1945) reported algae, foraminifera, corals, brachiopods, and crinoids either within or just outside of DINO on the west side of Hells Canyon, in the "Hells Canyon Formation". Untermann and Untermann (1949b) found possible algae, fusulinids and other foraminifera, sponges, bryozoans, brachiopods, ostracodes, crinoids, and echinoids in the middle and upper Morgan Formation of Whirlpool Canyon. Untermann and Untermann (1954) reported foraminifera, sponges, bryozoans, brachiopods, and crinoids in the Dinosaur Quarry area; the same assemblage in the Island Park area, including a *Chaetetes* sponge reef in their middle member; and bryozoans, brachiopods, and crinoids in the Jones Hole area. Fryberger (1979) reported brachiopods, crinoids, and tubular burrows in the formation at Sand Canyon. Broadhead and Driese (1994) found conodonts at their Mantle Cave and Warm Springs localities in DINO.

Over its entire depositional area, the fossil assemblage of the Morgan Formation includes microbial mats, phylloid algae, fusulinids and other foraminifera, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, ostracodes, trilobites, crinoids, echinoids, invertebrate burrows, conodonts, fish, and general bioturbation (disruption of sedimentary features by biological activity) (Driese and Dorr 1984). Most fossils are in the limestone beds; sandstone fossils are mostly invertebrate burrows and bioturbation (Driese and Dorr 1984).

Weber Sandstone (Middle Pennsylvanian–Lower Permian)

The Weber Sandstone is a thick, predominantly sandstone unit with sparse fossils. In the DINO area, the Weber Sandstone is composed of light-colored cross-bedded sandstone, between 230 and 330 m (755 and 1,080 ft) thick (Hansen et al. 1983; Gregson et al. 2010). There are also some thin limestone beds (Gregson et al. 2010), typically found near the base or top of the formation (Bissell and Childs 1958). The upper contact with the Park City Formation is sometimes described as disconformable (Fryberger 1979). This formation was first known as the Yampa Sandstone (Untermann and Untermann 1954). It is well-known as the reservoir rock for the Rangely oil field, and also forms the cliffs of Cliff Ridge and Split Mountain at DINO (Gregson et al. 2010).

The Weber Sandstone is mostly eolian in origin, from an area of sand dunes near a beach (Gregson et al. 2010), and represents the end of the Round Valley–Weber marine cycle (Driese and Dorr 1984; Broadhead and Driese 1994). The sand dunes seem to have been moving to the south over time (Driese 1985). Some of the limestone represents shallow interdune ponds, with brackish to hypersaline water (Driese 1985). The marine influence may have been most pronounced during the early and late stages of deposition (Fryberger 1979). Cyclic deposition has

been recognized in some places, similar to the deposition in contemporaneous basins such as the Paradox Basin to the south (Johnson et al. 1988). Foraminifera found at Split Mountain indicate a Middle Pennsylvanian–Early Permian age (Desmoinesian–Wolfcampian stages) for the formation (Bissell and Childs 1958; Bissell 1964), between approximately 308 and 280 Ma.

The Weber Sandstone is sparsely fossiliferous within DINO. Fusulinids have been found at Split Mountain, as well as sand-sized grains that are fragments of bryozoans, crinoids, and possible algae (Bissell 1964). Millimeter-scale burrows are known from Hog Canyon, becoming more common in the upper quarter of the formation. Similar burrows are found at Deer Lodge Park and Sand Canyon (Fryberger 1979). There are also root traces or invertebrate burrows near Echo Park (Driese 1985). The fossil assemblage of the Weber Sandstone over its entire area includes probable stromatolites (layered sedimentary structures created by algae and microbes) or microbial mats (Driese 1985), algae, fusulinids, stromatoporoids (coral-like sponges) (Bissell 1964), corals (Gregson et al. 2010), bryozoans (Bissell 1964), brachiopods (Gregson et al. 2010), crinoids, fecal pellets (Bissell 1964), and invertebrate burrows (Fryberger 1979). Fossils are locally found in the marine limestone beds (Gregson et al. 2010). Transported and detrital fossils of marine organisms are sometimes found in the eolian beds; these fossils are mostly fusulinids or sand-size fragments of larger fossils (Bissell 1964).

Park City Formation (Lower–Middle Permian)

The Park City Formation is a heterogeneous marine unit that has a complex relationship with the Phosphoria Formation across its depositional area (Peterson 1984). Within DINO, it is divided into two members: the lower member correlates to the Meade Peak Phosphatic Shale of the Phosphoria Formation and the lower Franson Member of the Park City Formation; and the upper member correlates to the rest of the Franson Member (Hansen et al. 1983; Gregson et al. 2010). This is similar to but less complex than the division seen in the Flaming Gorge region, where the Meade Peak Phosphatic Shale is explicitly distinguished, and there is an additional unit (the Grandeur Member of the Park City Formation) at the base of the interval (Sprinkel 2010; Sprinkel et al. 2010). Occasionally the Meade Peak Phosphatic Shale Member is explicitly distinguished at DINO (Schell and Yochelson 1966; Sumison 1976).

In the DINO area, the lower member is 10 to 14 m (33 to 46 ft) thick, and consists of gray-brown sandstone, dolomite, and limestone beds that are locally phosphatic and fossiliferous. The upper member is 10 to 21 m (33 to 69 ft) thick and includes grayish beds of limestone, siltstone, sandstone, and dolomite. The lower member forms cliffs, while the upper member forms slopes with the notable exception of a ledge-forming phosphatic dolomite or limestone bed (Hansen et al. 1983). The lower unit caps areas of the Weber Sandstone (Gregson et al. 2010). The contact with the overlying Moenkopi Formation at DINO has been described as a disconformity (Gregson et al. 2010), conformable (Untermann and Untermann 1954; Schell and Yochelson 1966), or even transitional (Untermann and Untermann 1954). Not surprisingly, the contact is difficult to identify (Schell and Yochelson 1966; Rowley et al. 1979b).

The Park City and Phosphoria formations were deposited as the result of a series of marine transgression-regression cycles of various scales. The Meade Peak Phosphatic Shale and Franson members share one large cycle, and there appears to be at least three smaller cycles with depositional hiatuses within the Franson Member, as well as two transgressions and one regression in the Meade Peak Phosphatic Shale Member (Peterson 1984). These cycles were identified from Wyoming rocks (Peterson 1984), so may not all be present in the DINO area (Schell and Yochelson 1966). The Meade Peak Phosphatic Shale Member appears to be related to cold water upwelling, while the Franson Member is regressive overall, becoming supratidal in places (Hendrix and Byers 2000). The climate during deposition of the Franson Member was probably hot and dry, and in some areas the seawater had high salinity (Peterson 1984). The Meade Peak Phosphatic Shale and Franson members date to the Lower and Middle Permian (Leonardian–Guadalupian) (Rigby et al. 2007), between 280 and 260 Ma.

The Park City Formation is fossiliferous at DINO. Untermann and Untermann (1954) found gastropods in this formation in the Dinosaur Quarry area, often replaced by chert. Schell and Yochelson (1966) described fossils from the upper part of the formation at Split Mountain, including bryozoans, brachiopods, bivalves, ammonite and nautiloid cephalopods, and gastropods. Some of these shells have been bored. Some of the fossils are silicified, which is a common phenomenon for Park City Formation fossils (Schmitt and Boyd 1981). Fossils that may be found in the DINO area include brachiopods, bivalves, cephalopods, gastropods, and scaphopods (tusk shells) (Rowley et al. 1979b; Gregson et al. 2010). They are more common in the western part of the area (Gregson et al. 2010).

Over its entire depositional area, the Meade Peak Phosphatic Shale Member of the Phosphoria Formation has yielded acritarchs (organic microfossils that cannot be classified as anything else) (Jacobson et al. 1982), sponges (Hendrix and Byers 2000), conulatans (extinct coral relatives that formed conical structures from rods), bryozoans, brachiopods, bivalves, ammonites, belemnites, other cephalopods, gastropods, ostracodes, crinoids (Yochelson 1968), invertebrate burrows and general bioturbation (Hendrix and Byers 2000), conodonts (Behnken et al. 1986), and fish (Yochelson 1968). Similarly, the Franson Member of the Park City Formation has yielded acritarchs (Jacobson et al. 1982), phylloid algae (Peterson 1984), sponges, solitary corals, bryozoans, brachiopods, bivalves, nautiloids, gastropods, scaphopods, ostracodes, crinoids, possible echinoids (Yochelson 1968), burrows (Hendrix and Byers 2000), conodonts (Jacobson et al. 1982), and fish (Yochelson 1968).

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location (Stewart et al. 1972a). In the DINO area, it is typically divided into five formal and informal members. From oldest to youngest, they are the Gartra Member, the mottled (or purple) member, the ocher (or ochre) siltstone member, the sandstone and conglomerate member, and the red siltstone or upper red member (Poole and Stewart 1964a, 1964b; Stewart et al. 1972a; Dubiel 1992; Jensen and Kowallis 2005; Erickson 2007). A sixth member, the upper member of Poole and Stewart (1964a, 1964b) and Stewart et al. (1972a), is now the Bell Springs Member of the Nugget Sandstone/formation of Bell Springs (Jensen and Kowallis 2005; Sprinkel et al. 2011a). The Chinle Formation is up to 110 m (360 ft) thick, and is found in areas of central and extreme eastern and western DINO. The upper contact with the Nugget Sandstone appears to be conformable (Rowley and Hansen 1979; Rowley et al. 1979a, 1979b; Jensen 2005; Sprinkel et al. 2011a); there is no obvious disconformity, at least (Untermann and Untermann 1954). The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) (Lucas et al. 1997, 2006a).

As with several of the other formations, the Chinle Formation of DINO has had a complicated nomenclatural history. Part of the issue is the fact that the Uinta Basin is the northernmost extreme of the Chinle Formation, so formation names from adjacent areas have also been applied (particularly the Popo Agie Formation; see for example McCormick and Picard [1969a, 1969b] or High et al. [1969]). Further complicating matters, intervals have been placed in or added from other formations. McCormick and Picard (1969a) split off the Gartra Member as its own formation and moved the red siltstone and upper members into the Nugget Sandstone. Lucas (1993) raised the Chinle Formation to a group, made the Gartra Member a formation, designated most of the rest of the previous Chinle Formation of the area as the Popo Agie Formation, and raised the Bell Springs Member of the overlying Nugget Sandstone to the Bell Springs Formation of the Chinle Group. Hunt et al. (1993), following Lucas (1993)'s work, used the red siltstone and upper members at DINO as the lower and upper members of the Bell Springs Formation. Other workers have, until recently, generally kept the Bell Springs interval in the Nugget Sandstone. Its rocks are now being mapped as a separate formation, including the former upper member of the Chinle Formation (Jensen and Kowallis 2005; Sprinkel et al. 2011a) but not including the red siltstone member (which is retained as part of the upper Chinle Formation) (Haddox et al. 2010a, 2010b).

The Gartra Member at DINO was described as the Shinarump Formation in early publications (Untermann and Untermann 1949a, 1950, 1954). It is a discontinuous unit mostly made up of light-colored sandstone, conglomeratic sandstone, and conglomerate, and fills channels eroded in the underlying rocks (Poole and Stewart 1964a, 1964b). Grain size becomes finer going up-section, with some siltstone and claystone in the upper part (McCormick and Picard 1969a). Deposition was fluvial, with west-flowing streams (Poole and Stewart 1964a, 1964b). The setting was a broad alluvial plain (McCormick and Picard 1969a), and over time, as the channels were filled in, the fluvial systems expanded laterally (Erickson 2007). The overlying mottled member continued the fluvial-floodplain deposition (Erickson 2007). It is composed of purple and red mudstone, and mottled sandstone and siltstone (Poole and Stewart 1964a, 1964b). It could represent in part a fossil soil zone (Stewart et al. 1972a). Erickson (2007) found a hiatus between the mottled member and younger rocks of the Chinle Formation, with the locally-overlying sandstone and conglomerate member as a fluvial cut and fill unit.

The ocher siltstone member of the DINO area is composed of siltstone, clay-rich siltstone, and silty claystone, thinning to the south and east. It is interpreted as representing deposition in a shallow water setting, such as a lake (Poole and Stewart 1964a, 1964b). Some of its sediments were derived from volcanoes (Stewart et al. 1972a). In some areas it is succeeded by the upper member, but in others the red siltstone member or the sandstone and conglomerate member is present between them. This is more common in the east, because the red siltstone member forms an east-thickening wedge between the ocher member and the Bell Springs interval ("upper member") (Poole and Stewart 1964a, 1964b; Stewart et al. 1972a). The sandstone and conglomerate member is only found locally in the easternmost part of the Uinta range, and is thought to be a basal coarse facies of the overlying red siltstone member. It includes sandstone, conglomerate, siltstone, and mudstone, generally becoming finer upward. Deposition was on a floodplain, with streams flowing to the southwest. The red siltstone member is mostly composed of its namesake rock, with some minor sandstone and conglomerate. It interfingers with the Bell Springs interval, and is interpreted as having been deposited in floodplain, delta, and lacustrine settings (Poole and Stewart 1964a, 1964b).

The Chinle Formation is fossiliferous at DINO, with fossils including petrified wood, invertebrate trace fossils, and tetrapod bone fragments and tracks. Invertebrate burrows have been found throughout the formation at several locations in DINO (Dubiel 1992). Erickson (2007) found petrified wood, root traces, crayfish and other invertebrate burrows, and general bioturbation at several locations, mostly in the Gartra, and mottled members. The Gartra Member had the most diverse fossils, with examples of all of the above, while the mottled and ocher members produced invertebrate burrows. Petrified wood is common in the Gartra Member at DINO (Untermann and Untermann 1949a), with specimens up to log size (Untermann and Untermann 1954).

The best-known Chinle Formation fossils at DINO are tetrapod tracks. 14 tracksites representing 17 stratigraphic levels are known from the red siltstone member of the Chinle Formation and overlying Bell Springs interval of the Nugget Sandstone, concentrated in the Split Mountain area. Seven types of vertebrate traces are known: *Rhynchosauroides* (made by lizards or sphenodontians [relatives of the modern tuatara]), ?*Apatopus* (made by trilophosaurian reptiles), *Gwyneddichnium* (made by tanystropheid or drepanosaurid reptiles), ?*Chirotherium* (made by phytosaurs, crocodile-like reptiles), *Brachychirotherium* (made by aetosaurs, armored herbivorous reptiles), *Agialopus* (made by theropod dinosaurs), and unnamed synapsid tracks. Additionally, invertebrate traces including *Kouphichnium* (horseshoe crab trails), *Scoyenia*, and *Isopodichnus* were found at some of these sites (Lockley et al. 1992). Similar trace localities are known from north of Vernal and the Vermillion Creek area 21 km (13 miles) north of DINO; the Vermillion Creek area has more diverse invertebrate traces than known DINO sites. Lockley (2006) described the *Gwyneddichnium* tracks in greater detail; they are of interest because they show the animals both walking and swimming, and show the webbing of the feet. Hunt et al. (1994) referred to dinosaur tracks at DINO in the Bell Springs Formation, including the theropod track types *Agialopus* and *Grallator*, and the prosauropod track types *Pseudotetrasauropus* and *Tetrasauropus* (=*Eosauropus*; Lucas et al. 2006a). Based on their and Lucas et al. (2006a)'s descriptions, these refer to tracks from the lower Nugget Sandstone described by Lockley et al. (1992), not rocks from the Bell Springs interval or Chinle Formation rocks included by Lucas (1993) or Hunt et al. (1993, 1994) in the Bell Springs Formation.

Similar fossil assemblages are known for the members in the vicinity outside of DINO. The Gartra Member has petrified wood, root traces, crayfish and other invertebrate burrows, and bioturbation (Erickson 2007). Petrified wood is locally abundant (McCormick and Picard 1969a). The mottled member has root traces, crayfish burrows, bioturbation (Erickson 2007), and, in the basal purple marl, there are amphibian bone fragments, phytosaur teeth, indeterminate bone fragments, and coprolites. Phytosaur teeth and coprolites were found in these beds in the Red Mountain area, 19 km (12 miles) northwest of DINO (Untermann and Untermann 1954), and bones of trilophosaurs and phytosaurs along with abundant coprolites were found in a coarse bed of the member north of Vernal (High et al. 1969; as the purple unit of the Popo Agie Formation). The ocher siltstone member has crayfish burrows, other invertebrate traces, and bioturbation (Erickson 2007). Phytosaur teeth and bone fragments are common in the sandstone and conglomerate member, especially in the coarser beds (Poole and Stewart 1964a). Bone fragments have been found in this unit in the Cliff Creek area just south of the western end of DINO, and phytosaur bones were collected just beyond DINO's eastern boundary in the Cross Mountain area (Stewart et al. 1972a). The red siltstone member has yielded petrified wood (Untermann and Untermann 1954), burrows, bioturbation (Erickson 2007), sparse bone fragments (Poole and Stewart 1964a, 1964b), and vertebrate tracks (Lockley et al. 1992). The Red Fleet State Park area, about 10 km (6 miles) northwest of DINO, has petrified wood in conglomeratic fluvial channels of the Gartra Member, and tetrapod tracks assigned to *Brachychirotherium* and *Agialopus* (Hamblin et al. 2000).

The various members of the Chinle Formation have similar types of fossils, but differ in the genera and species present and which are most commonly found. Plants are represented by charophyte algae (Lucas and Keitzke 1993), diverse palynomorphs (Gottesfeld 1972; Fisher and Dunay 1984; Parker 2004), and leaves, wood, and other remains of lycopods, horsetails, ferns and fern-like plants, cycads, bennettitales, ginkgoes, conifers, and taxa of uncertain classifications (Ash 1972, 1989, 2003, 2005a; Stewart et al. 1972a; Parker and Ash 2004). Fungal remains are also present (Ash 1972; Stewart et al. 1972a). Invertebrates are represented by tentaculites (enigmatic possible mollusks, misidentified as worm fossils [Ash 2005b]), bivalves, gastropods, ostracodes, conchostracans (clam shrimp) (Breed 1972; Stewart et al. 1972a), crabs (Rinehart et al. 2003), crayfish (Miller and Ash 1988), insects (Breed 1972; Stewart et al. 1972a), and various trace fossils (Caster 1944; Breed 1972; Gillette et al. 2003; Demko et al. 2005), including borings in wood (Walker 1938; Tapanila and Roberts 2012) and a variety of feeding traces on plants (Ash 1996). Reworked Paleozoic fossils, primarily of invertebrates, are occasionally found in the lowest part of the formation, and include foraminifera, algae, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, ostracodes, and crinoids (McKee 1937; Stewart et al. 1972a).

The Chinle Formation's vertebrate assemblage has been summarized in a number of documents, including Stewart et al. (1972a), Colbert (1972), Parrish (1989), Long and Murry (1995), Heckert et al. (2005a, 2005b), Irmis (2005), Parker (2005, 2009), Lucas et al. (2007), Loughney et al. (2011), and Parker and Martz (2011). Vertebrates include: freshwater sharks; ray-finned fish; lungfish; coelacanths; several types of enigmatic tetrapods known from isolated bones and teeth; metoposaur amphibians; procolophonids (stocky lizard-like herbivorous and insectivorous reptiles); sphenodontians; trilophosaurids (reptiles that resembled large herbivorous lizards); tanystropheids (long-necked piscivorous reptiles); the unusual aquatic reptile *Vancleavea*; drepanosaurids (beaked, possibly aquatic chameleon-like reptiles); several forms of crocodile-like phytosaurs, known under a variety of names; several forms of herbivorous armored aetosaurs, also known under a variety of names; the small unusual armored *Revueltosaurus*; "rauisuchian" crocodile relatives including poposaurids and rauisuchids, which resembled large carnivorous dinosaurs, and shuvosaurids (toothless forms that resembled ostrich-mimic dinosaurs); "sphenosuchians" (small terrestrial crocodile relatives); lagerpetonids (small bipedal dinosaur relatives); silesaurids (quadrupedal dinosaur relatives); several early dinosaur taxa; and dicynodonts (a type of non-mammalian therapsid, or "mammal-like reptile"). Lungfish, metoposaurs,

aetosaurs, and phytosaurs are the most common forms. Vertebrate trace fossils include coprolites (Hunt et al. 1998), swim traces, and footprints possibly attributable to sphenodontians, tanystropheids, phytosaurs, aetosaurs, carnivorous and herbivorous dinosaurs, and mammal relatives (Hunt and Lucas 2007). Bones accumulated in floodplains, bogs, small ponds, and in or near channels (Parrish 1989).

Nugget Sandstone (Upper Triassic–Lower Jurassic)

The Nugget Sandstone is a prominent resistant cross-bedded sandstone unit. It has also gone under the names Glen Canyon Sandstone, Navajo Sandstone, and Nugget–Navajo Sandstone, depending on the preference of the author. It appears to correlate to the entire vertical Glen Canyon Group sequence (Wingate to Kayenta to Navajo formations) (Poole and Stewart 1964a, 1964b; Dubiel 1992; Gregson et al. 2010; Sprinkel et al. 2011a). In the DINO area, it is between 180 and 200 m (590 and 660 ft) thick, and is mostly composed of gray to pink cross-bedded sandstone. Cross-bedding diminishes near the base, replaced by parallel bedding (Hansen et al. 1983; Gregson et al. 2010). This basal interval is 26 to 36 m (85 to 120 ft) thick and dominated by parallel bedding. The lower half of this interval, approximately 15 m (50 ft) thick, is made up of interbedded sandstone, siltstone, and mudstone, and the upper half is predominantly composed of siltstone and mudstone with two sandstone beds (Sprinkel et al. 2011a). This interval is sometimes identified as the Bell Springs Member of the Nugget Sandstone (Pipiringos 1968; Jensen and Kowallis 2005), but is increasingly distinguished as an informal separate formation (Haddox et al. 2010a, 2010b Sprinkel et al. 2011a), as mentioned above in the Chinle Formation section. Regardless of whatever nomenclature is eventually adopted, it appears to be transitional between the fluvial underlying Chinle Formation and the overlying eolian Nugget strata, potentially correlative to the Church Rock Member of the Chinle Formation (Sprinkel et al. 2011a), and might be more closely associated with the Chinle Formation (D. Sprinkel, pers. comm., January 2012), which would hearken back to its previous designation as the upper member of the Chinle Formation.

The great majority of the Nugget Sandstone is eolian, with winds from the north and northeast, but the basal Bell Springs interval was deposited by water (Poole and Stewart 1964a, 1964b). The Bell Springs interval is interpreted as a tidal flat unit (Jensen 2005; Jensen and Kowallis 2005) or predominantly fluvial and lacustrine (Haddox et al. 2010a, 2010b). Lithological cycles have been recognized in the Nugget Sandstone northwest of DINO, representing long-term fluctuations in precipitation; over time, the cycles became truncated as aridity increased (High and Picard 1975). In DINO, the Nugget Sandstone can be found in the east, west, and near the Red Rock anticline (Hansen et al. 1983). It is dated to the Late Triassic and Early Jurassic (Gregson et al. 2010), with the Bell Springs interval recognized as Late Triassic based on its fossils (Lockley et al. 1992), and the Late Triassic–Early Jurassic boundary low in the overlying eolian interval of the formation (Sprinkel et al. 2011a). The upper contact with the Carmel Formation is disconformable (Gregson et al. 2010).

The Nugget Sandstone is fossiliferous at DINO. The Bell Springs interval of the monument has yielded invertebrate trace fossils (Erickson 2007), fish scales (Hunt et al. 1993), phytosaur bones and teeth (Stewart et al. 1972a; Hunt et al. 1993), tetrapod tracks (Lockley et al. 1992), and vertebrate coprolites (Hunt et al. 1993). Two phytosaur teeth were collected from the upper part of the interval on the southwest side of Split Mountain within DINO (Stewart et al. 1972a). Fish scales are locally abundant, representing at least three taxa including a coelacanth (Hunt et al. 1993). Hunt et al. (1993) reported three phytosaur bone sites in the Bell Springs interval at DINO. These sites have yielded a partial fragmentary skeleton, a scute, and a probable tooth. They also noted phytosaur teeth at Bourdette Draw in the monument. Additionally, vertebrate coprolites are common in a 2 m (7 ft) thick interval near the top of the interval; they often include fish scales (Hunt et al. 1993). As noted above, several track sites have been found in the Bell Springs interval of the monument, with similar track types to those found in the underlying red siltstone member of the Chinle Formation (Lockley et al. 1992; the Bell Springs interval is the "purple member of the Popo Agie Formation" of their usage). Lockley et al. (1992) also described two track sites near the base of the upper Nugget Sandstone (about 7 m [20 ft] above the Bell Springs interval) in western DINO, on the southwest flank of Red Mountain. One site had *Brachychirotherium* tracks and a large circular to ovoid track, and the other had a three-toed track among general trampling. Similar tracks are known from the Vernal area. These tracks are thought to date from the Late Triassic. Hunt et al. (1994) and Lucas et al. (2006a) identify the ovoid track type as *Eosauropus* or its synonym *Tetrasauropus*, and the three-toed track as *Grallator*. Six other track horizons are known from the formation within DINO, mostly within 15 m (50 ft) of the upper contact. The upper assemblage is dominated by the prosauropod track type *Otozoum*, with small and medium theropod tracks also present (Gregson et al. 2010). Engelmann et al. (2010) reported a site in DINO with abundant *Brasilichnium* tracks (probable therapsid tracks) on multiple horizons. Over 300 such tracks were found on one surface. Also present at the site were burrows possibly from colonial arthropods, scorpion trackways, and tracks of a larger, unidentified tetrapod (Engelmann et al. 2010). More tracks were found at a comparable horizon near the Dinosaur Quarry (Anderson et al. 2011). Fossils are restricted to interdune rocks of the eolian upper Nugget Sandstone at DINO and in the vicinity; interdunes are best developed in the lower and upper thirds of the formation (Chure et al. 2009).

Similar fossils are known from the Nugget Sandstone of the vicinity of DINO. The Bell Springs interval has yielded beetle and clam burrows, bioturbation (Erickson 2007), and a four-toed track found in the Red Mountain area (Untermann and Untermann 1954). The upper Nugget Sandstone has a fossil

assemblage including possible microbial structures in interdune carbonates (Chure et al. 2009), plants (Britt et al. 2010; Chambers et al. 2011), molds of freshwater gastropods (Engelmann et al. 2009), invertebrate trace fossils (Faul and Roberts 1951; Engelmann et al. 2010), bones of sphenodontians and theropod dinosaurs, including a bonebed (Britt et al. 2010; Chambers et al. 2011), tetrapod tracks (Lockley et al. 1992; Hamblin et al. 2000), and possible synapsid (mammals and their closest relatives) burrows from the vicinity of DINO (Engelmann and Chure 2008). The Red Fleet State Park area has small and large three-toed dinosaur tracks and splayed quadrupedal tracks, possibly *Brasilichnium* (Hamblin et al. 2000). 350 examples of *Grallator* (small theropod tracks) and *Eubrontes* (large theropod tracks) were found in nine layers (Hamblin and Bilbey 1999). Another trace fossil location is south of Meeker, Colorado, 68 km (42 miles) southeast of DINO. Worm and gastropod trails, scorpion or spider tracks, and bipedal and quadrupedal reptile tracks were found here (Faul and Roberts 1951). The first body fossils in the formation were publicized in 2009 (Chure et al. 2009). Three stacked bonebeds were discovered in northeastern Utah, with most of the fossils representing a new coelophysoid theropod dinosaur. Over 2,700 bones and fragments belonging to at least 12 individuals have been found. A second distinct coelophysoid is represented by a partial skeleton, and a third larger type of theropod is represented by a tooth. Probable cycad foliage and sphenodontian bones were also found. The new finds are from the middle of the upper eolian unit. The organisms lived during a wet phase, possibly near an ephemeral lake (Britt et al. 2010; Chambers et al. 2011).

San Rafael Group: Carmel Formation (Middle Jurassic)

The Carmel Formation is a marine and terrestrial Middle Jurassic unit. The transition between these general depositional settings is near DINO. In the southwest Uinta Mountains, the laterally-equivalent fully marine Twin Creek Formation is present. This is replaced by marine, marginal marine, and terrestrial transitional rocks of the Carmel Formation by Vernal, which are in turn replaced by predominantly terrestrial Carmel Formation rocks by the Dinosaur Quarry area (Rigby 1964). The Carmel Formation thins to the east and pinches out a short distance into Colorado (Rigby 1964; Hansen et al. 1983). Within DINO, it is a nonresistant, typically red formation composed of shales, sandstone, siltstone, mudstone, and minor gypsum. It has a maximum thickness of about 40 m (130 ft), and is primarily found around Split Mountain in the monument; it does not extend to the eastern end of the monument (Hansen et al. 1983; Gregson et al. 2010). The Carmel Formation of DINO only represents the uppermost fraction of the section present to the west, and probably correlates to the Dewey Bridge Member of the Carmel Formation as seen at Arches National Park (ARCH) over 175 km (110 miles) to the south (Rigby 1964). The formation dates to the Middle Jurassic, with an age of approximately 169 to 162 Ma based on radiometric dates from central and southwestern Utah (Sprinkel et al. 2011b).

Both the transitional and terrestrial facies of the Carmel Formation of the DINO area display cyclic lithologies. In the transitional facies, the cycles include greenish-gray shales and limestones, red shales, and evaporites. Greenish-gray shales and limestones represent shallow marine deposition, red shales represent very shallow marine to mudflat deposition, and gypsum represents evaporitic conditions between transgression-regression cycles. The terrestrial facies features cyclic alternations of reddish sandstone, siltstones, and shales, and is interpreted as having been deposited in fluvial and coastal margin settings (Rigby 1964).

The Carmel Formation is reported as fossiliferous at DINO, but the occurrences should be documented formally. Loope and Simpson (1992) mentioned the presence of reptile tracks in the Carmel Formation just below the Entrada Sandstone at DINO, based on a personal communication. Douglas Sprinkel (pers. comm., January 2012) reported the presence of bivalves in marine beds and tracks in fluvial beds in the monument. Untermann and Untermann (1954) reported three-toed and four-toed tracks in the Carmel Formation along the Chew road near southwestern DINO; Sprinkel (pers. comm., January 2012) has seen these tracks, which are near the petroglyphs on the road, probably on BLM land.

Considering the whole of the Carmel Formation, most fossils are of marine origin. Marine fossils include dinoflagellates (a type of algae that moves using two flagella) (Sprinkel et al. 2011b), algae, foraminifera, radiolarians (another type of "shelled" protist), sponges, corals, bryozoans, brachiopods, bivalves, cephalopods, gastropods, ostracodes, asteroids (sea stars), crinoids, echinoids, and trace fossils including pellets and worm tubes (Imlay 1964; Blakey et al. 1983; Nielson 1990; Wilson 1998). Marine bivalves are abundant north of Vernal (Untermann and Untermann 1949a), and bivalves and gastropods have been found at Steinaker Draw on the northeast side of the city (Untermann and Untermann 1954). Echinoderms have also been found in the DINO area as well (Rowley et al. 1979a). Terrestrial fossils include spores, pollen (Sprinkel et al. 2011b), and tracks of crocodylomorphs and dinosaurs (Lockley and Hunt 1995). At least 60 dinosaur trackways have been found in the Jensen–Vernal area, including tracks named *Carmelopodus untermannorum* for the Untermanns (Lockley et al. 1998). Another nearby area with dinosaur tracks is Red Fleet State Park, where two sandstone beds among the mudflat and shoreline rocks have three-toed tracks including *Carmelopodus* (Hamblin et al. 2000). In the Uinta Mountains region, fossils are rare in the terrestrial facies, but marine fossils can be found in limestones in the transitional facies (Rigby 1964).

Stump Formation (Upper Jurassic)

The Stump Formation of DINO includes two distinct marine members that are otherwise known under different classifications over large areas: the lower Curtis Member, known as a formation in much of eastern and southern Utah into western Colorado; and the overlying Redwater Member, also

known as the Redwater Shale Member of the Sundance Formation in much of Wyoming and the Black Hills of South Dakota. In the DINO area, the Stump Formation and Redwater Member were not recognized until the mapping of the 1970s, and older references identify the entire interval as the Curtis Formation (Untermann and Untermann 1949a, 1954). Untermann and Untermann (1949a, 1954) divided the Curtis Formation into a lower third composed of greenish-gray to brown sandstone and an upper two-thirds composed of sandstone, limestone, and shale beds, which can be recognized as the modern Curtis and Redwater members of the Stump Formation. The Curtis Member is interpreted as a nearshore unit and the Redwater Member as a transgressive marine unit, with an overlying regressive member (the Windy Hill Member, also sometimes assigned to the Morrison Formation) in some areas such as Red Fleet State Park (Hamblin et al. 2000). The marine body transgressed from the west (Brenner and Davies 1974), and sediment arrived to the DINO area from the east (Patterson-Wittstrom 1980). The Stump Formation dates to the early Late Jurassic (Oxfordian stage) (Kvale et al. 2001; Wilcox 2007; Hintze and Kowallis 2009), approximately 161 to 156 Ma.

Stump Formation: Curtis Member (Upper Jurassic)
The Curtis Member of DINO is typically a ledge-forming gray, cross-bedded sandstone unit between 7 and 30 m (23 and 100 ft) thick, thinning west to east. It and the overlying Redwater Member are limited to small areas of western and extreme eastern DINO (Hansen et al. 1983; Gregson et al. 2010). The upper contact with the Redwater Shale is disconformable, and the two have an irregular relationship, with the proportion of one to the other ranging widely from location to location. They represent different marine transgressions as well (Pipiringos and Imlay 1979). The sea that deposited the Curtis Member was on the order of 30 m (100 ft) deep at its greatest in Utah, and retreated during deposition (Hoggan 1970). When it first advanced, the sea caused the collapse of the massive dunes of the Entrada Sandstone erg (estimated as up to 100 m [300 ft] tall), leading to the formation of a disrupted basal sandstone layer for part of the Curtis Member (Eschner and Kocurek 1986).

The Curtis Member is probably fossiliferous within DINO. Untermann and Untermann (1954) reported the presence of guards from the belemnite *Pachyteuthis* and casts of gastropods in the upper part of the Curtis Member in the Dinosaur Quarry area, and found bivalves and gastropods just outside of the monument to the east. Invertebrate trace fossils are also known from the area (Gregson et al. 2010). Specimens of *Pachyteuthis* have been found at several archeological sites within or just outside of DINO, and are probably locally derived from this member or the Redwater Member (Breternitz 1970).

As a transitional unit, both marine and terrestrial fossils have been reported from the Curtis Member/Formation. Because the Curtis Member is marine in origin at DINO, the marine assemblage is more relevant for the purposes of this document. The marine fossil assemblage of the Curtis Member may include dinoflagellates (Evitt 1961), foraminifera, coral-like organisms, brachiopods, bivalves, ammonite and belemnite cephalopods, gastropods, serpulid worms, possible crustaceans, ostracodes, crinoids, echinoids, holothurians (sea cucumbers), invertebrate trace fossils and fragments of vertebrates (Hoggan 1970). It is likely that at least some of Hoggan's fossils, which come from the Flaming Gorge area, pertain to the Redwater Member instead; the section lacks a basal resistant sandstone unit, and the lithology more closely resembles the Redwater Member. However, in the absence of first-hand evaluation, it would be foolish to attempt a division. Fossils that can be specifically attributed to the Curtis Member include bivalves, belemnites, gastropods (Untermann and Untermann 1954), crinoids, echinoids, and trace fossils (Pipiringos and Imlay 1979).

Stump Formation: Redwater Member (Upper Jurassic)
The Redwater Member of DINO is a slope-forming unit mostly composed of greenish shale and siltstone, with interbeds of sandstone and fossiliferous limestone. This unit is 22 to 40 m (70 to 130 ft) thick in the area, thinning west to east (Hansen et al. 1983; Gregson et al. 2010). The upper contact with the Morrison Formation is disconformable (Gregson et al. 2010). The lower part of the member represents a transgression, while the upper represents a regression (Brenner and Davies 1974). The sea had normal marine salinity and oxygenation (Feldmann and Titus 2006), and transgressed rapidly (Imlay 1947). The Redwater Member dates to the early Late Jurassic, approximately 159.5 to 156 Ma (Kvale et al. 2001).

The Redwater Member is probably fossiliferous at DINO because it is extensively fossiliferous in the area. Untermann and Untermann (1954) identified limestone beds in the Dinosaur Quarry area with abundant fossils, including bivalves, ammonites, belemnites, and gastropods; most of the outcrops of the Redwater Member in this study area are within DINO. Bivalves and belemnites have been found just east of the monument as well (Untermann and Untermann 1954). Fossils of three superficially dolphin-like ichthyosaurs (probably *Ophthalmosaurus*), mostly vertebrae, were found just west of DINO in this unit in a pipeline cut (Chure 1993; Gregson et al. 2010).

Overall, the Redwater Member has a very diverse fossil assemblage. Plants are represented by wood (Brenner and Davies 1974; Pipiringos and Imlay 1979). Invertebrates are abundant, and marine vertebrates are occasionally found (O'Keefe and Street 2009). Fossils are sufficiently abundant to form local coquina (shell hash)-like layers (Imlay 1947; Brenner and Davies 1973). Marine invertebrate remains may have been concentrated by storm events (Brenner and Davies 1973; Specht and Brenner 1979). The invertebrate fossil assemblage of the Redwater Member includes foraminifera (Loeblich and Tappan 1950), brachiopods (Brenner and Davies 1973), bivalves, ammonite and belemnite cephalopods (Imlay 1947), gastropods, scaphopods (Specht and Brenner 1979), the lobster-like decapod

crustacean *Eryma* (Wahl 2004), ostracodes (Swain and Peterson 1951, 1952), asteroids (Blake 1981), crinoids, echinoids, polychaete worm tubes, and other invertebrate trace fossils (Specht and Brenner 1979). Vertebrates known from the member include hybodont sharks (Wahl 2005), the ichthyosaur *Ophthalmosaurus* (formerly *Baptanodon*) (Massare et al. 2006), the small long-necked plesiosaurs *Pantosaurus* (O'Keefe et al. 2009) and *Tatenectes* (O'Keefe and Street 2009), the pliosaur (short-necked and large-skulled plesiosaur) *Megalneusaurus* (Wahl et al. 2007), a crocodylomorph (Marsh 1895), and indeterminate reptile bones (Specht and Brenner 1979).

It is likely that the first named reptile specimen from northeastern Utah came from the Redwater Member. In 1870, Othniel Charles Marsh collected a crocodylomorph humerus from Sheep Creek Canyon, 64 km (40 miles) northwest of DINO, which he later described as the type specimen of the new species *Diplosaurus nanus* (Marsh 1895; Norell and Storrs 1989; Bilbey and Hamblin 1992). While the area has outcrops of the Entrada Sandstone and both members of the Stump Formation (Bilbey and Hamblin 1992), a Redwater Member identification is most probable, given Marsh's use of the "*Baptanodon* beds" name (Norell and Storrs 1989). The specimen is now considered nondiagnostic and the species dubious (Norell and Storrs 1989). Closer to DINO are reports of marine reptiles from near Red Fleet Reservoir (Bilbey et al. 1990; Hamblin et al. 2000).

Morrison Formation (Upper Jurassic)

The Morrison Formation is a widespread heterogeneous and variegated unit. At DINO, these "rainbow beds" (Untermann and Untermann 1949a) include shale, claystone, siltstone, and sandstone, with lesser amounts of limestone and conglomerate; generally, lithology become finer up-section. It is between 200 and 300 m (650 and 1,000 ft) thick in the monument. Exposures can be found in the southwestern part of DINO on the south side of Split Mountain, an area near Rainbow Park and Island Park, and the Deerlodge Park area in the far east of the monument (Hansen et al. 1983; Gregson et al. 2010). Unsurprisingly, it forms both ridges and valleys (Gregson et al. 2010). The upper contact with the Cedar Mountain Formation is unconformable (Gregson et al. 2010; Kirkland et al. 2011a), but difficult to identify in locations where the Cedar Mountain Formation's Buckhorn Conglomerate Member is absent (Gregson et al. 2010). The Morrison Formation dates to the middle Late Jurassic, between about 155 and 148 Ma (Foster 2007); the oldest part of the unit at DINO has been dated to about 154.8 Ma (Kowallis et al. 1998).

The formation is often divisible into several members. At DINO, it is usually depicted as either undivided (Gregson et al. 2010; most maps) or as having two members: the Salt Wash and overlying Brushy Basin members (Turner and Peterson 1999; Foster 2003). More divisions may be present: Chure (2000a) distinguished two thin (both less than 10 m [33 ft] thick) units at the base of the formation, and assigned them to the Windy Hill and Tidwell members; these probably at least in part account for the basal "friable (crumbly) gray sandstone" mentioned by Gregson et al. (2010). Bilbey et al. (2005), in their description of pterosaur and dinosaur tracks from the Windy Hill Member of Flaming Gorge Reservoir, assigned the Windy Hill Member to the underlying Stump Formation instead of the Morrison Formation. The Windy Hill interval would represent the final stage of the marine deposition begun in the Stump Formation (Chure 2000a), while the rest are terrestrial, with a combination of lacustrine, fluvial, floodplain, and possible deltaic and eolian depositional environments (Untermann and Untermann 1949a). Sandstone beds represent fluvial channels, and shales and mudstones represent floodplains and bodies of water; bentonitic beds derive from ash erupted from volcanoes to the west (Gregson et al. 2010). The climate of the Morrison Formation is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004).

The Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites, fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans, ostracodes, and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods, gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians, lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods,

diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks (Turner and Peterson 1999; Foster 2003, 2007) and many of these genera are known from multiple specimens collected from DINO.

Morrison Formation paleontology at DINO: history

The first paleontological work in the DINO area was done by Othniel Charles Marsh, who would later become famous for his work on dinosaurs of the Morrison Formation in Colorado and Wyoming. In 1870, he collected a theropod tooth from the Morrison Formation just beyond what would become the northwestern boundary of DINO (Marsh 1871; Bilbey and Hall 1999; Chure 2000a, 2000b). It probably pertains to *Allosaurus* (Chure 2000b), and represents the second recorded dinosaur discovery from Utah, the first to be published, and the first record of theropods from the state (Bilbey and Hall 1999).

The Dinosaur Quarry site was discovered August 17, 1909 by Carnegie Museum of Natural History (CM or CMNH) paleontologist Earl Douglass (DINO Figure 5). Douglass would not only lead excavation at the site, but would help it become a national monument, and envisioned the *in situ* fossil relief concept that became the monument's leading attraction (Chure and McIntosh 1990). The first fossils he found were eight articulated tail vertebrae of an *Apatosaurus*, which are now recognized as part of the holotype (name-bearing specimen) of *Apatosaurus louisae*, on display at the Carnegie Museum. Between 1909 and 1922, museum personnel extracted over 320,000 kg (700,000 lbs.) of fossils from the Dinosaur Quarry, mostly from dinosaurs, but also from turtles and crocodylomorphs (Chure and McIntosh 1990). Carnegie Museum officials soon began to worry that the site would be claimed by speculators or attract competition, so they filed a claim on the land for its mineral rights. This was denied on the grounds that fossils were not strictly minerals. With this avenue closed, the direction then turned to the National Park Service, and on October 4, 1915, President Woodrow Wilson signed Dinosaur National Monument into existence (Chure and McIntosh 1990; Elder 1999).

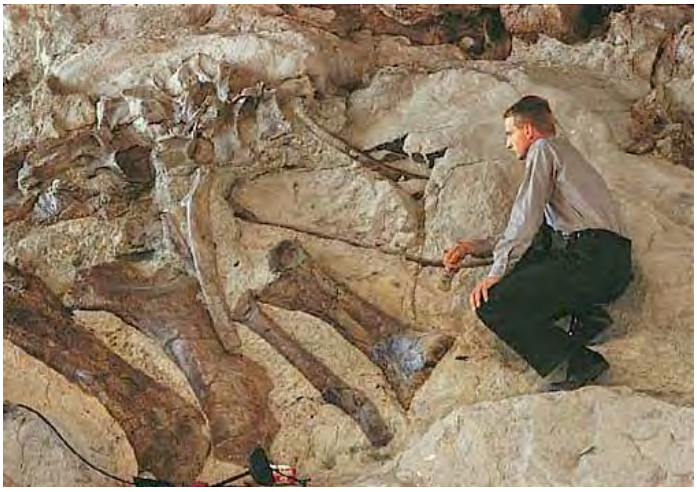


DINO Figure 5. Earl Douglass with Diplodocus *bones at the Dinosaur Quarry, circa 1923 (National Park Service photo collection).*

1922 was the last year of excavation for the Carnegie Museum (Chure and McIntosh 1990; Elder 1999). They were followed in 1923 by the National Museum of Natural History (USNM) and the University of Utah (UUVP or UMNH). Both institutions collected 33 crates of fossils, with the USNM collection including a skeleton of *Diplodocus* now on display there, and the UUVP collection including an *Allosaurus*, which was on display before being sent back to DINO in the early 1980s (Chure and McIntosh 1990). The University of Michigan also collected at the quarry before excavation ceased in 1925 (Elder 1999). With such large quantities of fossils already collected, it is understandable that Douglass faced skepticism when he proposed that an unexcavated section be prepared to show the bones in relief, *in situ*.

Very little was done at DINO between 1925 and 1952, with the exception of some non-paleontological excavation work at the quarry done at the suggestion of American Museum of Natural History paleontologist Barnum Brown (Chure and McIntosh 1990; Elder 1999). This work was performed between 1933 and 1938 by the laborers from the Civil Works Administration, then the Transient Relief Service, and finally the Works Progress Administration (Chure and McIntosh 1990). The quarry once again became active in 1953, when Douglass' plan was brought back to life as part of the wider Mission 66 initiative (Elder 1999). Work to expose bones on the quarry face began under the direction of Theodore White. Any doubt about whether scientifically important specimens were still present was answered with White's publication (White 1958) on a new partial skull of *Camarasaurus* (Chure and McIntosh 1990), the same year the visitor center was dedicated. In the end, over 1,400 bones were prepared for viewing on the quarry face (DINO Figure 6) (Elder 1999), and the *in situ* bone relief display was adopted by several other heavily fossiliferous sites,

such as Ichthyosaur–Berlin State Park in Nevada, the Mammoth Site in South Dakota, and the Dashanpu Dinosaur Museum in China (Chure and McIntosh 1990). The visitor center had to be replaced in the first decade of the 2000s due to structural damage brought on by underlying bentonitic shrink-swell clays (Chure 2010), and was reopened in October 2011.



DINO Figure 6. Fossils exposed on the wall of Dinosaur Quarry (NPS photo).

Paleontological work in the Morrison Formation of DINO began to go beyond Dinosaur Quarry in the 1980s. Surveying of DINO's Morrison Formation took place in the 1980s and 1990s. A preliminary survey in 1984 found more than 100 fossil localities, including what would become the two very productive microsites in Rainbow Park. A more extensive survey began in 1989, and within two field seasons had found over 260 localities. More than half had dinosaur bone, usually scrappy. The next most common find was petrified wood. The survey found that the Salt Wash Member had abundant fossils in its sandstone beds, but the material was limited to wood and dinosaur bone. The overlying Brushy Basin Member had much more diverse fossils, in sandstone and finer beds (Engelmann 1992). One of the best theropod specimens from the monument (DINO 11541) was found in the Salt Wash Member in 1990 by George Engelmann (Chure 2000a). Initially headless, its skull was located in 1996 by Ray Jones using a shielded gamma scintillation detector (Jones et al. 1998; Chure 2000a). It represents a new species of *Allosaurus* (Chure 2000a). A year later, a partial skeleton of an embryonic *Uteodon* (=*Camptosaurus*) was found high in the formation by Scott Madsen (Chure et al. 1994). Reinvestigation of old material has also yielded important discoveries, such as a partial skeleton of the rare theropod *Marshosaurus* that was collected in 1912 but not studied until the 1990s (Chure et al. 1997).

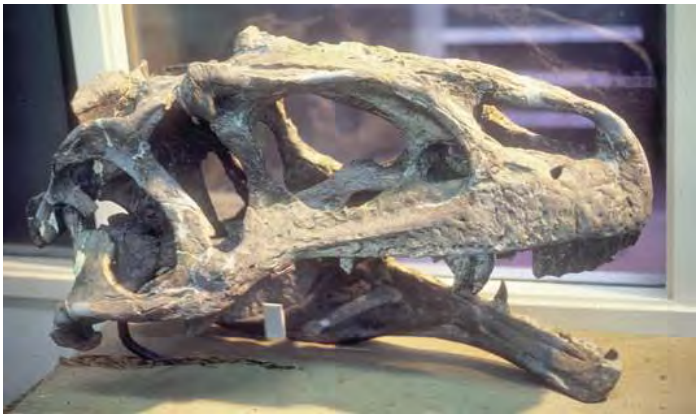
Morrison Formation paleontology at DINO: localities and material

The Morrison Formation at DINO is most famous for dinosaurs, but other fossils are not uncommon; for example, small vertebrate fossils may be encountered though minor disturbances of the surface, such as trenching for geological sections (Chure 1992). Plant fossils are historically poorly known at DINO (Ash 1993; Gregson et al. 2010), but are represented by root casts (Lawton 1977), microbial mounds (Bilbey et al. 1974), charophytes, spores, pollen (Chure 1992; Gregson et al. 2010), and body fossils of horsetails (Chure 1989), ferns (Chure 1992; Ash 1993), cycads, ginkgoes (Chure 1992; Gregson et al. 2010), czekanowskiales, and conifers (Chure 1992; Ash 1993; Gregson et al. 2010). Plants have been found as compressions and impressions, and petrified (Ash 1993). Petrified wood in and around DINO mostly pertains to conifers. Some specimens are logs up to 1 m (3 ft) in diameter (Untermann and Untermann 1954), or over 20 m (67 ft) long (Engelmann 1992). Invertebrates include bivalves, gastropods, conchostracans, ostracodes, possible insect borings on wood (Chure 1992), invertebrate burrows (Chure 1989), and dermestid beetle borings on dinosaur bones (Hasiotis et al. 1999). Vertebrates represented at DINO include the following: a bowfin; a pycnodontid ray-finned fish; unidentified ray-finned fish; the lungfish *Ceratodus*; at least five taxa of frogs including *Enneabatrachus* and *Rhadinosteus*; *Iridotriton* and a second unnamed genus of salamander; the turtles *Dinochelys* and *Glyptops*; the sphenodontians *Opisthias* and *Theretairus*; the lizards *Dorsetisaurus*, *Paramacellodus*, and *Schilleria*; a possible snake; the choristodere *Cteniogenys*; the crocodylomorphs *Eutretauranosuchus*, *Fruitachampsa*, *Goniopholis*, and *Hoplosuchus*; the theropod dinosaurs *Allosaurus* (two species), *Ceratosaurus*, *Koparion*, *Marshosaurus*, *Torvosaurus*, and an unnamed possible troodontid; the sauropods *Apatosaurus*, *Barosaurus*, *Camarasaurus*, *Diplodocus*, and possibly *Haplocanthosaurus*; the stegosaurid *Stegosaurus*; the ornithopods *Dryosaurus* and *Uteodon* (=*Camptosaurus*); the docodont mammal *Docodon*; the triconodont mammals *Priacodon*, *Triconolestes*, and an unnamed form; the multituberculate mammals *Ctenacodon* (three species), *Glirodon*, and *Psalodon*; an unidentified spalacotheriid mammal; the dryolestoid mammals *Amblotherium* (two species), *Araeodon*, *Dryolestes*, *Euthlastus*, and *Herpetarius* (Foster 2003; Gregson et al. 2010; *Iridotriton* after Evans et al. [2005]); and dinosaur eggshell (Bray and Hirsch 1998). There are also reports of undescribed unusual theropods, currently only documented in abstracts (Chure et al. 1993; Chure 1995). The great majority of the vertebrate diversity comes from three sites: Dinosaur Quarry, Rainbow Park Microsite DNM 94, and Rainbow Park Microsite DNM 96 (Foster 2003). Locality-level taxonomic lists for Morrison Formation dinosaurs can be found in Turner and Peterson (1999), and similar compilations for all vertebrates can be found in Foster (2003).

Dinosaur Quarry, in the Brushy Basin Member (Lawton 1977; Fiorillo 1994), is famous for the abundance, variety, and preservation quality of its fossils (Untermann and Untermann 1949a; Gregson et al. 2010). Some 5,000 bones have been found at the quarry. It covers an area of 1,700 m^2 (18,300 ft^2), giving a

density of 2.9 bones/m^2 (0.27 bones/ft^2) (Dodson et al. 1980). The setting was a braided channel in a well-drained floodplain (Dodson et al. 1980), with flow direction to the east (Gregson et al. 2010). The fossil accumulation is interpreted as the result of three or four closely-spaced depositional events (Lawton 1977; Fiorillo 1994). Taking into account the taphonomy of the specimens, the quarry's fossils probably accumulated on the order of six months to 19 years, most likely closer to the shorter extreme (Fiorillo 1994). Deposition occurred during brief periods of high flow, with short transportation distances for the bones (Lawton 1977). It has been suggested that the remains represent attritional and non-catastrophic mass mortality from droughts (Carpenter 2010). Carcasses would dry and decay during dry periods, and then be swept into the channel by high-precipitation events. The presence of dermestid beetle borings on dinosaurs bones indicates a few weeks to months between death and burial (Hasiotis et al. 1999).

Non-vertebrates from the quarry include root casts, petrified wood (Lawton 1977), carbonized logs, other plant fragments, the horsetail *Equisetum* (Chure 1989), bivalves (Lawton 1977), gastropods (White 1964), and dermestid borings (Hasiotis et al. 1999). Freshwater unionid bivalves are abundant in the bone-bearing channel sandstone, indicating that there was usually flowing water (Dodson et al. 1980). Vertebrates include the following (minimum number of individuals in parenthesis): *Opisthias* (1); *Dinochelys* (1); *Glyptops* (10); *Goniopholis* (3); *Ceratosaurus* (1); *Torvosaurus* (1); *Allosaurus* (8, including 3 juveniles; DINO Figure 7); *Diplodocus* (29, including 14 juveniles); *Apatosaurus* (16, including 4 juveniles and 1 hatchling); *Barosaurus* (5); *Camarasaurus* (22, including 6 juveniles; DINO Figure 8); ?*Haplocanthosaurus* (2); *Stegosaurus* (14, including 3 juveniles); *Dryosaurus* (4, including 2 juveniles); and *Uteodon* (=*Camptosaurus*) (8, including 3 juveniles) (Foster 2003, with number correction for *Uteodon* after Carpenter and Wilson [2008]). Sauropods are often found as segments of a skeleton. *Stegosaurus*, while common, is only found as isolated bones (White 1964).



DINO Figure 7. Allosaurus fragilis *skull from the Dinosaur Quarry (VLS).*



DINO Figure 8. A Camarasaurus *skull exposed at Dinosaur Quarry. Sauropod dinosaur skulls are historically rare (VLS).*

Two of the best-known specimens from the quarry are CMNH 11338, a nearly complete and articulated juvenile specimen of *Camarasaurus lentus* (Gilmore 1925a), and CMNH 3018, a nearly complete skeleton of *Apatosaurus louisae* which is one of the defining specimens of this genus thanks to the extensive monograph by Charles Gilmore (Gilmore 1936). This specimen is part of the *Apatosaurus* skull controversy because of CMNH 11162, a skull found near it. Holland thought that this skull was potentially that of CMNH 3018, which would give *Apatosaurus* a slender *Diplodocus*-like skull in opposition to the prevailing opinion that *Apatosaurus* had a squarish *Camarasaurus*-like skull. Gilmore (1936) dismissed Holland in part because he thought that Holland had gotten CMNH 11162 confused with CMNH 11161, a smaller *Diplodocus* skull found much farther away in the quarry; thus, he thought that the skull near CMNH 3018 was the smaller skull, which could not have belonged to the large skeleton. It later became apparent that Gilmore was the mistaken party, and CMNH 11162 is indeed the probable skull of CMNH 3018 (McIntosh and Berman 1975; Berman and McIntosh 1978). Other notable specimens from the quarry include: juvenile specimens such as a baby *Dryosaurus* (Carpenter 1994) and a juvenile *Stegosaurus* (Galton 1982); thirteen dinosaur skulls, from *Allosaurus* (two), *Apatosaurus* (one), *Camarasaurus* (five), *Diplodocus* (three), and *Dryosaurus* (two) (Chure and McIntosh 1990; Gregson et al. 2010); and the best specimens of *Dryosaurus* and *Uteodon* (=*Camptosaurus*) (Chure and McIntosh 1990).

Two microvertebrate sites, now known as Rainbow Park Microsites DNM 94 and 96 (Turner and Peterson 1999; Foster 2003), were discovered in Rainbow Park by Robert West in 1984 (Chure 1989). They come from the upper Brushy Basin Member, with DNM 94 below DNM 96 (Turner and Peterson 1999). DNM 94 represents a mud slurry deposit and DNM 96 represents quiet water deposition (Chure 1989). The DNM 94 assemblage includes ray-finned fish, *Glyptops*, *Opisthias*, lizards, a possible snake, *Goniopholis*, *Allosaurus*, *Koparion*

(Foster 2003), an unnamed troodontid (Turner and Peterson 1999), *Diplodocus*, *Camarasaurus*, *Stegosaurus*, *Dryosaurus*, *Uteodon* (=*Camptosaurus*), *Docodon*, unidentified multituberculates, and unidentified mammals (Foster 2003). The DNM 96 assemblage includes ray-finned fish, *Enneabatrachus*, *Rhadinosteus*, unidentified frogs, *Iridotriton*, *Glyptops*, *Opisthias*, *Theretairus*, lizards, *Eutretauranosuchus*, *Goniopholis*, *Camarasaurus*, *Stegosaurus*, unidentified multituberculates, *Priacodon*, *Triconolestes*, *Euthlastus*, *Dryolestes*, and unidentified mammals (Foster 2003; *Iridotriton* after Evans et al. [2005]). Over one hundred mammal teeth were found at these sites (Engelmann et al. 1993). Fossils from at least one of the two sites (unspecified) include possible seeds (Engelmann et al. 1989), invertebrate burrows, freshwater gastropods, eggshell fragments, and possible eggs (Chure 1989). Early reports of shark material (for example Chure [1989]) appear to have been in error, as sharks are not identified in more recent reports (Foster 2003; Gregson et al. 2010).

A number of taxa have been named from the Morrison Formation of DINO (see also DINO Table 2). In chronological order, they are: the sauropod dinosaur *Apatosaurus louisae*, based on CMNH 3018 from Dinosaur Quarry (Holland 1915; Gilmore 1936); the turtle *Glyptops utahensis*, based on CMNH 3412 from Dinosaur Quarry (Gilmore 1916; now considered a synonym of *G. plicatulus*); the sauropod *Uintasaurus douglassi*, based on CMNH 11069 from Dinosaur Quarry (Holland 1919, 1924a; now considered a synonym of *Camarasaurus lentus*); the small crocodylomorph *Hoplosuchus kayi*, based on CMNH 11361 from a road excavation near Dinosaur Quarry (Gilmore 1926); the sauropod *Camarasaurus annae*, based on CMNH 8942 from Dinosaur Quarry (Ellinger 1950; now considered a synonym of *C. lentus*); the turtle *Dinochelys whitei*, based on DINO 986 to 991 from Dinosaur Quarry (Gaffney 1979); the possible troodontid theropod *Koparion douglassi*, based on DINO 3353 from Rainbow Park Microsite DNM 94 (Chure 1994); the frog *Rhadinosteus parvus*, based on DINO 14693 from Rainbow Park Microsite DNM 96 (Henrici 1998a); the triconodont mammal *Triconolestes curvicuspis*, based on DINO 10780 from Rainbow Park Microsite DNM 96 (Engelmann and Callison 1998); the lizard *Schilleria utahensis*, based on DINO 14720 from DNM 375 (Evans and Chure 1999); the multituberculate mammal *Glirodon grandis*, based on DINO 10822 from an unspecified locality (Engelmann and Callison 1999); the salamander *Iridotriton hechti*, based on DINO 16453a-b from Rainbow Park Microsite DNM 96 (Evans et al. 2005); and the ornithopod dinosaur *Uteodon aphanoecetes*, based on CMNH 11337 from Dinosaur Quarry (McDonald 2011; began as *Camptosaurus aphanoecetes* of Carpenter and Wilson 2008). In addition, as mentioned above, an as-yet unofficially-named species of *Allosaurus* is represented by an excellent skull and skeleton (DINO 11541) discovered in Salt Wash Member beds at DINO (site DNM 116) (Chure 2000a), and there are multiple undescribed amphibian and mammal taxa (Gregson et al. 2010).

Cedar Mountain Formation (Lower Cretaceous)

The Cedar Mountain Formation is a heterogeneous unit similar to the Morrison Formation. In the DINO area, it includes two main divisions: the basal Buckhorn Conglomerate Member, composed of light gray to tan sandy pebble conglomerate, pebbly sandstone, and sandstone; and an unnamed overlying unit composed of purple, gray, and green shale, mudstone, and minor limestone (Rowley et al. 1979a; Currie 1998; Kirkland and Madsen 2007; Currie et al. 2008; Kirkland et al. 2011a). The formation is generally 30 to 60 m (100 to 195 ft) thick, with the basal conglomerate contributing up to 15 m (50 ft), but both the conglomerate and the formation are locally absent. Like the Morrison Formation, the Cedar Mountain Formation forms both valleys and ridges, with the Buckhorn Conglomerate forming a prominent hogback where present (Gregson et al. 2010). Like the other Cretaceous units, exposures can be found in the southwestern part of DINO on the south side of Split Mountain, near Rainbow Park and Island Park, and the Deerlodge Park area in the far east of the monument (Hansen et al. 1983). The upper contact with the Dakota Sandstone is disconformable (Hansen et al. 1983). The Cedar Mountain Formation was primarily deposited in fluvial and lacustrine settings (Gregson et al. 2010). The Buckhorn Conglomerate Member represents a northeast-flowing river system out of the San Rafael Swell (Kirkland and Madsen 2007). The climate became more humid over time, with high temperatures (Garrison et al. 2007). Deposition of the formation ended during the Early Cretaceous at DINO; locality DNM 16 (the *Abydosaurus* site; see below) is no older than 104.46 ± 0.95 Ma (Chure et al. 2010), and the overlying Dakota Sandstone includes a date of 101.4 ± 0.4 Ma (Kirkland et al. 2011a). At the base of the formation, the Buckhorn Conglomerate is usually thought of as dating to the Barremian stage of the Early Cretaceous (Dickinson and Gehrels 2008), approximately 130 to 125 Ma. However, its age is controversial (Kirkland and Madsen 2007). There are places where it appears to intertongue with the Brushy Basin Member of the Morrison Formation, which would make it at least partially Late Jurassic in age (Kirkland et al. 1999).

The Cedar Mountain Formation name was not used at DINO during the early geological work (Untermann and Untermann 1949a, 1950, 1954). Untermann and Untermann (1954) reported the presence of 4 m (13 ft) of white chert and jasper pebble conglomerate and 22 m (72 ft) of shale in their upper Morrison Formation that could represent the Buckhorn Conglomerate and upper Cedar Mountain Formation, which would account for at least some of the formation. The upper Cedar Mountain Formation may be divisible, although this hasn't been done formally. Several names from members established in central Utah have been applied. These include from oldest to youngest the Yellow Cat Member (Chure et al. 2010), the Ruby Ranch Member (Maxson et al. 2003; Tremain 2003; Maxson 2011), and the Mussentuchit Member (Chure et al. 2010) (see the Arches National Park [ARCH] and Capitol Reef National Park [CARE] summaries in this report for discussions). However, Kirkland et al. (2011a) cautioned against applying these terms to

the Cedar Mountain Formation of northeastern Utah. In particular, the portion that has been correlated to the Mussentuchit Member is several million years older than that member, and the time span the Mussentuchit Member occupies elsewhere in Utah is occupied by the Mowry Shale in the DINO area (Kirkland et al. 2011a).

The Cedar Mountain Formation of DINO is fossiliferous. It is best-known for a sauropod dinosaur bonebed (DNM 16) discovered near the Dinosaur Quarry. This site has yielded four skulls of the *Brachiosaurus*-like sauropod *Abydosaurus mcintoshi*, including the holotype specimen (DINO 16488) (DINO Figure 9). The site also includes a number of sauropod postcranial bones that probably belong to the same species (Chure et al. 2010). Other fossils from DINO include microbial mats, root casts, gastropods, ostracodes, burrows (Kosmidis et al. 2002), and an undescribed dromaeosaurid (Gregson et al. 2010). Abundant trace fossils are also known from the Cedar Mountain Formation of DINO and the immediate vicinity. These include root traces, worm burrows, bivalve burrows, crayfish burrows, ant and termite nests, wasp cocoons, dung beetle balls, other invertebrate traces, and vertebrate tracks. Different trace fossils have been found to correspond to different facies (Tremain et al. 2004). A partial skeleton of an ornithopod similar to *Tenontosaurus* was recently recovered from a site northwest of DINO along a pipeline corridor (Kirkland et al. 2011b). Fossils of a dromaeosaur and ankylosaur are known from the Vernal area (Bilbey and Hamblin 1992).



DINO Figure 9. Holotype skull of Abydosaurus mcintoshi *(DINO 16488) (VLS).*

Regionally, the Buckhorn Conglomerate Member is sparsely fossiliferous, but it has yielded cycad stems (Furniss and Tidwell 1972), petrified logs, poorly preserved mollusk shells, fragmentary dinosaur bones (Stokes 1950), a partial skeleton of an ankylosaurian dinosaur (which would suggest a Cretaceous age; Kirkland and Madsen 2007), and reworked Paleozoic fossils (Kirkland and Madsen 2007). In the interest of space, and because there is some question about the correlations, the fossil assemblages of the three units sometimes thought to be present at DINO (the Yellow Cat, Ruby Ranch, and Mussentuchit member) are not included here. They are documented in the ARCH and CARE summaries in this report. The Cedar Mountain Formation above the Buckhorn Conglomerate Member is known for its sequence of vertebrate assemblages. These usually include hybodont sharks, ray-finned fish, lungfish, sphenodontians, crocodylomorphs, large and small theropods, various types of sauropods, diverse armored dinosaurs, ornithopods, and sometimes mammals (Kirkland et al. 1999; Kirkland and Madsen 2007). Diverse plants are also known from the Cedar Mountain Formation (Tidwell et al. 1976; Thayn et al. 1983).

Mowry Shale (Upper Cretaceous)

The Mowry Shale is the oldest dominantly marine Cretaceous unit at DINO. Depending on the source, it may be identified as the Mowry Shale Member of the Mancos Formation in the DINO area (see for example most of the geologic maps of DINO quadrangles from the 1970s and 1980s). As noted by Molenaar and Wilson (1990), historically, north of the Uinta Mountains this unit and the Frontier Formation are given formational rank, and south of the mountains they are members of the Mancos Shale. In the DINO area, it is composed of hard dark gray silica-rich shale with bentonite. It weathers to a lighter silvery gray, typically into thin sheets (fissile), and ranges from 10 to 65 m (33 to 220 ft) thick, thinning from west to east (Hansen et al. 1983; Gregson et al. 2010). Untermann and Untermann (1949a, 1954) suggested that an interval of softer shale at the base could represent the Thermopolis Shale.

The Mowry Shale was deposited offshore within the Western Interior Seaway (Franczyk et al. 1992). It belongs to the Mowry transgression-regression cycle (Cole 1987). During this time, the DINO area was under the northern part of the seaway (Stewart and Hakel 2006), and it is uncertain if the northern and southern parts of the Seaway system were connected; they may have been connected near the end of Mowry Shale deposition (Eicher 1961). Water depth in the area may have been greater than 150 m (490 ft) (Byers and Larson 1979). The upper contact with the Frontier Formation in this area of Utah is disconformable, and eliminates several million years of marine deposition of the early Late Cretaceous (parts of the Cenomanian and Turonian stages) (Molenaar and Wilson 1990; Franczyk et al. 1992). While marine conditions are usually excellent for deposition, in this case it is thought that a submarine structural high appeared in the Vernal area before the hiatus, locally interrupting the sediment record for a few million years (Molenaar and Wilson 1990; White et al. 2002). Often dated to the Early Cretaceous in older references, it is now known to date to the Late Cretaceous (Gregson et al. 2010). Ash beds have permitted the formation to be dated precisely. The base of the formation dates to around 98.52 ± 0.41 Ma (Oboh-Ilkuenobe et al. 2007), and the top is dated to 97.17 ± 0.69 Ma (Vakarelov et al. 2006).

The Mowry Shale is fossiliferous at DINO. One of its primary constituents is fish scales (Untermann and Untermann 1949a, 1954; Anderson and Kowallis 2005). Shark teeth have been found in the formation within the monument (Hunt et al. 2006). Two sites in the monument include constituents of a northeastern Utah Mowry Shale fossil assemblage of angiosperm leaves, teleosts (derived ray-finned fish), fish scales, possible plesiosaurs, and possible crocodylomorphs (Stewart et al. 1994). Untermann and Untermann (1954) reported an ammonite from the Dinosaur Quarry area, and mentioned fish scales in the Island Park area and the Lily Park area east of the monument. Nearby, the type locality for the crinoid *Uintacrinus* is probably just west of DINO (Bilbey and Hall 1999). Isolated occurrences of vertebrate fossils have been found 6 km (4 miles) north of Vernal on Highway 191. These include shark teeth, teleost jaw and vertebral fragments, teleost teeth (mostly of *Enchodus*), and some possible teleost micro-coprolites (Anderson and Kowallis 2005).

The Mowry Shale has a small but growing fossil assemblage. Non-animal organisms are represented by foraminifera, radiolarians, dinoflagellates, spores, pollen (Oboh-Ilkuenobe et al. 2007), coal (Uyger and Picard 1985), angiosperm leaves (Stewart et al. 1994), and other plant remains (Uyger and Picard 1985). Invertebrates include bryozoans (Maione 1971), bivalves (Gregson et al. 2010), ammonites (Maione 1971; Molenaar and Wilson 1990), impressions of a squid-like animal (Connely and Talbot 2005), burrows, and local bioturbation (Davis et al. 1989; Merewether et al. 2010). The bottom sediments often show little evidence of burrowing animals, indicating anaerobic bottom waters; DINO is within the zone of little benthic activity (Byers and Larson 1979). Vertebrates are represented by sharks and other cartilaginous fish (Cicimurri 2004; Hamblin et al. 2000; Gregson et al. 2010), teleosts (typically represented by scales, teeth, and isolated bones) (Hamblin et al. 2000), the ichthyosaur *Platypterygius americanus* (Nace 1939; Massare and Dain 1989), a long-necked elasmosaurid plesiosaur, a short-necked plesiosaur (possibly the polycotylid *Trinacromerum*), a crocodylomorph (Massare and Dain 1989), a pterosaur (Bennett 1989), and possible vertebrate trace fossils (Burford 1985; Connely and Talbot 2005; Trumbull and Connely 2010). Fish scales and bone fragments are abundant (Maione 1971; Molenaar and Wilson 1990).

Frontier Formation (Upper Cretaceous)

The Frontier Formation (sometimes called the Frontier Sandstone Member of the Mancos Shale) is a mixed terrestrial/marine unit with heterogeneous rocks. In the DINO area, it includes informal lower and upper sections separated by a shale interval up to 15 m (50 ft) thick. The lower part is composed of lime-rich shale, silty shale, and siltstone. This interval is fossiliferous, and is dark gray when fresh, but weathers to a light yellowish gray. It ranges from 24 to 40 m (80 to 130 ft) thick. The upper part is composed of light brown to gray lime-rich sandstone, and ranges from 24 to 44 m (80 to 145 ft) thick. The lower part is nonresistant, while the upper part forms ridges. Overall, the formation was deposited near the coastline in environments such as deltas, swamps, and offshore sandbars (Gregson et al. 2010). It includes three tongues of continental material interposed from the west into the Western Interior Seaway rocks to the east (Ryer 1977); in the DINO area, only one of these regression-transgression cycles is present (Molenaar and Wilson 1990). Much of the formation in the DINO area was part of a large delta system known as the Vernal Delta (Maione 1971). This was not a true delta, but a number of individual deltas that formed on the uplift that had appeared a few million years prior. The Vernal Delta jutted far out into the Western Interior Seaway, reaching as far west as the DINO area (Ryer and Lovekin 1986). Unlike the underlying Mowry Formation, the northern and southern parts of the Seaway were connected during Frontier Formation deposition (Scott et al. 2001b). Longshore drift moved south along the coast while other currents moved north (Ryer 1976).

The Frontier Formation is stratigraphically complex (Ryer 1976, 1977; Ryer and Lovekin 1986; Merewether et al. 1979). It includes a number of internal hiatuses (Vakarelov et al. 2006; Merewether et al. 2010). The formation existed for different amounts of time in different areas. In the DINO area, it dates to the latter half of the Turonian stage (Maione 1971; Molenaar and Wilson 1990; Franczyk et al. 1992; Johnson 2003), about 91 to 89 Ma. Older rocks are missing because of the uplift that disrupted deposition during the preceding few million years. The Frontier Formation of the DINO area appears to correlate, at least in part, to rocks of the Tununk and overlying Ferron members of the Mancos Shale in central Utah (Untermann and Untermann 1954; Ryer and Lovekin 1986; Molenaar and Wilson 1990; Johnson 2003); see the ARCH and CARE summaries in this document for more information on these units.

The Frontier Formation is fossiliferous within DINO. The basal part of the formation is conspicuously fossiliferous (Sumison 1976), and yields petrified wood, bivalves, ammonites, and gastropods (Gregson et al. 2010). The upper part has yielded coal and invertebrate burrows (Gregson et al. 2010). Mollusk fossils are often found in "cannonball" calcareous (lime-rich) concretions (Untermann and Untermann 1954). Shark teeth have been recovered from anthills in the monument (Gregson et al. 2010). In the Dinosaur Quarry area, the formation includes fossiliferous concretions and examples of plant stem casts, deciduous leaf impressions, and bivalve casts. In the Island Park area, the Frontier Formation has yielded bivalves, ammonites, gastropods, and shark teeth (Untermann and Untermann 1954). Bivalves and shark teeth have been found at Lily Park just east of the monument (Untermann and Untermann 1954). Coal beds and plant fossils have also been found in the area (Untermann and Untermann 1949a). Maione (1971) reported a fossiliferous site just outside of the monument on the south side of Split Mountain. This locality, in sandstone deposited as part of a barrier island, has wood impressions, bivalve molds, the giant bivalve *Inoceramus*, the shrimp burrow *Ophiomorpha*, and other

trace fossils. To the northwest, dinosaur tracks have been found in the Frontier Formation at Red Fleet State Park, and include three-toed ornithopod tracks and possible four-toed tracks (Hamblin et al. 2000).

Overall, the Frontier Formation includes diverse fossils, befitting its transitional coastal origin. Palynomorphs and inorganic microfossils include microspheres (Upshaw 1963), foraminifera (Young 1951; Maione 1971; Molenaar and Wilson 1990), charophytes (Jones 1954), dinoflagellates and other microplankton, spores, and pollen (Upshaw 1964). The formation has a significant floral assemblage, first collected by the explorer John C. Frémont in 1843. It includes *Equisetum*, ferns, cycadophytes, conifers, and monocot and dicot angiosperms (Knowlton 1917; Berry 1929; Andrews and Pearsall 1941). Coal and wood are common (Uygur and Picard 1985; Moslow and Tillman 1986; Merewether et al. 2010), and root traces have also been found (Moslow and Tillman 1986). Invertebrates are represented by bryozoans (Merewether and Cobban 2008), brachiopods (Merewether et al. 2010), bivalves, ammonites, gastropods (Young 1951; Upshaw 1964; Merewether et al. 1984; Merewether and Cobban 2008), nonmarine ostracodes (Jones 1954), crabs (Merewether and Cobban 2008), echinoids (Upshaw 1964; Merewether and Cobban 2008), and trace fossils, including worm tubes (Merewether and Cobban 2008), *Ophiomorpha*, other burrows (Merewether et al. 1979; Molenaar and Wilson 1990), and general bioturbation (Maione 1971). Vertebrates are limited, and aside from the shark teeth and dinosaur tracks of the DINO area, many records are from Frontier Formation rocks not present at the monument. The Frontier Formation of Wyoming has vertebrates including sharks, rays, gars, pycnodont fish, turtles, squamates, and dinosaurs (Breithaupt 1985). Vertebrates including plesiosaurs, crocodylomorphs (Williston 1905), and two armored dinosaur specimens, are known from the Belle Fourche Member of the lower Frontier Formation (Carpenter and Kirkland 1998), which is not present at DINO (Molenaar and Wilson 1990). Rocks from the Vernal-area hiatus in northern Utah have produced vertebrate fragments and eggshells (Eaton 2004).

Mancos Shale (Upper Cretaceous)

The Mancos Shale is a marine unit largely made up of shale with some sandy, silty, or carbonate-rich layers. In the DINO area, it is mostly composed of dark gray silty or clayey lime-rich shale that weathers light gray or yellow. There are also some bentonitic and limestone layers in the lower part, and siltstone and sandstone beds in the upper part (Hansen et al. 1983; Gregson et al. 2010). Only the lower part of the formation can be found in DINO (Gregson et al. 2010). About 150 m (500 ft) are present in eastern DINO (Rowley et al. 1979b), and 300 m (1,000 ft) in western DINO (Rowley et al. 1979a), but the full section can be 1,500 to 1,700 m (4,920 to 5,580 ft) thick in the immediate vicinity (Gregson et al. 2010). The upper contact with the Bishop Conglomerate is disconformable (Gregson et al. 2010).

The Mancos Shale is divided into a number of different members depending on location, but with the exception of authors who included the Mowry and Frontier formations in the Mancos Shale, this has not been done at DINO. The lower Mancos Shale of the DINO area is equivalent at least in part to the Blue Gate Shale Member found to the south and west (Molenaar 1991; Johnson 2003) (see the ARCH and CARE summaries in this document for more information on the Blue Gate Shale). A sandy unit sometimes called the Mancos B or Prairie Canyon Member interrupts the Mancos Shale in northwestern Colorado (Kass 1999; Johnson 2003). The part of the Mancos Shale below this sandy interval was deposited during the Coniacian and early to middle Santonian stages (Johnson 2003), between roughly 89 and 85 Ma. This probably approximates the age of deposition for the truncated DINO section.

The Mancos Shale is fossiliferous in the DINO area, and is probably fossiliferous within the monument. Untermann and Untermann (1954) found bivalves and ammonites in the Mancos Shale in the Dinosaur Quarry area. This study area does include some exposures of the Mancos Shale outside of DINO, but given that bivalves and ammonites are fairly abundant in the area (Untermann and Untermann 1949a), it seems more prudent to regard the formation as fossiliferous within the monument until proven otherwise instead of the opposite. Coniacian-age ammonites and bivalves have been found in the lower Mancos Shale on Ashley Creek, 8 km (5 miles) southeast of DINO, and Brush Creek, 11 km (7 miles) northwest of DINO (Kennedy and Cobban 1991).

Diverse marine fossils are known from the Mancos Shale across the Colorado Plateau, with the most abundant being ammonites and bivalves. Marine microorganisms and invertebrates are represented by acritarchs, dinoflagellates (Crowley et al. 1994), nannoplankton (Cohen 1959), radiolarians (Olesen 1991), foraminifera (Bagshaw 1977; Leckie et al. 1997), corals (Ball et al. 2010), brachiopods (Leckie et al. 1997), bryozoans (Hill 1982), ammonites, bivalves (Merewether et al. 2006; Ball et al. 2010), gastropods (Leckie et al. 1997), crustaceans (Kues 1980; Scott et al. 2001a), ostracodes (Bagshaw 1977), crinoids (Leckie et al. 1997), echinoids (Hill 1982), and worm tubes and other invertebrate trace fossils (Leckie et al. 1997). Marine vertebrates are represented by sharks and other cartilaginous fish (Ekren and Houser 1965; Leckie et al. 1997), teeth, scales, and bones of bony fish (Leckie et al. 1997), turtles (Irby 1995), plesiosaurs (Carter 1991; Spielmann and Lucas 2006), mosasaurs (marine reptiles related to monitor lizards) (Kass 1999; Spielmann and Lucas 2006), and sea birds (Lucas and Sullivan 1982). Terrestrial fossils are present near the western margin of the continental seaway, and include terrestrial microfossils (spores, and pollen from gymnosperms and angiosperms) (Crowley et al. 1994), root traces, coal, leaves (Hill 1982), ferns, gymnosperms, angiosperms (Young 1987), crocodilian bones (Irby 1995), dinosaur bones (Hill 1982) and at least one partial dinosaur skeleton (Lucas et al. 2006b).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary sediments of DINO include a variety of sediments ranging from silt to boulders in size. Three major processes are responsible for their deposition: flowing water formed alluvial and fluvial deposits, wind deposited sand in sheets and dunes, and gravity is responsible for talus slopes and large landslides throughout the monument (Hansen et al. 1983; Gregson et al. 2010). Quaternary deposits can reach thicknesses of 15 to 18 m (50 to 60 ft) in DINO (Untermann and Untermann 1949a; Sumison 1976).

Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). DINO's Quaternary fossil resources, a series of packrat middens, fall into the latter category. Summaries of Quaternary fossil sites in Utah and the vicinity include Anderson (1974), Walker (1987), Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008). Reworked Paleozoic and Mesozoic fossils may also be found, and have been found on the Quaternary-age Jensen pediment near Vernal (Bilbey 1992b).

Packrat middens are collections of plant material and food waste constructed by packrats (*Neotoma* spp.) and cemented by their viscous urine. They can be well-preserved in dry caves, and illustrate the environment of the builder's foraging range. Middens are important tools for reconstructing the ecology and climate of the late Pleistocene and Holocene of the southwestern United States (Strickland et al. 2001). Five packrat middens have been described from Sand Canyon Alcove in eastern DINO (Sharpe 1991, 2002; Anderson et al. 2000). Of the five sampled middens, one was modern, and the other four yielded dates of 9,870 ± 80 radiocarbon years before present (11,620–11,160 calendar years before present), 9,050 ± 120 radiocarbon years before present (10,510–9,780 calendar years before present), 8,460 ± 100 radiocarbon years before present (9,630–9,230 calendar years before present), and 3,000 ± 100 radiocarbon years before present (3,400–2,920 calendar years before present) (Sharpe 1991, 2002) (Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 [http://calib.qub.ac.uk/calib/] where appropriate). Sharpe (2002) found that the location underwent a transition from a cool mesophytic community to a more xerophytic and warmer-adapted community early in the Holocene, between the 9,050- and 8,460-radiocarbon-year middens. Blue spruce, limber pine, and common juniper are not found after the 9,050-radiocarbon-year midden, and the former two taxa are not present at DINO today. The modern pinyon-juniper woodland was present before the construction of the youngest fossil midden. The floral assemblage indicates that since the oldest midden was built, the January temperature has increased by 1 to 3 °C, the July temperature has increased by 3 to 10 °C, the January precipitation has decreased by half, and the July precipitation has decreased by two-thirds (Sharpe 2002). The only other paleoecological site in the vicinity is another fossil midden series, this one from Dutch John Mountain, about 48 km (30 miles) northwest of the northern tip of DINO. This series is substantially longer, dating back 40,000 years (Lyford et al. 2003; Jackson et al. 2005). Because DINO has caves (Graham 2006), it is also possible that Quaternary cave fossils will be found in the monument in the future. Santucci et al. (2001) presented an overview and cited selected examples of National Park Service fossils found in caves.

There are relatively few Pleistocene- or Holocene-age bone sites within about 100 km (60 miles) of DINO. Faunmap (http://www.ucmp.berkeley.edu/faunmap/index.html) and the Neotoma database (http://ceiwin5.cei.psu.edu/weborb30/myapps/neotoma/#) show about ten Quaternary fossil sites within this radius in northeastern Utah, northwestern Colorado, and south-central Wyoming. Most of these are archeological sites from the Fremont culture, which existed in the area between about 1,400 and 500 years ago (Bilbey 1992a). Two bison skulls thought to pertain to the extinct *Bison antiquus* have been found a few km/miles outside of DINO (Untermann and Untermann 1954). Of the other pre-Fremont sites, two have extinct mammals. Pine Spring, about 50 km (31 miles) northwest of the northern tip of DINO in Wyoming, has fossils of mountain sheep, bison, camels, and pronghorns (Sharrock 1966; Walker 1987). It is also a Paleoindian site, with a date of 9,695 ± 195 radiocarbon years before present (11,720–10,500 calendar years before present) (Frison 1978). The Stauffer Chemical Company site, about 80 km (50 miles) northwest of DINO in Wyoming, has camel fossils (Anderson 1974; Walker 1987). A third site, the Maxon Ranch site, about 45 km (28 miles) northwest of DINO in Wyoming, has bones from birds, rodents, rabbits, felid and mustelid carnivorans, bison, and pronghorns (Harrell and McKern 1986). This Early Plains Archaic culture site dates to 6,480 ± 90 radiocarbon years before present (7,570–7,250 calendar years before present) (Frison 1978).

Finally, Pleistocene gastropods are not uncommon in the region. A few specific reports include: the valley of the White River near Meeker, Colorado, 70 km (43 miles) southeast of DINO (Evanoff 1983); near the confluence of Antelope Creek and the Duchesne River, about 80 km (50 miles) southwest of DINO (Reheis et al. 1991); and near Craig, Colorado, 82 km (51 miles) east of DINO (Reheis et al. 1991).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within DINO. However, they are known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from one or more of them.

Uinta Mountain Group (Neoproterozoic)

The Uinta Mountain Group is a thick unit restricted to the Uinta Mountains and the immediate vicinity (Condie et al. 2001). Within DINO, about 2,130 m (7,000 ft) of the uppermost part are exposed, a relatively small fraction of a total thickness approaching 7,000 m (20,000 ft) (Gregson et al. 2010). Most of the DINO section is composed of red quartz-rich sandstone, with some shale and siltstone intervals, some of which are thick enough to be mapped (Hansen et al. 1983; Gregson et al. 2010). Division of the Uinta Mountain Group has been mostly informal, with distinct nomenclature for the eastern and western Uinta Mountains (Condie et al. 2001; Dehler et al. 2005, 2007; Nagy and Porter 2005). Historically, the easternmost Uinta Mountain Group, as exposed in DINO, has been mapped as undivided (Hansen et al. 1983; Sprinkel 2006); the presence of shaly intervals at DINO offers some potential for correlation. The Uinta Mountain Group makes up much of the bedrock in the northern prong of DINO, continuing farther south in the Canyon of Lodore (Hansen et al. 1983; Gregson et al. 2010).

The Uinta Mountain Group was first thought to be as old as 1100 Ma (Vidal and Ford 1985), but is now known to be substantially younger. The lower part of the group predates 766.4 Ma (Dehler et al. 2010), and the upper part predates rifting around 700 Ma, with a Middle Neoproterozoic time frame of 770 to 742 Ma suggested for the group (Dehler and Sprinkel 2005; Dehler et al. 2007). Deposition occurred without lengthy hiatuses (Dehler et al. 2007). The upper contact with the Lodore Formation is an angular unconformity representing a substantial depositional hiatus combined with structural tilting (Gregson et al. 2010).

Deposition occurred in a basin that existed where the Uinta Mountains exist today (Hansen 1986a), east of a shallow sea (Condie et al. 2001). The eastern rocks of the Uinta Mountain Group were generally deposited as a braided river system, flowing to the south and southwest, while the western rocks include deltaic and marine deposition (Dehler et al. 2007). At least two extensive marine incursions reached the eastern half of the Uinta Mountain Group's basin (Dehler et al. 2010). The climate was subtropical to tropical (Condie et al. 2001).

The Uinta Mountain Group is fossiliferous, but fossils have not yet been reported from DINO (Gregson et al. 2010). Fossils are not yet as well known from the eastern section as they are from the west, and in general Uinta Mountain Group assemblages are not as diverse as those of contemporaneous units, perhaps suggesting a more hostile environment (Nagy and Porter 2005). The fossil assemblage of the group includes filaments, the smooth-walled acritarch *Leiosphaeridia*, the aggregate *Bavlinella*, ornamented acritarchs, vase-shaped microfossils, the macroscopic disc-like compression fossil *Chuaria* (Nagy and Porter 2005), and possible organic sedimentary structures known as sandolites (sand stromatolites) (Filkorn 2009). The first three types of fossils are most common (Nagy and Porter 2005). In the east, filaments, *Leiosphaeridia*, *Bavlinella*, ornamented acritarchs, and possible vase-shaped microfossils have been found (Sprinkel and Waanders 2005; Nagy and Porter 2005; Dehler et al. 2007). Most fossil discoveries for the eastern units have been made in the Flaming Gorge area (Nagy and Porter 2005).

Moenkopi Formation (Lower Triassic)

The Moenkopi Formation is a heterogeneous unit present in several western states, and is known for its redbeds. The Lower Triassic redbeds of the DINO area were once called the Red Wash Formation (Williams 1945), or identified with the Woodside, Thaynes and Ankareh formations (Williams 1945; Untermann and Untermann 1954), but are usually identified as the Moenkopi Formation. The formation has been divided into several members, depending on location, but in the DINO area it is an undivided redbed sequence, becoming coarser going east (Stewart et al. 1972b). Within DINO, the Moenkopi Formation is mostly composed of shale, mudstone, and siltstone, with some sandy or gypsum-rich beds. The rocks are typically red, with some brown, gray, and green intervals, and are not resistant to erosion. The thickness of the unit ranges from 170 to 240 m (550 to 790 ft). Most exposures are found in the Split Mountain and Red Rock anticline areas (Hansen et al. 1983; Gregson et al. 2010).

The depositional environment of the Moenkopi Formation is regarded as continental to nearshore to marine in the DINO area, with freshwater, normal marine, and evaporitic deposition (Rowley et al. 1979b). The basic setting may have been a tidal flat (Hansen et al. 1983). The upper contact with the Chinle Formation is disconformable (McCormick and Picard 1969a). The Moenkopi Formation can date to the Early and early Middle Triassic (Heckert et al. 2005), between approximately 250 and 240 Ma, but is regarded as Early Triassic in age at DINO (Gregson et al. 2010). If the Park City–Moenkopi contact is conformable, then part of the Moenkopi Formation at DINO would have to be Permian in age as well (Schell and Yochelson 1966; Stewart et al. 1972b).

The Moenkopi Formation is very fossiliferous elsewhere, but is not known to be fossiliferous at DINO. There are several reports of fossils in the Uinta Mountains region, though. The earliest reports come from Peabody (1948, 1956), who described reptile tracks (possibly the common form *Chirotherium*) found at Brush Creek north of Vernal. Some of the Peabody tracks include tetrapod swimming traces (Santucci et al. 1998). More recently, a number of tetrapod tracks have been found in the Red Fleet State Park area. These include tracks of amphibians, possible phytosaurs, and undetermined small tetrapods. Swimming traces are more common than terrestrial traces (Hamblin 1992; Hamblin et al. 2000). Finally, marine mollusks

(Rowley et al. 1979b; Gregson et al. 2010), bioturbation, and invertebrate burrows, such as crayfish burrows, are known in the DINO area (Erickson 2007). The most likely fossils to be present in DINO's Moenkopi beds are probably similar terrestrial and shallow subaqueous trace fossils.

San Rafael Group: Entrada Sandstone (Middle Jurassic)

The Entrada Sandstone is dominated by red sandstone, with some silt (Harshbarger et al. 1957). In the DINO area, it is a light-colored cross-bedded sandstone unit similar to the Nugget Sandstone. It is 12 to 50 m (40 to 165 ft) thick, thinning west to east across the monument (Hansen et al. 1983; Gregson et al. 2010). While not as prominent as the Nugget Sandstone, it does form extensive cliffs in the Island Park area. The upper contact with the Stump Formation is disconformable (Gregson et al. 2010), and is the local expression of the regional J-3 unconformity (Pipiringos and O'Sullivan 1978).

The Entrada Sandstone lacks age-diagnostic fossils and is regarded as late Middle Jurassic in age (Hoggan 1970). The climate was semi-arid to arid (Ekdale and Picard 1985) and hot, with sparse vegetation. The Entrada dunes made up the last major erg of the area, and formed when southern Utah was about 22° N (Loope 2008). In the DINO area, the Entrada Sandstone is interpreted as a wet eolian system near a shore to the west (Crabaugh and Kocurek 1993). Occasional marine incursions reworked areas of sand (Gregson et al. 2010).

The Entrada Sandstone is not known to be fossiliferous at DINO, and it is very sparsely fossiliferous overall. Many reports of fossils from the Entrada Sandstone are from the Moab Member, which is now regarded as part of the Curtis Formation (Curtis Member of the Stump Formation in the DINO area) (Doelling 2010). These include the well-known megatracksite of the Moab area (Lockley 1991; Lockley et al. 2007a) and the crocodylomorph *Entradasuchus* (Hunt and Lockley 1995). Hoggan (1970) reported bivalves and crinoids in marine facies of the Entrada Sandstone in the Flaming Gorge area. Trace fossils have been reported from the Escalante Member in south-central Utah, including rare root traces, insect burrows (Loope 2008), vertebrate burrows, traces of sand-swimming vertebrates (Graverson et al. 2007), and theropod (large *Megalosauripus* and smaller *Therangospodus*) and sauropod (cf. *Brontopodus*) dinosaur tracks (Foster et al. 2000). It is possible that deformation structures found in the Entrada Sandstone in the DINO area represent trampling (Loope and Simpson 1992), but this is far from certain (Kocurek and Crabaugh 1993; Loope and Simpson 1993).

Dakota Sandstone (Lower Cretaceous)

The Dakota Sandstone (also Formation or Group) is a heterogeneous rock unit with outcrops as distant as Minnesota. Because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in areas on the western side of the seaway may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). To further complicate matters, there are apparently two different intervals called the Dakota Sandstone in Utah: in northeastern Utah, on the north side of the Uinta Basin (an area including DINO), the Dakota Sandstone is an upper Lower Cretaceous unit that is overlain by the Mowry Shale, which then is separated from the Frontier Formation by a hiatus representing several million years of the early Late Cretaceous. On the south side of the basin, the Dakota Sandstone is above a similar unconformity and dates to the early Late Cretaceous (Molenaar 1991; Franczyk et al. 1992).

The Dakota Sandstone of DINO includes gray, white, and yellow sandstone and lesser amounts of conglomerate and dark carbonaceous shale, typically forming an extensive hogback. It ranges from 12 to 30 m (40 to 100 ft) thick, usually around 15 m (50 ft) (Gregson et al. 2010). A conglomerate is found at the contact with the Mowry Shale (Kirkland et al. 2011a). Untermann and Untermann (1949a) split off the lower part as the Cloverly Formation, with an overlying Fuson Shale for 11 m (36 ft) of red shale and underlying Lakota Sandstone for 6 m (18 ft) of yellow sandstone. Untermann and Untermann (1954) refrained from applying those names, but noted their probable correlations.

The Dakota Sandstone of the DINO area represents shifting deposition, from fluvial to shoreline settings, as the Western Interior Seaway encroached (Gregson et al. 2010). This is generally true for the entire Dakota system: the name is usually applied to rocks deposited during the initial encroachment of the seaway at a given location. The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but usually it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma (see Titus et al. [2005] for discussion of the age). The middle carbonaceous shale at DINO has a date of 101.4 ± 0.4 Ma (Kirkland et al. 2011a), and the top of the formation in the area predates 98.52 ± 0.41 Ma (Oboh-Ikuenobe et al. 2007). The northern Uinta basin version of the formation may be related to the Skull Creek marine cycle (late Albian stage) (Kirkland et al. 2011a), the earliest cycle of the Western Interior Seaway (Cole 1987).

The Dakota Sandstone of DINO is not known to be fossiliferous, but fossils are known from the vicinity. Bird tracks have been found near the top of the formation just outside of the monument, near the northwest corner at the north end of Pipeline Road (Anfinson et al. 2009). They were initially thought to be a new species of *Aquatilavipes* (Anfinson et al. 2004), but are now known to be examples of *Koreanaornis* (Anfinson et al. 2009). The depositional setting at the time the bird tracks were made is interpreted as marginal marine (Anfinson et al. 2009). Petrified wood and twig casts have been found in the Lily Park area east of DINO (Untermann and Untermann 1954). The Utah Field House of Natural History in Vernal has dinosaur bones collected from the basal Dakota

Sandstone just west of the monument (Santucci and Kirkland 2010). Marine invertebrate trace fossils are known at Red Fleet State Park (Hamblin et al. 2000). Coal (Bilbey and Hamblin 1992) and white shell fragments are also common Dakota Sandstone fossils in the area (Gregson et al. 2010).

Over its entire depositional area, and keeping in mind the stratigraphic questions, the fossils of the Dakota Sandstone are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. The best-known terrestrial fossils are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Sandstone (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Sandstone of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Sandstone for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). The floral assemblage in Utah includes coal (Molenaar and Cobban 1991), palynomorphs (including moss spores, fern spores, and gymnosperm and angiosperm pollen; Nichols 1997), the tree fern *Tempskya* (Ash and Read 1976), *Equisetum*, possible gymnosperms, and angiosperms (Tidwell et al. 1976). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Sandstone rocks in Utah reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals; locations included areas near Bryce Canyon National Park (BRCA; see the BRCA summary in this report for more details) (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Sandstone marine fossils include algae, foraminifera (Tibert et al. 2003), dinoflagellate cysts (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Sandstone may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera, corals, bryozoans, brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Bishop Conglomerate (Lower Oligocene)

The Bishop Conglomerate is a middle Cenozoic conglomerate limited to the flanks of the eastern half of the Uinta Mountains (Hansen 1986b). In the DINO area, it is composed of poorly sorted and loosely cemented conglomerate and coarse sandstone. Most of its rocks are derived from the weathering of the Uinta Mountain Group and Paleozoic limestone. It can be 60 to 150 m (200 to 500 ft) thick (Hansen et al. 1983; Gregson et al. 2010). Exposures of the Bishop Conglomerate are scattered throughout DINO, with the largest areas between Island Park and Whirlpool Canyon (Hansen et al. 1983). The Bishop Conglomerate represents a bajada (alluvial fan complex) that formed around the eastern Uinta Mountains. It was deposited under a drying climate on the large Gilbert Peak erosion surface (Hansen 1986b). Ash beds from the formation have been dated to 34.07 ± 0.04 Ma and 30.54 ± 0.22 Ma. The younger ash bed near the top is probably related to the Cottonwood Wash Tuff, erupted from a caldera complex on the Nevada–Utah border (Kowallis et al. 2005). The upper contact with the Browns Park Formation is disconformable, with a hiatus of several million years driven by the tectonic activity (Hansen 1984, 1986b).

Several erroneous interpretations complicated the picture of how the Bishop Conglomerate and overlying Browns Park Formation were deposited. Originally, it was thought that the Bishop Conglomerate was younger than the Browns Park Formation (Sears 1924). In part, this was because the Browns Park Formation was deposited after faulting that postdated the Bishop Conglomerate, resulting in areas where the Browns Park Formation would be found in basins next to high areas capped by the Bishop Conglomerate (Hansen 1986b). Sears (1924) correctly recognized that the Bishop Conglomerate was older, but thought that it was a basal conglomerate of the Browns Park Formation. The Browns Park Formation does have a basal conglomerate in some areas, but it is distinct from the Bishop Conglomerate (Carey 1955; Hansen 1984, 1986b).

The Bishop Conglomerate is not known to be fossiliferous (Sears 1924). With a conglomeratic unit such as this, the most likely fossils are reworked from older units. The Paleozoic limestone clasts are good candidates. *In situ* fossils would probably be limited to extremely durable materials like petrified wood and large mammal bones, particularly teeth.

Browns Park Formation (Miocene)

The Browns Park Formation is an eolian, fluvial, and lacustrine unit deposited in the Browns Park trough, a structural feature that formed on the northeastern flank of the Uinta Mountains during the Miocene (Hansen 1986b; Gregson et al. 2010). At

DINO, it is up to about 90 m (300 ft) thick (Sumison 1976), but reaches thicknesses of 490 m (1,600 ft) elsewhere (Gregson et al. 2010). It is light-colored and heterogeneous, mostly composed of sandstone and ash beds, but with chert, mudstone, siltstone, conglomerate, and limestone (Gregson et al. 2010). The only area where it can be seen in DINO is north of the Gates of Lodore (Hansen et al. 1983; Gregson et al. 2010). A basal conglomerate distinct from the Bishop Conglomerate is present in some areas (Hansen 1984, 1986b), including DINO (Sumison 1976).

Generally, rocks of the eastern Browns Park Formation are eolian in origin, while those in the west are more fluvial (Hansen 1986b). Dunes probably lined the course of the west-flowing ancestral Yampa River (Honey and Izett 1988). Much of the formation is made up of ash tuffs, deposited in both eolian and lacustrine settings. Occasionally, the basin was flooded by lakes (Hansen 1986b). There are many potential sources for the tuffs. For the older tuffs (30 to 10 Ma), most sources are in Nevada, while the tuffs younger than 10 Ma derived from more widely distributed volcanic centers to the west (Luft 1985). The Browns Park Formation was deposited over a long period, with dates ranging from 26 to 8.25 Ma (Lockley et al. 2007b). Deposition was not continuous. There are internal unconformities, and the structural activity that began before deposition continued through deposition, if slower (Hansen 1986b). As a result, relative height of beds is not reliable for correlation or relative dating (Luft 1985). It is likely that most of the formation was deposited after 15 to 12 Ma, in Browns Valley at least (Hansen 1986b). It does not correlate to the Uinta Formation, as originally thought (Carey 1955).

The Browns Park Formation is not known to be fossiliferous at DINO, but there are fossiliferous locations nearby. The first good fossils from the Browns Park Formation were discovered south of Sunbeam, 28 km (17 miles) east of DINO. They consisted of four skulls and partial skulls of small carnivorans, and a tooth assigned to a small camel (Anonymous 1923; Peterson 1924). More recently, fossils of the proboscidean *Gomphotherium* and the rhinocerotid *Aphelops* were found at Douglas Mountain, just outside of the monument (Honey and Izett 1988).

The fossil assemblage of the formation includes diatoms (a type of algae with cell walls of silica) (Hansen 1986b), root traces (Luft and Thoen 1981; Honey and Izett 1988), ostracodes (Hansen 1986b), insect and spider traces (Lockley et al. 2007b), possible invertebrate burrows (Luft and Thoen 1981), borophagine dogs (Wang et al. 1999), gomphothere proboscideans, horses (Honey and Izett 1988), rhinocerotids (Honey and Izett 1988; Prothero 2005), sheep-like oreodonts, camelids, pronghorns (Honey and Izett 1988), and tracks of jumping rodents and camels (Lockley et al. 2007b). Diatoms and ostracodes are limited to lacustrine rocks (Hansen 1986b). Due to the lengthy deposition of the formation, multiple distinct local faunas of mammals can be recognized. The mammals span parts of four different North America Land Mammal Ages (Arikareean to Clarendonian) (Hansen 1986b; Honey and Izett 1988; Wang et al. 1999; Lockley et al. 2007b). Early reports by Peterson (1924, 1928) included some additional mammal taxa, but several were later found to be indeterminate or, in the case of a possible chalicothere, actually from another formation (Honey and Izett 1988). Some fossils from south-central Wyoming were assigned to the Browns Park Formation (McGrew 1951), but actually pertain to the North Park Formation (Izett 1975).

Park Collections

DINO has substantial collections of fossils, and has retained its own material since paleontological work has resumed. Outside of DINO, the largest collections of fossils from the monument are at the Carnegie Museum of Natural History (CM or CMNH) in Pittsburgh, the National Museum of Natural History (USNM) in Washington, D.C., and the American Museum of Natural History (AMNH) in New York City. The University of Utah (UUVP) in Salt Lake City formerly held a substantial collection from DINO as well, but most of their fossils were sent back to DINO in the early 1980s (Chure and McIntosh 1990). Also, a dinosaur vertebra from DINO is in the collections of Yosemite National Park (YOSE), probably used at one point as an educational specimen (Santucci and Kenworthy 2007). So much material was collected that preparation has taken decades; the CMNH was still finding unexpected fossils like the lower jaw of a *Stegosaurus* into the 1980s (Berman and McIntosh 1986), and a field jacket from the original excavation was opened at the Utah Field House of Natural History in Vernal for the centennial of the field work, revealing a *Stegosaurus* tail vertebra (Finlayson et al. 2010). Schell and Yochelson (1966)'s collection from Split Mountain was put into a U.S. Geological Survey collection. The U.S. Geological Survey has several drawers' worth of Paleozoic marine invertebrates from DINO collected during the 1940s and 1950s.

The early years of collection (1909–1924) produced about 20 specimens that are complete enough to mount for display (Chure and McIntosh 1990). Museum that exhibit DINO specimens include the AMNH, which has a *Barosaurus* obtained via the CMNH, the USNM, and the University of Utah (Norell et al. 1995); the CMNH, which displays DINO fossils of the turtle *Glyptops*, the crocodylomorph *Hoplosuchus*, the theropod *Marshosaurus*, the sauropods *Apatosaurus*, *Camarasaurus*, and *Diplodocus*, *Stegosaurus*, and the ornithopods *Dryosaurus* and *Uteodon* (=*Camptosaurus*) (Chure 2010); the Denver Museum of Nature and Science (DMNS or DMNH) in Denver, which has

a *Diplodocus* obtained via the CMNH (Stokes 1949; Markman 1961); the Royal Ontario Museum (ROM) in Toronto, which in 1962 acquired a partial *Barosaurus* skeleton collected from DINO by the CMNH, and put it on exhibition in 2007 (Anonymous 2007); the USNM, which has a *Diplodocus* (Stokes 1949); the Museum of Paleontology at the University of Michigan (MPUM) in Ann Arbor (Chure and McIntosh 1990); and the University of Nebraska State Museum (UNSM) in Lincoln (Griesemer et al. 1976; Chure and McIntosh 1990). The University of Utah formerly exhibited an *Allosaurus* from DINO, but it was transferred back to DINO with the bulk of their collection from the monument, and the skull is on display there (Chure and McIntosh 1990). A skeleton of *Apatosaurus* was sent to the Natural History Museum of Los Angeles County (LACM), in Los Angeles (McIntosh and Berman 1975), but the museum no longer has any DINO fossils on display (Chure and McIntosh 1990). Another noteworthy specimen returned to DINO is the *Camarasaurus* braincase described in White (1958) (Chure 1981). The AMNH and CMNH also have mountable skeletons in storage (Chure and McIntosh 1990). A table of type specimens from DINO is included below.

Taxon	Publication	Catalog Number	Provenance	Notes
Abydosaurus mcintoshi	Chure et al. 2010	DINO 16488	Cedar Mountain Formation, DNM 16	sauropod
Apatosaurus louisae	Holland 1915	CMNH 3018	Morrison Formation, Dinosaur Quarry	sauropod
Camarasaurus annae	Ellinger 1950	CMNH 8942	Morrison Formation, Dinosaur Quarry	sauropod, now assigned to *Camarasaurus lentus*
Camptosaurus aphanoecetes	Carpenter and Wilson 2008	CMNH 11337	Morrison Formation, Dinosaur Quarry	ornithopod, renamed *Uteodon aphanoecetes* by McDonald (2011)
Dinochelys whitei	Gaffney 1979	DINO 986 to 991	Morrison Formation, Dinosaur Quarry	turtle
Glirodon grandis	Engelmann and Callison 1999	DINO 10822	Morrison Formation, unspecified locality	multituberculate mammal
Glyptops utahensis	Gilmore 1916	CMNH 3412	Morrison Formation, Dinosaur Quarry	turtle, now assigned to *Glyptops plicatulus*
Hoplosuchus kayi	Gilmore 1926	CMNH 11361	Morrison Formation, road excavation near Dinosaur Quarry	crocodylomorph
Iridotriton hechti	Evans et al. 2005	DINO 16453a-b	Morrison Formation, Rainbow Park Microsite 96	salamander
Koparion douglassi	Chure 1994	DINO 3353	Morrison Formation, Rainbow Park Microsite 94	theropod
Rhadinosteus parvus	Henrici 1998a	DINO 14693	Morrison Formation, Rainbow Park Microsite 96	frog
Schilleria utahensis	Evans and Chure 1999	DINO 14270	Morrison Formation, DNM 375	lizard
Triconolestes curvicuspis	Engelmann and Callison 1998	DINO 10780	Morrison Formation, Rainbow Park Microsite 96	triconodont mammal

Taxon	Publication	Catalog Number	Provenance	Notes
Uintasaurus douglassi	Holland 1919	CMNH 11069	Morrison Formation, Dinosaur Quarry	sauropod, now assigned to *Camarasaurus lentus*

DINO Table 2. Type specimens from DINO. The taxon column presents the original name used; because taxonomy is subjective and under constant revision, names will vary from source to source. Significant alternatives are included in the notes column.

Cultural Resource Contexts

As noted before, people have been in the Uinta Mountains region since the late Pleistocene (Bilbey 1992a). DINO and its immediate vicinity have a number of archeological sites, mostly pertaining to the Fremont culture of the late first and early second millennia AD. A number of these sites have yielded fossils used for cultural purposes. Two common fossils collected by the Fremont people include crinoid stems, which were sometimes used for beads, and guards of the Jurassic belemnite *Pachyteuthis*, both of which can be found throughout the monument. Other fossils found at these sites include petrified wood and mollusk shells, which were sometimes fashioned into pendants (Breternitz 1970). In more recent times, the fossils of DINO have become part of the modern culture of Vernal and surrounding areas, and dinosaur-related tourism is an important part of the local economy (pers. obs.). Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory workshop for DINO during August 1998 (NPS 1988) and produced a geologic resource evaluation report for the park in 2006 (Graham 2006). It would be worthwhile to take stock of actions made regarding recommendations in the two documents. Both suggested a comprehensive study of the Cedar Mountain Formation in the monument. NPS (1998) reported the need for paleontological maps, and Graham (2006) added studies of the Moenkopi, Chinle, Nugget, Dakota, Mowry, and Frontier formations to the list of stratigraphic units of interest. Graham (2006) also noted the need for improved storage facilities, especially in regard to ventilation of radon gas, a product of mildly radioactive fossils.

- DINO reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 431 localities, 334 of which were reported to be in good condition in 2010. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- The abundance and publicity of fossils at DINO puts them at increased risk from visitors. Parks such as FOBU and Petrified Forest National Park have investigated casual collecting of fossils and the social science driving the behavior. Contact the Geologic Resources Division for further information or assistance.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Within the NCPN, the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and

increase understanding of this important fossil-bearing unit. A similar USGS–NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Dinosaur National Monument Digital Geologic Map
2006–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=1038936.

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Royal Ontario Museum
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National Museum of Natural History
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University of Colorado–Boulder Dinosaur Tracks Museum collection (formerly at University of Colorado–Denver)
1985–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by museum staff; status: Active.

Museum of Paleontology, University of Michigan
1924–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by MPUM staff; status: Active.

University of Nebraska State Museum
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Utah Field House of Natural History State Park Museum
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Fossil Butte National Monument

Fossil Butte National Monument (FOBU), Wyoming, was established to protect a portion of an extremely fossiliferous Eocene lake deposit, including the geologic landmark known as Fossil Butte (FOBU Figure 1). These rocks, deposited over 50 million years ago in a lake and associated floodplains, are world-famous for their well-preserved fish fossils, and associated plants, insects, mollusks, and rare tetrapods. FOBU was established October 23, 1972. It encompasses 3,318 ha (8,198 acres), all of which are under federal administration (FOBU Figure 2).

Geologic Background

FOBU is located near the Wyoming–Utah border in southern Lincoln County, southwestern Wyoming. It is within the narrow, north-south trending Fossil Basin (known as Fossil Syncline in some older works), bounded on the east by Oyster Ridge and on the west by a series of small ranges including the Crawford Mountains and the Tunp Range. FOBU is about 16 km (10 miles) west of Kemmerer, Wyoming, and about 64 km (40 miles) north-northwest of Evanston, Wyoming. To the east of Oyster Ridge is the Green River Basin, which occupies most of the rest of southwestern Wyoming.

Fossil Basin has been the subjects of numerous paleontological and geological publications, abstracts, theses, and dissertations. A selection of noteworthy documents includes: Peale (1879); Lesquereux (1883); Cope (1884); Scudder (1890); Veatch (1907); Schultz (1914); Knowlton (1923); Brown (1936, 1937); Thorpe (1938); Hesse (1939); Bradley (1948); Gazin (1952, 1956a, 1962); Tracey and Oriel (1959); Schaeffer and Mangus (1965); Stucker (1966); Rubey et al. (1968, 1975); Brodkorb (1970); Oriel and Tracey (1970); McGrew (1975); McGrew and Casilliano (1975); Edwards (1976); Olson (1977); Grande (1980, 1982a, 1982b, 1984, 1989, 1994, 2001); Podorsky (1981); Feldmann et al. (1981); Dayvault and Bersch (1983); Cushman (1983, 1998, 1999); Dexter (1983); Beard (1988); Breithaupt (1990); Grande and Bemis (1991); M'Gonigle and Dover (1992, 2004); Buchheim (1993, 1994a, 1994b, 1998); Grande and Buchheim (1994); Ambrose et al. (1997); Buchheim and Eugster (1998); Leggitt et al. (1998); Loewen and Buchheim (1998); Rieppel and Grande (1998); Biaggi and Buchheim (1999); Zonneveld et al. (2000); Gunnell et al. (2002); Koch and Santucci (2002); Carvalho et al. (2004); von Koenigswald et al. (2005); Hunt et al. (2006); Conrad et al. (2007); Simmons et al. (2008); Smith et al. (2008); Aase (2009); Santucci et al. (2009); Martin et al. (2010); McMurran and Manchester (2010); Weidig (2010); and Buchheim et al. (2011). The Geologic Resources Division of the National Park Service (NPS) coordinated a geologic resources inventory workshop for FOBU during May 2002 (NPS 2002). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The geologic record exposed at FOBU includes an episode of marine and marginal marine deposition during the Early Triassic, floodplain and lake deposition during the Eocene, and recent surficial processes (see the appendix for a geologic time scale). The oldest rocks of the monument are represented by a small area of outcrops south and west of Rubey Point (McGrew and Casilliano 1975; M'Gonigle and Dover 1992, 2004). These Triassic rocks were deposited between about 250 and 245 Ma (million years ago) and include the Woodside Redbeds and overlying Thaynes Limestone. These two formations represent part of a series of marine transgressions (marine advances) and regressions (retreats) during the Early Triassic. The Woodside Redbeds were deposited during a marine regression, while the Thaynes Limestone was deposited during a rapid transgression from the west (Paull and Paull 1987; Paull et al. 1989). The sea that deposited the Thaynes Limestone was the most extensive shallow seaway of the Early Triassic in the United States (Boyer et al. 2004). The seaway underwent smaller-scale transgressions and regressions. FOBU was in an area that was usually situated in the outer shelf to inner shelf belts, with a deep basin to the west (Pisera et al. 1996). Paleolatitude for the area was only a few degrees north of the equator. The sea was warm and low-energy, and the coast was open (James 1980).

A number of rock units were deposited in the vicinity of FOBU after the Thaynes Limestone but before the Wasatch Formation, but none of them are exposed within FOBU (McGrew and Casilliano 1975); nearly 200 million years of geologic history between the two formations are not represented at the surface within FOBU. Toward the end of this gap, during the Late Cretaceous, a major mountain-building event known as the Laramide Orogeny began. Like most geologic structures in southwestern Wyoming, Fossil Basin was formed as a result of compressional stresses during the Laramide Orogeny. It is within the Overthrust Belt of western Wyoming (Roehler et al. 1977). Fossil Basin began subsiding ("sinking") during the latest Cretaceous; the subsidence persisted into the Eocene (Oriel and Tracey 1970). By the Eocene, the mountain ranges and associated basins of southwestern Wyoming were in place (Bradley 1964). The present landscape is generally an exhumed Eocene landscape (Oriel and Tracey 1970). Three thrust faults (from east to west, the Absaroka, Tunp, and Crawford faults) were active during the Eocene in Fossil Basin, shedding debris flows and providing sediment for part of the Wasatch Formation (Hurst and Steidtmann 1986). Paleozoic and Mesozoic rocks were thrust east over younger Mesozoic rocks (Rubey et al. 1975). With the appearance of Fossil Basin and the surrounding uplifts, the geographic stage was set for the creation of Fossil

Lake, and the deposition of the interrelated Wasatch and Green River formations.

Deposition in Fossil Basin began as it was forming. The Late Cretaceous-age Adaville Formation and Late Cretaceous–Paleocene-age Evanston Formation record the early history of the basin (Oriel and Tracey 1970). Above them, the Wasatch and Green River formations of Fossil Basin date to the early part of the Eocene, with deposition between about 54 and 50 Ma. These formations represent two parts of a system in the basin: the fluvial and floodplain environments of the Wasatch Formation, adjacent to the lacustrine environment of Fossil Lake preserved as the Green River Formation. This is a simplification, because Wasatch Formation deposition began before the appearance of Fossil Lake and continued after it, but the paleoenvironmental relationship can be recognized for the entire history of Fossil Lake. The Wasatch-Green River fluvial-lacustrine system is present in two other nearby basins (Smith et al. 2008). To the east, on the opposite side of Oyster Ridge (Oriel and Tracey 1970), was Lake Gosiute, which occupied portions of the greater Green River Basin. To the south, on the other side of the Uinta Mountains, was Lake Uinta (Oriel and Tracey 1970; Smith et al. 2008). Fossil Lake was the smallest and shortest-lived of these three lakes (Breithaupt 1990; Smith et al. 2008). Lake Gosiute and Lake Uinta were shallow, playa-type lacustrine systems that frequently became partly desiccated or saline (Grande 1989; Smith et al. 2008). Fossil Lake also became periodically saline, but even during these times its margins are thought to have remained relatively fresh (Buchheim 1994a). The persistence of fresher water may be responsible for Fossil Lake having a more diverse fish fauna than the other two lakes despite its much smaller size (Grande 1989). Altogether, the three lakes existed on the order of 10 million years during the late Paleocene and Eocene (roughly 54 to 43 Ma; Smith et al. 2008) and occupied 87,000 km^2 (34,000 $miles^2$) (Grande 1989). They are renowned for their fossils: the Green River Formation has several biotas that qualify as lagerstätten, due to their diversity and preservation (Grande 1989). One of the most common fish species from the Fossil Lake sediments is *Knightia eocaena*, the state fossil of Wyoming (Hunt et al. 2006).

Fossil Lake dates to the early part of Green River Formation deposition (Smith et al. 2008). The lake was approximately 1,500 km^2 (580 $miles^2$) at its greatest extent, and produced over 120 m (395 ft) of lake sediments (Buchheim and Eugster 1998). Overall, with fluvial and deltaic incursions of the Wasatch Formation included, the Fossil Lake section is about 200 m (650 ft) thick (Buchheim et al. 2011). The lake underwent several transgression-regression cycles, with rapid transgressions followed by slow regressions (Buchheim 1994a; Buchheim and Eugster 1998). Drainage entered the lake primarily from the south and southwest, forming deltas (Buchheim and Eugster 1998), with one large inflow from the northeast (Buchheim 1998). The surrounding Wasatch Formation was a vegetated floodplain with rivers and streams (Buchheim 1998). Uplifted areas on the Crawford and Tunp faults formed islands (Buchheim et al. 2011). Forested mountains rose in the vicinity (Cushman 1999), attaining heights of 1,200 to 1,500 m (4,000 to 5,000 ft) above sea level (Roehler 1993). The basin continued to be tectonically active during deposition of the Green River Formation, with uplift reactivated around the onset of Fossil Lake (Leggitt et al. 1998).

The general history of Fossil Lake is that of a small floodplain lake that became a large lake, and eventually became hypersaline before disappearing (Buchheim and Eugster 1998). The first two-thirds of its history were mostly freshwater, and it became hypersaline in the final third; it is thought that the water was alkaline throughout (Buchheim 1994a). Three major phases of the lake's history can be identified. The early history of Fossil Lake featured a shallow, usually freshwater lake, probably never more than 5 m (16 ft) deep (Buchheim 1998). This phase of the lake was concentrated in the southern part of Fossil Basin, which initially was divided into north and south segments by a structural barrier (Biaggi and Buchheim 1999). This lake became hypersaline at least twice (Buchheim 1998). The end of the initial phase of Fossil Lake is marked by a rock unit known as the sandstone tongue of the Wasatch Formation, representing a major advance of a delta from the south (Biaggi and Buchheim 1999). Once the southern segment of the basin filled in, the succeeding phase of Fossil Lake was able to expand over a much wider area (Biaggi and Buchheim 1999).

The middle phase of Fossil Lake featured greater precipitation, feeding the much larger lake. At this time, the large freshwater lake was populated by abundant fish. Occasional arid periods are recorded in its minerals (Buchheim 1998). The end of the middle phase is defined by a transition from laminated micrites (microcrystalline calcite) to more massive (lacking structures like laminations) carbonate rocks (made up of carbonate minerals such as calcite and dolomite) with interbeds of evaporites and siliciclastic rocks (made up of clasts [fragments] of rocks made of silicate minerals) (Buchheim et al. 2011). The transition is a few meters above the thickest volcanic tuff of the Fossil Lake rocks, the k-spar tuff (Buchheim et al. 2011). This bed is dated to 51.66 ± 0.09 Ma (Smith et al. 2008). More than 130 ash and tuff beds are present in the rocks of Fossil Lake; at least 21 of them are thick and potentially datable (A. Aase, FOBU paleontologist, pers. comm., October 2011). Volcanic material may have come from the Absaroka, Challis, and Lowland Creek volcanic fields to the north and northwest (Smith et al. 2008). The final phase of Fossil Lake was mostly hypersaline, with a fresher period at the end just before being completely filled by the encroaching rivers of the Wasatch Formation (Buchheim 1998). The basin became closed to outside drainage during the final phase (Loewen and Buchheim 1998). Fish are only found in the margins of the lake, where a freshwater apron seems to have persisted (Buchheim 1998). Stromatolites and other microbial sedimentary features (microbialites) increased; microbial debris was probably the main component of the formation's kerogen-rich laminated

micrite (informally known as oil shale) (Loewen and Buchheim 1998). After Fossil Lake was filled in by sediment, small lakes occasionally appeared in the basin, represented by limestone beds within the overlying Wasatch Formation (Buchheim 1998).

Fossil Basin during the time of Fossil Lake is interpreted as having had a transitional climate, fluctuating between humid subtropical and drier warm temperate conditions (Cushman 1999). Climate began as warm temperate. Humidity increased and the temperature warmed to subtropical, then cooled toward the end, possibly warming again just before the end (Buchheim 1998). Aridity also increased during the late part of its history (Loewen and Buchheim 1998; Wilf 2000). Oriel and Tracey (1970) proposed a humid subtropical to tropical climate, with a mean annual temperature around 18 °C (65°F) and mean annual precipitation of about 100 cm (40 in). MacGinitie (1969), writing primarily about Lake Uinta, suggested something similar but more detailed: a subhumid subtropical climate with a mean annual temperature of 20 °C (67 °F) and precipitation of 61 to 89 cm (24 to 35 in) at lake level, with relatively dry winters, rainy late springs and early summers, and dry and hot late summers and autumns. Different authors have compared the Green River Formation climate to the Gulf Coast or much drier southeastern California (Eugster 1982), or found it unlike anything present today (MacGinitie 1969). Evaporation was not as significant at Fossil Lake as at the other lakes, and the elevation of the site was probably 330 m (1,000 ft) or less (Oriel and Tracey 1970). The neighboring Green River Basin was probably frost-free during Fossil Lake time (Wilf 2000). Southwestern Wyoming was about 35° N, which is about 5° to 8° south of the present latitude. Winds blew from the west and northwest (Roehler 1993). The Paleocene–Eocene Thermal Maximum, a brief global temperature spike, occurred more or less at the onset of Wasatch Formation deposition in Fossil Basin (Wing et al. 2003), and the Cenozoic Global Climate Optimum, the warmest interval of the Cenozoic, was beginning just as Fossil Lake was coming to an end (Clyde et al. 2001).

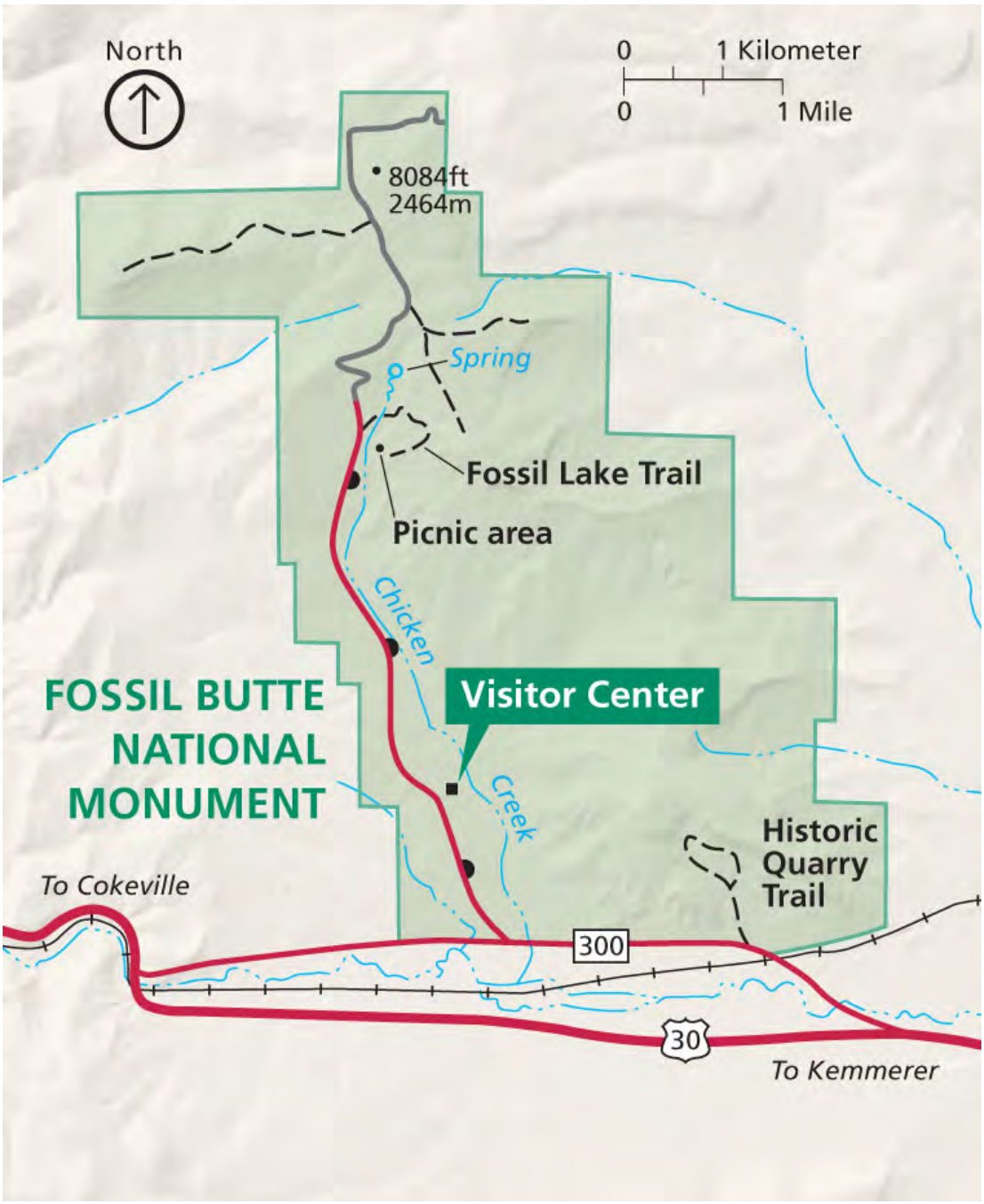
Climate forcing by orbital parameters were also in effect: Milankovitch cycles have been identified in roughly contemporaneous rocks from Lake Gosiute (Roehler 1993; Clyde et al. 2001).

There are no rock exposures at FOBU that represent the period after the end of the Green River–Wasatch system in Fossil Basin (McGrew and Casilliano 1975; M'Gonigle and Dover 1992, 2004). It is known that multiple episodes of uplift and faulting occurred following the main period of Laramide uplift in the region (Bradley 1964), and seismic activity has persisted to the present, with Holocene fault movement on the Rock Creek Fault, paralleling the Tunp Thrust west of FOBU (McCalpin 1993). Humans were present in the Bonneville Basin to the east perhaps before 12,300 years ago, and had reached the east side of Lake Bonneville over 11,000 years ago (Rhode et al. 2005).

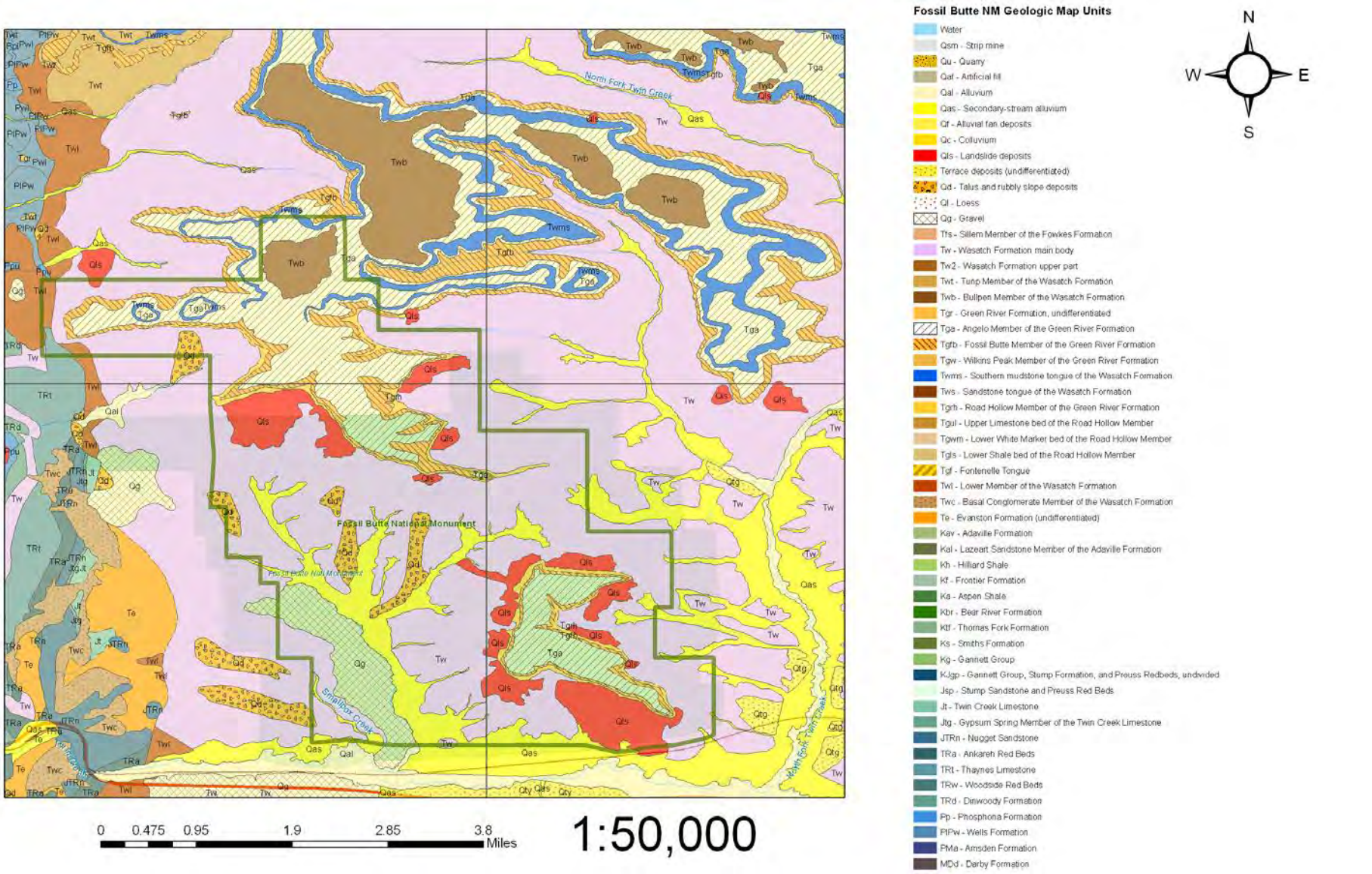
Rock units exposed within FOBU include, from oldest to youngest: the Woodside Red Beds (Lower Triassic); the Thaynes Limestone (Lower Triassic); and the intertongued Wasatch and Green River formations (Lower Eocene). The intertongued units include, in stratigraphic order, the lower member and main body of the Wasatch Formation, the Road Hollow and Fossil Butte members of the Green River Formation, the mudstone tongue of the Wasatch Formation, the Angelo Member of the Green River Formation, and the Bullpen Member of the Wasatch Formation. Surficial units include a variety of unconsolidated Quaternary deposits (see FOBU Figure 2 for a geological map) (Pleistocene–Holocene) (McGrew and Casilliano 1975; M'Gonigle and Dover 1992, 2004; Road Hollow Member after Buchheim et al. 2011). The main body of the Wasatch Formation, the Green River Formation, and the Quaternary sediments are known to be fossiliferous at FOBU; of the others, most are usually not extensively fossiliferous and/or are exposed over small areas, but are potentially fossiliferous (FOBU Table 1).



FOBU Figure 1. Fossil Butte (VLS).



FOBU Figure 2. Geography of FOBU.



FOBU Figure 3. Schematic geological map of FOBU. Member-level divisions of the Green River Formation have been revised by Buchheim et al. (2011). Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1044844.

Formation	Age	Fossils Within FOBU	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Bison bones	Fluvial and landslide deposits
Green River Formation	Early Eocene	Stromatolites and other microbialites, plant fossils, freshwater invertebrates, insects, rays, bony fish, rare tetrapods, and abundant vertebrate coprolites	Freshwater to evaporative lacustrine settings (Fossil Lake)
Wasatch Formation	Early Eocene	Invertebrate trace fossils and locally abundant fragmentary fossils of turtles, lizards, crocodilians, and mammals	Fluvial and floodplain system, associated with Fossil Lake for much of the formation's history
Thaynes Limestone	Early Triassic	None to date; marine invertebrates are most likely, and brachiopods, bivalves, ammonites, gastropods, and crinoids have been found in Fossil Basin	Outer shelf to inner shelf marine settings
Woodside Redbeds	Early Triassic	None to date; fossils are very rare in this formation in general	Terrestrial coastal settings

FOBU Table 1. Concise tabulation of FOBU stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of FOBU present opportunities for education, interpretation, and continued or future scientific research in the monument.

The Hayden Survey worked in the vicinity as early as 1868, and reached Fossil Basin by 1872 (Schultz 1914). Fossil Butte first appears in the literature in 1879, when it was described by A. C. Peale for the Hayden survey (1879). Fossils collected by Peale were described by Edward Drinker Cope (1877, 1878, 1884) (vertebrates), Leo Lesquereux (1883) (plants), and Samuel H. Scudder (1890) (insects).

Cope probably had quarries in the FOBU area by the late 1870s (Dayvault and Bersch 1983). His locality information leaves something to be desired, but it is possible that his "main line of the Wasatch Mountains" locality (Cope 1877) represents the Fossil Butte area (Hesse 1939). The Twin Creek locality (Cope 1884) is definitely from the Fossil Butte area (Hesse 1939; Grande 1980), and part of the Twin Creek Valley is within FOBU. The butte itself is sometimes known as "Fish Cliffs" in old references (e.g. Hesse 1939, Stucker 1966). Parties from Princeton University began collecting Wasatch Formation mammals from Fossil Butte and the vicinity in 1939. The Smithsonian Institution collected other mammal fossils from west-facing exposures on the saddle north of the butte, and on southern exposures of the butte itself (Gazin 1962).

The Fossil Butte area was long preserved as public lands (Hesse 1939), which helped establish widespread commercial fossil collecting. Commercial collecting has gone on in the Fossil Basin's Green River Formation since the late 19th century (Grande and Buchheim 1994), and has greatly expanded in the decades since the 1960s. It is possible that a half million fish skeletons were removed between 1965 and 1990 (Grande 2001). Perhaps half of the accessible outcrops of the most famous layer, the 18-inch layer (see below), is being worked by commercial collectors (Buchheim 1998). A handful of collectors (Robert Lee Craig, the Haddenham family, Rick Hebdon, the Tynsky family, and the Ulrich family) are responsible for many of the fossils now in institutions and private hands (Grande 1980; A. Aase, pers. comm., October 2011). FOBU cooperates with local commercial quarries to raise awareness of the fossils of the basin (Hunt et al. 2006), and worked with them on a paleobotanical project begun in 2004 (Aase 2009). Some former commercial sites are within FOBU, such as the Haddenhams site (Olson 1977) and the Larson Fish Quarry (Stucker 1966).

Wasatch Formation (Lower Eocene)

The Wasatch Formation of Fossil Basin is a variegated, heterogeneous fluvial unit (M'Gonigle and Dover 1992, 2004). As with the neighboring Green River Basin, the Wasatch and Green River formations of Fossil Basin have a complex intertonguing relationship. At the bottom of the sequence is a basal conglomerate assigned to the Wasatch Formation. This is followed by an unnamed lower member of the Wasatch Formation, and then the formation's main body. The Green River Formation, representing Fossil Lake, appears in the basin during the deposition of the main body, and the two units intertongue laterally in response to changes in the size of the lake (Oriel and Tracey 1970; McGrew and Casilliano 1975). One particularly extensive sandstone tongue interposes on the lower Green River Formation from the south (Oriel and Tracey

1970; McGrew and Casilliano 1975), separating the Road Hollow Member of the Green River Formation from the overlying Fossil Butte Member (Biaggi and Buchheim 1999; Buchheim et al. 2011). Higher in the section, an extensive mudstone tongue interposes from the north into the Angelo Member of the Green River Formation. After Fossil Lake ceased to exist, the Bullpen Member of the Wasatch Formation was deposited across the entire basin. An additional member of the Wasatch Formation, the Tunp Member, was deposited during the entire Wasatch–Green River sequence along the periphery of the basin, associated with the thrust faults (Oriel and Tracey 1970; McGrew and Casilliano 1975). Four of the Wasatch Formation members can be observed at FOBU. From oldest to youngest, they are the lower member, the main body, the mudstone tongue, and the Bullpen Member (McGrew and Casilliano 1975; M'Gonigle and Dover 1992, 2004).

The lower member is only found near Rubey Point, as part of a faulted section that also includes the monument's Triassic rocks (McGrew and Casilliano 1975). It is up to 100 m (330 ft) thick in the vicinity of FOBU, and is composed of a combination of gray, brown, and red mudstone, sandstone, claystone, and limestone (M'Gonigle and Dover 1992, 2004). Depending on the location, it may grade into the main body, or have an angular unconformity (a contact including both a hiatus in deposition and structural deformation, so that beds of the lower unit are not parallel to those in the overlying unit) as the contact (Oriel and Tracey 1970).

The main body makes up much of the bedrock at FOBU. It is at least 450 m (1,475 ft) thick in the FOBU area. The unit is variegated and heterogeneous, including mudstone, sandstone, siltstone, claystone, conglomerate, and marlstone (Rubey et al. 1975). Lenticular bodies of limestone appear near the Green River Formation. Rocks near the center of the basin (the old Knight Formation; see below) are finer-grained, while those on the periphery (the old Almy Formation) are coarser (Oriel and Tracey 1970).

The mudstone tongue is only present in northern FOBU (McGrew and Casilliano 1975). Historically, it has been used to mark the contact between the Fossil Butte and Angelo members of the Green River Formation, but this contact has been revised to be lower than the mudstone tongue (Buchheim et al. 2011). This unit is about 20 m (65 ft) thick at most, and pinches out going east and south. It is composed of red, green, and brown mudstone and shale (M'Gonigle and Dover 1992, 2004). The color becomes more green and gray moving into the basin (Oriel and Tracey 1970). This tongue appears to have come from the Tunp Member (M'Gonigle and Dover 1992, 2004) and represents a regressive phase of Fossil Lake (Buchheim and Eugster 1998).

The Bullpen Member caps the topography of northern FOBU (McGrew and Casilliano 1975). This unit is similar to the main body (Rubey et al. 1975). It is up to about 120 m (395 ft) thick, and is composed of variegated mudstone, sandstone, conglomerate, limestone, and shale (Rubey et al. 1975). It has a gradational contact with the underlying upper Green River Formation (Oriel and Tracey 1970).

In the early scientific history of Fossil Basin, the Wasatch Formation was described as a group, with up to three formations: the basal Almy Formation, the middle Fowkes Formation, and the upper Knight Formation (Veatch 1907). It was soon recognized that the Fowkes Formation was misplaced (Schultz 1914). This unit is now known to be above the Wasatch Formation (Tracey and Oriel 1959; Oriel and Tracey 1970; McGrew and Casilliano 1975). Later, it was shown that the Almy Formation was the coarse peripheral facies of the Knight Formation, and the two names were abandoned (Tracey and Oriel 1959; Gazin 1962; Oriel and Tracey 1970). Although no longer in use, the names can be found in documents published between 1907 and the early 1960s, including two important publications on Wasatch Formation paleontology by Charles Gazin (Gazin 1952, 1956b).

The duration of the Wasatch Formation in Fossil Basin is shorter than in the Uinta and greater Green River basins. Smith et al. (2008) depicted the main body as beginning around 53.5 Ma and the Bullpen Member as ending before 50 Ma, which is entirely within the Early Eocene. Under the North American Land Mammal Age system, this range would include part of the Wasatchian NALMA (late Graybullian, Lysitean, and Lostcabinian subages, ending approximately 51 Ma) and part of the Bridgerian NALMA (early Gardnerbuttean) (Smith et al. 2008). The exact duration of the formation in Fossil Basin is not certain, because of the paucity of chronostratigraphically-useful fossils from the lowest and uppermost beds (Oriel and Tracey 1970), but the main body is known to have Lysitean and Lostcabinian fossils (Ambrose et al. 1997; Gunnell et al. 2002). The basal conglomerate may include rocks as old as the Late Paleocene (Hurst and Steidtmann 1986). The main body of the Wasatch Formation in Fossil Basin overlaps the main body in the greater Green River Basin in age, and the upper part of the formation is contemporaneous in part with the New Fork and Cathedral Bluffs tongues in the greater Green River Basin (Smith et al. 2008). The depositional settings of the Wasatch Formation included streams, rivers, and alluvial plains, with occasional mudflows and rock slides. Climatically, it has been compared to a savanna (Oriel and Tracey 1970).

The great majority of fossils in the Wasatch Formation of Fossil Basin come from the main body, and it is the only unit of the formation that is known to be fossiliferous at FOBU. The other units are much less fossiliferous overall. The lower member includes some microbial ("algal") limestone (Oriel and Tracey 1970; M'Gonigle and Dover 1992, 2004), charophytes (green algae), pollen, spores, gastropods, ostracodes (seed shrimp), gars, and the small herbivorous mammal *Meniscotherium* (a phenacodontid condylarth, one of many extinct lineages present in the Wasatch Formation) (Oriel and Tracey 1970). The

mudstone tongue includes fossils of logs and branches that fell into water and were encrusted by layers of calcium carbonate-producing cyanobacteria ("blue-green algae") (Oriel and Tracey 1970; McGrew and Casilliano 1975); these fossils are sometimes known as algal logs, algal-encrusted logs, and similar terms, although strictly speaking the encrusting organisms are no longer described as algae. Gastropods, gars, and the extinct fish *Asineops* are known from the Bullpen Member, with the fish found in limestone beds (Oriel and Tracey 1970). The Powerline *Presbyornis* nesting site, previously assigned to the Angelo Member of the Green River Formation (Leggitt et al. 1998), is now thought to pertain to a ponding event in overlying Wasatch Formation beds (A. Aase, pers. comm., October 2011), which may be in the Bullpen Member.

The Wasatch Formation within FOBU has locally abundant fragmentary terrestrial fossils, which are collected by monument staff (Santucci et al. 2009). The best-known Wasatch Formation fossils in the monument come from outcrops of the main body on Fossil Butte. The mammals have been discussed in several publications and abstracts (Gazin 1952; Tracey and Oriel 1959; Gazin 1962; Oriel and Tracey 1970; McGrew and Casilliano 1975; Beard 1988; Breithaupt 1990; Ambrose et al. 1997; Gunnell et al. 2002). Gazin (1952, 1962) and Breithaupt (1990) published faunal lists, which are now partially superseded by work done during the late 1990s and early 2000s by Albion College and the University of Michigan. Breithaupt's list, the most recent, included the leptictid *Diacodon*, a small semi-bipedal insectivore; the semi-aquatic and hippo-like pantodont *Coryphodon*; the lemur-like primate *Pelycodus* (specimen reassigned to *Copelemur australotutus*; Beard [1988]); the rodent *Reithroparamys*; the hyaenodontid creodont *Sinopa*, an early mammalian carnivore; two early marten- or civet-like carnivorans, the viverravid *Didymictis* and the miacid *Vassacyon*; the hyopsodontid condylarth *Hyopsodus*, an elongate small ungulate; a possible occurrence of the phenacodontid condylarth *Meniscotherium*; the horse *Hyracotherium*; the helatoid (tapir relative) *Heptodon*; and the early artiodactyl (even-toed ungulate) *Hexacodus* (Breithaupt 1990). Keeping in mind changing taxonomy and additional material, this list is close to those provided by Gazin. The only differences between the lists above the family level are reports of the stylinodontid taeniodont *Ectoganus* (a burrowing mammal) and a possible occurrence of the mesonychian *Pachyaena* (another early carnivore) in Gazin (1962). At least two species have been named from material found at Fossil Butte. The primate *Coepelemur australotutus* was named from a mandible fragment (USNM [National Museum of Natural History, Washington, D.C.] 22261) collected in 1953 on the saddle of Fossil Butte (Beard 1988). The artiodactyl *Hexacodus uintensis* was named from a specimen (PU [Princeton University, Princeton, NJ] 16175) collected about 5 km (3 miles) northwest of Fossil (Gazin 1952), which probably puts it within FOBU.

The recent work confirms Gazin's (1962) observation that the vertebrate assemblage changes going stratigraphically up the butte, and has shown that Wasatch Formation fossils are present in distinct intervals at Fossil Butte. Ambrose et al. (1997) found the lowest 10 m (33 ft) of the section and the uppermost 30 m (100 ft) to contain fossil vertebrates. The lower interval, composed of silty and sandy beds, is interpreted as lake margin and fluvial settings with some paleosol (fossil soil) development. The upper interval is composed of fluvial mudstones with greater paleosol development (Ambrose et al. 1997). Gunnell et al. (2002) distinguished a third vertebrate-producing interval just below the Green River Formation, interpreted as channel and crevasse splay sandstones interbedded with floodplain mudstones. Eight or nine reptile species are present in the intervals: three glyptosaurine lizards (armored lizards), a dermatemydid turtle (river turtle), an emydid turtle (pond or marsh turtle), two trionychid turtles (softshelled turtles), an alligatorid, and a crocodylid (Ambrose et al. 1997; Gunnell et al. 2002; the former paper reported four turtle species and one crocodilian, while the latter reported three turtles and two crocodilians). The mammal fauna includes nearly three dozen species, pertaining to proteutherians (tree shrews and relatives), palaeanodonts (scaly anteater-like), taeniodonts, rodents, primates, pantodonts, creodonts, carnivorans, mesonychians, condylarths, perissodactyls (odd-toed ungulates), and artiodactyls (Ambrose et al. 1997; Gunnell et al. 2002). Most of the fossil taxa pertain to the Lysitean subage, but at least one, the perissodactyl *Lambdotherium* from near the top of the section, is known from the succeeding Lostcabinian, giving a Lysitean-Lostcabinian age for the Wasatch Formation at Fossil Butte (Ambrose et al. 1997). Invertebrate trace fossils from multiple depositional settings are also found in the main body at FOBU. There are three main assemblages: a flood basin and alluvial plain assemblage with *Planolites* and *Skolithos* burrows and possible *Celliforma* (wasp and bee nests); a diverse crevasse splay assemblage with a mix of vertical, subvertical, and horizontal burrows; and a fluvial channel assemblage dominated by vertical probable crayfish burrows (Zonneveld et al. 2000).

The main body in the greater Fossil Basin has a similar mammal assemblage to Fossil Butte. The most recent assessment, by Breithaupt (1990), identified leptictids, esthonychid tillodonts (extinct mammals), the pantodont *Coryphodon*, primates (including adapids and microsyopids), hyaenodontid creodonts, carnivorans (including miacids and viverravids), rodents, several condylarths (including hyopsodontids and phenacodontids), mesonychians, horses, tapiroids, and early artiodactyls (Breithaupt 1990). Other productive locations in Fossil Basin include an area west of Elk Mountain (20 km [12 miles] south-southwest of FOBU) low in the formation and sites on Bear River near Knight Station (75 km [47 miles] south-southwest of FOBU), closer in horizon to Fossil Butte (Gazin 1952, Gazin 1962). Other fossils from the main body in the basin include plants, bivalves, gastropods, and fish (Veatch 1907).

In general, the fossil assemblage of contemporaneous Wasatch Formation units associated with Lake Gosiute is similar to that associated with Fossil Lake. In addition to equivalents of the invertebrate trace fossils, reptiles, and mammals found in the Wasatch main body of Fossil Basin, the Wasatch main body and its tongues in the Green River Basin have the following: oncolites (microbialite balls) (Oriel 1962), stromatolites (Lawrence 1963), wood (Dorr and Gingerich 1980), seeds (Dorr 1978), snakes (Holroyd et al. 1997), the duck-flamingo-shorebird *Presbyornis* (Feduccia and McGrew 1974; Foster 2001) and other birds (Stidham and Holroyd 2004), multituberculates (rodent-like early mammals) (Dorr and Gingerich 1980), marsupials (West and Dawson 1973; Breithaupt 1990), dinoceratans (rhinoceros-like mammals) (Gazin 1962; West 1969), brontotheres (another group of rhinoceros-like mammals) (Foster 2001), enigmatic mammals (Zack 2004), salamander tracks, and eggshell fragments (Foster 2001).

Finally, reworked fossils have also been found in the Wasatch Formation. Reworked Jurassic fossils including bivalves have been found in the more conglomeratic peripheral beds ("Almy Formation") of southwestern Wyoming (Schultz 1914), and reworked Cretaceous-age fish and dinosaur teeth have been found in the Wasatch Formation of eastern Utah (Eaton et al. 1989). Fossils as old as the Paleozoic have been found in cobbles in conglomeratic Wasatch Formation rocks (Bradley 1964).

Green River Formation (Lower Eocene)

The Green River Formation of Fossil Basin is a heterogeneous lacustrine unit, including beds of limestone, marlstone, claystone, mudstone, siltstone, sandstone, oil shale, chert, and windblown volcanic debris (Oriel and Tracey 1970). Historically, the Green River Formation of FOBU and Fossil Basin is divided into the Fossil Butte and overlying Angelo members (Oriel and Tracey 1970; McGrew and Casilliano 1975; Rubey et al. 1975; Grande 1980; M'Gonigle and Dover 1992, 2004). More recently, it has been found that three units are present: the Road Hollow Member, Fossil Butte Member, and Angelo Member (Buchheim 1994a, 1998; Buchheim et al. 2011). These units correspond to the three phases of the lake described in the Geologic Background. The Road Hollow Member had previously been the portion of the Fossil Butte Member below the sandstone tongue of the Wasatch Formation (Buchheim et al. 2011). The contact between the Fossil Butte and Angelo members is revised as well under the tripartite division (Buchheim et al. 2011). The revised Fossil Butte and Angelo members are well-represented at Fossil Butte (Loewen and Buchheim 1998), but only a short section of the Road Hollow Member is exposed (less than 5 meters [18 ft]) (Buchheim et al. 2011).

Overall, the Green River Formation of Fossil Basin is drab in color, with gray to tan limestone and gray to brown shale as common general lithologies (M'Gonigle and Dover 1992, 2004). Thin laminations are abundant (Oriel and Tracey 1970). The laminations have often been described as varves, which imply even annual deposition, but it has been found that there are more and thicker laminations in nearshore sediments than in the center of the basin, which indicate that they are not varves. The laminations probably formed in response to multiple processes, including storm-related inflow (Buchheim 1994b; Buchheim and Eugster 1998). The formation becomes slightly thicker going west to east, from 160 m (525 ft) to 175 m (575 ft). It interfinger with adjacent Wasatch Formation rocks (M'Gonigle and Dover 1992, 2004). Lithologies differ markedly from location to location, based on conditions in Fossil Lake over time. Perennial lake, ephemeral lake, and other settings and subsets of settings have been studied in detail, providing a picture of the lake's history (Buchheim 1994a, 1994b; Buchheim and Eugster 1998; Buchheim et al. 2011). Historically, the age of the formation in Fossil Basin has been estimated from the age of mammals in the Wasatch Formation, but as noted in the Geologic Background the k-spar tuff in the upper Fossil Butte Member has been dated to 51.66 ± 0.09 Ma (Smith et al. 2008).

The Road Hollow Member is composed of a variety of lithologies, including sandstone, siltstone, mudstone, claystone, and carbonate rocks. The Fossil Butte Member is primarily composed of laminated micrite, with or without the organic compound kerogen, which can be used to produce fuel (hence the general name "oil shale"). The Angelo Member is primarily composed of dolomicrite, which has dolomite instead of calcite as the primary carbonate mineral. Large fluctuations in deposition, precipitation, and lake size occurred during the deposition of the Angelo Member (Loewen and Buchheim 1998).

Fossils of the Green River Formation in Fossil Basin

There are numerous fossiliferous layers known from the Green River Formation in Fossil Basin. Most of the known fossils are from the Fossil Butte Member (Breithaupt 1990). Locally, fossils may make up the greater part of a bed. These include concentrations of gastropods and ostracodes, and microbial limestones (Oriel and Tracey 1970).

Fossil Butte Member fossils are described in terms of F-1 and F-2, designations introduced by Grande (1980). These terms refer to two groups of localities. F-1 localities include all those sites with the "18-inch layer," including Fossil Butte itself and a number of quarries in the vicinity of the monument (Boundy, Craig, Haddenham, Jackson, Scott-Sommers, and Ulrich quarries, and most "Fossil, Wyoming" sites, among others) (Grande 1980). F-2 sites were first described from an area northeast of the F-1 cluster, in the northeast part of the basin. They include sites with "split-fish" layers, particularly most fossils from the Hebdon and Tynsky quarries (Grande 1980). These two designations represent specific taphonomic conditions, and broadly correspond to distinct settings. The F-1 sites are from the center of the basin, while the F-2 sites are

generally described as nearshore facies, but can be found in more central locations as well (A. Aase, pers. comm., October 2011). The F-1 sites are more stratigraphically limited than the F-2 sites, because there is only one 18-inch layer but multiple split-fish layers (Buchheim 1998), above and below the 18-inch layer (A. Aase, pers. comm., October 2011). Compared to the F-2 sites, F-1 sites have much more kerogen (organic chemicals) in their rocks, almost no bottom-dwelling (benthic) animals (Grande 1989), less diverse vertebrates (Grande and Buchheim 1994), and better-preserved fish (Buchheim and Eugster 1998; this difference can be attributed to F-1 fossils being found still covered with matrix, while F-2 specimens tend to split on the fossil, thus breaking their bones [A. Aase, pers. comm., October 2011). The F-1 sites also probably represent a shorter interval of time, perhaps a few hundred years, while the F-2 sites may represent thousands of years (Grande and Buchheim 1994). Fossil exploration is much older in the F-1 sites, with commercial operations in the F-2 sites only going back a few decades compared to roughly 120 years of F-1 collection (Grande 1989). Since the original work, other rock-fossil-depositional setting correlations have been found in the Green River Formation. For example, fossil fish are best preserved in laminated beds; the center of the lake has the best plant and insect fossils, with the exception of some thin shale-rich layers in split-fish "sandwich layer" beds (A. Aase, pers. comm., October 2011); the dolomitic beds representing hypersaline conditions sometimes have abundant ostracodes; bioturbated beds (heavily disturbed by organisms) from ephemeral lake and mudflat settings have mollusks, ostracodes, and fish; and sandstone and siltstone beds have mollusks and burrows (Buchheim 1994b; Buchheim and Eugster 1998).

The Fossil Butte Member is extensively fossiliferous within FOBU, although determining which finds came from within the monument's boundaries is complicated by vague, inaccurate, or entirely absent locality information, particularly for work done during the 19th century. For example, publications on Green River Formation plants by Lesquereux (1883) and Brown (1936) are presented as reports on specific locations when they are actually of floras from mixed locations (Grande 1980). Of the roughly two dozen known fish taxa (Grande 2001), eight have been identified from the monument: *Diplomystus dentatus*, *Knightia eocaena*, *K. alta*, *Mioplosus labracoides*, *Phareodus encausatus*, *P. testis*, *Priscacara serrata*, and "P." *liops*. *Knightia eocaena* is the most abundant (Hunt et al. 2006). Fossils of the other Fossil Lake fish taxa would probably be found at FOBU with sufficient exploration, with the exception of the rare genus *Esox* (A. Aase, pers. comm., October 2011).

The fossils of the Green River Formation of Fossil Basin and the Fossil Butte Member in particular are expansive topics. In brief, the Fossil Butte Member has a fossil assemblage including microbialites, spores, pollen, leaves, bivalves, gastropods, ostracodes, crayfish, prawns, insects, invertebrate traces, rays, diverse bony fish, salamanders, turtles, lizards, snakes, crocodilians, birds, and mammals including bats and ungulates (Oriel and Tracey 1970; McGrew and Casilliano 1975; Grande 1980, 1984; Breithaupt 1990; Grande and Buchheim 1994; Buchheim and Eugster 1998). In more detail:

Compared to Lake Gosiute and Lake Uinta, plants are poorly known from Fossil Lake (Grande 1980). Some descriptions and floral lists are provided by Brown (1936), Grande (1980, 1984), and Breithaupt (1990). More than 250 leaf types have been found in Fossil Basin, as well as more than 100 types of fruits, flowers, and other reproductive structures (FOBU Figure 4) (A. Aase, pers. comm., October 2011). Angiosperms are well-represented, and include water lilies (Grande 1994), monocots such as palms and cattails (FOBU Figure 5) (Grande 1980; Breithaupt 1990), and dicots such as oaks, sumacs, maples, tropical chestnuts (Brown 1936), poplars, roses, Keaki trees, soapberries, trees of heaven, legumes, torchwoods, spurges, and verbenas (Grande 1980; Breithaupt 1990). Ferns are also known from leaves, with at least 12 types present (A. Aase, pers comm., October 2011). Palynomorphs (organic microfossils such as pollen and spores) show the presence of ferns and conifers in the vicinity, as well as microscopic algae, acritarchs (organic microfossils that cannot be classified as anything else), and dinoflagellates (a type of algae that moves using two flagella) (Cushman 1998, 1999). 176 palynomorph forms are known from the Fossil Butte and lower Angelo members (Cushman 1998). The palynomorphs illustrate the types of vegetation growing around the lake over time. In general, a mixed mesophytic forest (not specialized for arid or humid conditions) grew near the lake. Moist lowlands and floodplains surrounded the lake and were in turn surrounded by upland forest in the mountains. Early in the lake's history, a dense forest grew. By the middle phase, low-elevation vegetation had partly displaced the upland vegetation under a possibly warmer and more humid climate. Finally, during the late phase the upland forest expanded again, under cooler, drier conditions (Cushman 1999).



FOBU Figure 4. A plant reproductive structure (VLS).



FOBU Figure 5.A palm leaf on display at FOBU (VLS).

Along with microbial forms, gastropods and ostracodes can be the dominant constituents of limestone in the basin. Beds of microbial, gastropod, and ostracodal limestone are mostly found around the basin margin between Sillem Ridge and the Absaroka fault (M'Gonigle and Dover 1992, 2004). Crustaceans are represented by the prawn *Bechleja rostrata* and the crayfish *Procambarus primaevus*, and are known from nearshore locations (Feldmann et al. 1981), with the occasional specimen found in weathered rock in 18-inch layer sites (A. Aase, pers. comm., October 2011). The insect assemblage is dominated by the bibionid fly *Plecia pealei* (march fly), which can make up 80% to 96% of insect specimens (Grande 1980). Other forms present include planthoppers, snout beetles (Grande 1980), odonates (dragonflies and damselflies), crane flies, bees, wasps, and lepidopterans (butterflies and moths) (A. Aase, pers. comm., October 2011). Insect fossils are most abundant in the oil shale known as the bottom capping layer underneath the 18-inch layer. Much more diverse insect assemblages are known from the other lakes, with dozens of family-level taxa reported (Grande 1980). Caddisfly larval cases have been found at shore sites (Loewen et al. 1999), in the Angelo Member (Buchheim et al. 2011). Arachnids are represented by several spider fossils (A. Aase, pers. comm., October 2011). Sediment-feeding invertebrates left traces in the lower part of the member (Buchheim 1998). Annelids are represented by trace fossils (Grande 1980). Nematode trails are known from Lake Uinta (Moussa 1970), but have not yet been found in Fossil Lake (Grande 1980). Other invertebrates found in the other lakes but not yet at Fossil Lake include sponges, conchostracans (clam shrimp), and mites (Grande 1980).

Fish are the most famous component of the Fossil Butte Member fossil assemblage (FOBU Figure 6), and Fossil Lake has the most diverse fish assemblage of the three lakes. The most recent taxonomic list was provided by Grande (2001), and the following discussion is primarily based on that source, with additions and modifications where noted. Cartilaginous fish in Fossil Lake are represented by two genera and species of rays, *Asterotrygon maloneyi* (Carvalho et al. 2004) and *Heliobatis radians*. Rays are uncommon, and are generally restricted to nearshore sites (Grande 2001), but have been found in the 18-inch layer (Carvalho et al. 2004). Ray-finned fish are represented by 21 named species. They include: the paddlefish *Crossopholis magnicaudatus*; several species of gar, including *Masillosteus janeae*, the short-snouted gar *Lepisosteus bemisi*, and the long-snouted gars *Atractosteus simplex* and *A. atrox* (Grande 2010); the bowfins *Amia pattersoni* and *Cyclurus gurleyi*; the mooneye *Hiodon falcatus* (Hilton and Grande 2008), a relatively rare nearshore taxon; the bonytongues *Phareodus encaustus* and *P. testis*; the herring-like *Diplomystus dentatus*; the herrings *Knightia eocaena* and *K. alta*; the gonorynchid (beaked salmon) *Notogoneus osculus*, known primarily from the 18-inch layer; the pike *Esox kronneri*, represented by one specimen; the trout-perch *Amphiplaga brachyptera* (trout-perch); rare *Asineops squamifrons*, from a now-extinct lineage; and several perch-like forms of uncertain relationships, including *Mioplosus labracoides*, *Priscacara serrata*, "P." *liops*, "P." *hypsacantha*. There are also two possible new unnamed perch-like species, one similar to "P." *hypsacantha* (Grande 2001), and one similar to *P. serrata* (A. Aase, pers. comm., October 2011), and an unnamed *Asineops*-like form (A. Aase, pers. comm., October 2011). The herrings and herring-like fish (clupeomorphs) are the most abundant forms. For example, of a set of 6,667 Fossil Lake fish, these fish made up 67% to 70% (Grande 2001), while Breithaupt (1990) found 25% to 50% of fish to be *Knightia* or *Diplomystus*.

It is likely that our limited knowledge of these fish as living animals conceals greater diversity: the family-level diversity is comparable to modern lakes, but many fewer species are recognized (Grande 2001). Only a few taxa in the Green River system are not found at Fossil Lake. These include the herring *Gosiutichthys parvus*, the rare sucker *Amyzon gosiutensis*, the catfish *Astephus antiquus* (potentially represented in Fossil Lake by a single lost specimen) and *Hypsidoris farsonensis*, and the rare trout perch *Erismatopterus levatus*, all of which are known from Lake Gosiute (Grande 2001), and the gar *Cuneatus cuneatus* in Lake Uinta (Grande 2010). Lake Gosiute in general is known for having more bottom-feeding fish such as suckers

and catfish than Fossil Lake (Grande 2001). Fish mass-death accumulations are known from Fossil Lake, as well as examples of aspiration (one fish died eating another) (Hunt et al. 2006). Two mass-death sites are known for *Priscacara*, and several are known for *Knightia* (Dayvault and Bersch 1983). The mass-death events may have occurred in late summer and early fall, as the result of microbial blooms (McGrew 1975). The Fossil Butte Member also has fish trace fossils, including the first reports of fish swim marks (*Undichna*) from between the Mesozoic and the Quaternary (Loewen and Gibert 1999; Martin et al. 2010), and delicate rope-like fish coprolites (fossil feces) (Edwards 1976); although fish body fossils are more famous, the most abundant fossils from Fossil Lake are coprolites, from fish and tetrapods (A. Aase, pers. comm., October 2011). The fish of the Green River Formation are sufficiently well-preserved to permit using them in studies of the evolution of modern groups, making the Green River Formation one of the few early Cenozoic sites where this is feasible (Grande 1989, 2001). They show a distinct Asiatic influence (Grande 1989).



FOBU Figure 6. As of this writing, there are nearly two dozen named species of Fossil Lake fish, with more to be described (VLS).

Amphibians and reptiles are represented by a handful of forms. The salamander *Paleoamphiuma tetradactylum* was named from an F-2 locality (Rieppel and Grande 1998). An undescribed Fossil Lake frog is in the collections of the Field Museum of Natural History (A. Aase, pers. comm., October 2011). Amphibians are better known from the other two lakes, although there are some additional Fossil Lake amphibian fossils in private hands (Gardner 1999). Turtles include chelydrids (snapping turtles) (Grande 1980; Breithaupt 1990), the emydids *Baptemys* (A. Aase, pers. comm., October 2011) and *Echmatemys*, the extinct baenids, and trionychids (Grande 1980; Breithaupt 1990), with *Trionyx* as the most abundant form (FOBU Figure 7) (Grande 1980). Three types of lizards have been found in the Fossil Butte Member: the anole *Afairiguana avius*, the shinisaurid (crocodile lizard) *Bahndwivici ammoskius*, and the monitor *Saniwa ensidens* (Conrad et al. 2007). Fossil Butte Member snakes are represented by the boa *Boavus idelmani* (Grande 1980) and an undescribed form, both known from only one specimen (A. Aase, pers. comm., October 2011). Crocodilian fossils include bones and coprolites (FOBU Figure 8) (Grande 1980). Crocodilian taxa present include *Borealosuchus* (formerly *Leidyosuchus*) *wilsoni* (Breithaupt 1990; Brochu 1997) and the small alligator *Procaimanoidea* (A. Aase, pers. comm., October 2011). Lake Uinta has large skin impressions and tracks from lizards, not yet known from Fossil Lake (Grande 1980).



FOBU Figure 7. A Trionyx*, a common turtle of Fossil Lake (VLS).*



FOBU Figure 8. A crocodilian coprolite. Coprolites are abundant in Fossil Lake rocks (VLS).

The bird assemblage includes a small but growing group of taxa. The most recent list of Green River Formation birds was published in Weidig (2010). Those known from the Fossil Butte Member include the following: the lithornithid *Pseudocrypturus cercanaxius* (Houde 1988); the frigatebirds *Limnofregata azygosternon* (the holotype specimen of which comes from an old commercial site now within FOBU; Olson 1977) and *L. hasegawai* (Olson and Matsuoka 2005); the presbyornithid *Presbyornis pervetus*; the landfowl *Gallinuloides wyomingensis*; the bittern-like *Messelornis nearctica* (Hesse 1992); the frogmouth *Fluvioviridavis platyrhamphus* (Nesbitt et al. 2011); the oilbird *Prefica nivea* (Olson 1987); *Primobucco mcgrewi* (Brodkorb 1970) and *Eozygodactylus americanus* (Weidig 2010), related to perching birds; parrot relatives *Avolatavis tenens* (Ksepka et al. 2012), *Cyrilavis colburnorum* (Ksepka et al. 2011), *Tynskya eocaena* (Mayr 2000), and "Primobucco" *olsoni* (Feduccia and Martin 1976); the mousebirds *Anneavis annae* (Houde and Olson 1992) and *Celericolius acriala* (Ksepka and Clarke 2010); and forms of uncertain relationships, including *Foro panarium* (Olson 1992) and *Plesiocathartes major* (Weidig 2006). Feathers and nests with eggs have also been found; FOBU has feathers in its collections (Grande 1980), including one assigned to the giant ground bird *Diatryma* (A. Aase, pers. comm., October 2011). *Presbyornis* is particularly abundant. Two nesting sites for this bird have been found near the base of the Road Hollow Member (Leggitt et al. 1998). The *Presbyornis* mass-death sites in Fossil Basin may have resulted from botulism outbreaks (Loewen and Buchheim 1998). Bird tracks have been found at Lake Uinta (Moussa 1968; Grande 1980) and Lake Gosiute (Sullivan 1980; Scott et al. 2007).

Mammal fossils are historically extremely rare in the Fossil Butte Member. The mammal assemblage includes: the arboreal insectivore *Apatemys* (von Koenigswald et al. 2005); an unnamed pantolestid, interpreted as an otter analogue (von Koenigswald et al. 2004); the bats *Icaronycteris index* (Grande 1980) and *Onychonyctris finneyi* (Simmons et al. 2008); the condylarth *Hyopsodus* (A. Aase, pers. comm., October 2011); the perissodactyl *Lambdotherium* (Froehlich and Breithaupt 1997, 2002); the horses *Orohippus* (Cushman 1998) and either *Protorohippus* or *Hyracotherium*, found in 2003, as well as two isolated limbs (A. Aase, pers. comm., October 2011); and an indeterminate forelimb (Grande and Buchheim 1994). More diverse assemblages are known from the other lakes, including marsupials, several families of primates, rodents, miacid carnivorans, tillodonts, hyopsodontid condylarths, brontotheres, and hyrachyid perissodactyls (Grande 1980); unsurprisingly, the assemblages are similar to those of the surrounding Wasatch Formation. Mammal tracks are known from Lake Uinta (Moussa 1968; Grande 1980).



FOBU Figure 9. Bats are among the rare mammals known from Fossil Lake (VLS).

As noted above, the Fossil Butte Member is the most fossiliferous member. The Road Hollow Member includes leaves, plant debris, bivalves gastropods, ostracodes, a handful of fish taxa (*Diplomystus*, *Knightia*, *Phareodus*, and *Priscacara*), bones turtles, birds, and mammals, crocodile teeth, and fish coprolites. Ostracodes are the most abundant fossils (Buchheim et al. 2011). The Angelo Member also has a reduced assemblage compared to the Fossil Butte Member. The diversity and abundance of fossils decrease going up-section (Loewen and Buchheim 1998), as well as the quality of preservation (Oriel and Tracey 1970). The upper Green River Formation (the Fossil Butte Member above the k-spar tuff and the Angelo Member) has an assemblage including stromatolites and tufa (a type of limestone) from cyanobacteria and algae, "algal-encrusted logs", acritarchs, dinoflagellates, fern spores, gymnosperm and angiosperm pollen, leaf and wood fragments,

palm fronds (*Sabalites*), seed pods, gastropods, ostracodes (Loewen and Buchheim 1998), caddisfly larvae (Buchheim et al. 2011), fish (*Asineops*, *Diplomystus*, *Knightia*, *Lepisosteus*, "Priscacara" *hypsacantha*, and "P." *liops*), the turtle *Trionyx*, crocodilians, *Presbyornis*, mammal bone fragments, large coprolites (Loewen and Buchheim 1998), and bird egg shells and tracks (Buchheim et al. 2011). Fish preferred the margins of the lake (Loewen and Buchheim 1998). During this phase, stromatolites became common, "Priscacara" *hypsacantha* was the most abundant fish (outside of several beds dominated by *Asineops* [A. Aase, pers. comm., October 2011), and *Lepisosteus* was the main predator (Loewen and Buchheim 1998).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary sediments of FOBU include a variety of sediments up to gravel in size, mostly deposited by fluvial processes or small local landslides (McGrew and Casilliano 1975; M'Gonigle and Dover 1992, 2004). They can be more than 5 m (16 ft) thick (M'Gonigle and Dover 1992, 2004). Most of the sediments are thought to be Holocene in age (M'Gonigle and Dover 1992, 2004), but there may be some gravels dating to the late Pliocene or early Pleistocene in the vicinity (Rubey et al. 1975). The Quaternary sediments of FOBU are fossiliferous, having yielded bison bones (A. Aase, pers. comm., October 2011).

Common Quaternary fossils of the Colorado Plateau and Great Basin, excluding large lakes like Lake Bonneville in western Utah, include bones of large mammals and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Faunmap (http://www.ucmp.berkeley.edu/faunmap/index.html) and the Neotoma database (http://ceiwin5.cei.psu.edu/weborb30/myapps/neotoma/#) show about two dozen Quaternary fossil sites within approximately 100 km (60 miles) of FOBU in southwestern Wyoming, northeastern Utah, and southeastern Idaho. However, most Quaternary sites are fairly distant from FOBU, with a much smaller subset within 50 km (30 miles). Interestingly, with the exception of a horse molar from the Fort Bridger area about 63 km (39 miles) southeast of FOBU (Hay 1924; Anderson 1974), the Wyoming sites are all Holocene in age, while the Utah and Idaho sites include Pleistocene material, which tend to be isolated terrestrial vertebrate finds associated with Lake Bonneville shorelines (Miller 1982; Nelson and Madsen 1980, 1987; Gillette and Miller 1999). The Pleistocene sites have yielded bones belonging to members of the extinct Pleistocene megafauna, such as ground sloths, sabre-toothed cats, short-faced bears, mastodons, mammoths, horses, camels, llamas, extinct mountain goats, and musk oxen (Heaton 1999).

The best-studied Quaternary fossil site near FOBU is probably Bear Lake. At its closest, Bear Lake is about 40 km (25 miles) west of FOBU. It has a detailed fossil and sedimentological record documenting at least the last 225,000 years (Jiménez-Moreno et al. 2007). Its fossils include diatoms (a type of algae with cell walls of silica) (Moser and Kimball 2009), pollen and other palynomorphs (Jiménez-Moreno et al. 2007; Doner 2009), bivalves, gastropods, ostracodes (Laabs and Kaufman 2003), mammoths (Gillette and Miller 1999; DeBlieux 2010), bighorn sheep, bison, and possibly musk oxen (Gillette and Miller 1999).

There are about a half-dozen Holocene sites within about 50 km (30 miles) of FOBU. Benton (2001) briefly described Holocene-age fossils from a fissure on Fossil Ridge, just south of FOBU. Material here includes seeds, bones of packrats and deer mice, packrat middens, and carnivoran droppings and raptor pellets; the latter fossils were probably brought in by packrats for their middens (Benton 2001). Otherwise, the nearest published packrat middens are from Oneida Narrows, Idaho, about 90 km (55 miles) northwest of FOBU (Smith and Betancourt 2003). Charcoal dating to a little over 3,000 years ago, rodents, rabbits, and an unidentified artiodactyl have been found at 48Ln1469, 20 km (12 miles) southeast of FOBU (McKern 1987). Nearby, fish, rodent, dog, pronghorn, and bison material has been found at the late prehistoric Skull Point archeological site (48Ln317), 30 km (19 miles) southeast of FOBU (McGuire 1977). The Taliaferro Site, an archeological site about 50 km (30 miles) east of FOBU, has a fossil record encompassing the last 5,000 years. Its fossils include pollen, charcoal, seeds, bivalves, fish, birds, rodents, rabbits, deer, and pronghorns (Smith and Creasman 1988). The Woodruff Bison Kill (42Ri1) is about 48 km (30 miles) southwest of FOBU. It is over 1,300 years old, and includes human-modified bison remains (Shields 1978). Finally, Holocene-age gastropods are known from Whitney Canyon, approximately 50 km (30 miles) southwest of FOBU near the Utah–Wyoming border (Gibbons 1983).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within FOBU. However, they are known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from one or more of them.

Woodside Redbeds (Lower Triassic)

The Woodside Redbeds (also known as Formation, Sandstone, Shale, or Siltstone) represents a terrestrial episode between marine regressions. The redbeds date to the early Early Triassic (Dienerian) (Paull et al. 1989), approximately 250.4 to 249.7 Ma. In Fossil Basin near FOBU, the Woodside Redbeds are composed of red siltstone and claystone, with some thin red

sandstone and interbeds of gray limestone (M'Gonigle and Dover 1992, 2004). They thicken going west to east, from 130 m (425 ft) to 200 m (650 ft), and are not resistant. The upper contact with the Thaynes Limestone is intertonguing (Rubey et al. 1975). Veatch (1907) thought that the Woodside Redbeds and Thaynes Limestone of Fossil Basin were "Permo–Carboniferous" in age, but Schultz (1914) recognized them as Triassic in age. The Woodside Red Beds were not reported as present in FOBU by McGrew and Casilliano (1975), but are present with the Thaynes Limestone in a small area southwest of Rubey Point in FOBU (M'Gonigle and Dover 1992, 2004).

Fossils are very rare in the Woodside Redbeds, and are not known from Fossil Basin (Schultz 1914). Reports include bivalves near the top of the formation, below the Thaynes Limestone, in the Fort Hall area of southeastern Idaho (Mansfield 1916, 1920).

Thaynes Limestone (Lower Triassic)

The Thaynes Limestone (or Formation) is a widespread marine unit, exposed from eastern Nevada to southwestern Montana. At FOBU, it can only be found in a small area southwest of Rubey Point (McGrew and Casilliano 1975; M'Gonigle and Dover 1992, 2004). Within Fossil Basin, the Thaynes Limestone is a resistant formation with distinct lower and upper sections. The lowest third of the formation is composed of green or gray lime-rich siltstone, silt-rich claystone, and silt-rich limestone (Rubey et al. 1975); these rocks weather to a brown color (McGrew and Casilliano 1975). The remainder of the formation includes more limestone (often silt- or sand-rich) and weathers to a pale yellow or gray color. The thickness of the formation changes substantially in the basin, from nearly 400 m (1,300 ft) in the northwest to only 215 m (700 ft) in the southeast (Rubey et al. 1975). The Thaynes Limestone is sometimes divided into a number of members depending on location (Kummel 1943, 1954; Picard et al. 1969; Massare and Callaway 1994), but this has not been done in Fossil Basin. Recent work in southern Utah has advocated raising the Thaynes Limestone to a group and incorporating several other dominantly marine units (Lucas et al. 2007).

At its most extensive, the Thaynes Limestone was deposited during the Early Triassic (the Smithian and Spathian stages) (Solien 1979; Solien et al. 1979), approximately 250 to 245 Ma. The Thaynes Limestone of Fossil Basin may be somewhat shorter in duration, as it is not as thick as neighboring areas to the west (Kummel 1954). This marine formation was deposited by a sea that spread instantaneously relative to geologic time (Paull et al. 1989). FOBU was in an area that was usually within the outer shelf to inner shelf belts (Pisera et al. 1996). Depositional environments in the formation ranged from open marine to tidal flat settings, with shallower environments appearing over time (James 1980).

The Thaynes Limestone is not known to be fossiliferous within FOBU at this time, but its proven fossil resources elsewhere make it a good candidate. This formation has long been known as extensively fossiliferous. Veatch (1907) reported it as "very fossiliferous" in Fossil Basin, with its assemblage including brachiopods, bivalves, ammonites, gastropods, and crinoids. Schultz (1914) wrote of the existence of many Thaynes Limestone fossil sites in the basin.

The fossil assemblage of the Thaynes Limestone is dominated by shelled marine invertebrates, particularly mollusks (Schubert and Bottjer 1995). Algae, microbes, and other unicellular organisms are represented by organic mounds (Picard et al. 1969), microbial limestone, stromatolites (James 1980), wrinkle structures (Pruss et al. 2004), and foraminifera (amoeba-like protists that form "shells"). Invertebrates are represented by sponges, brachiopods (lamp shells), bivalves, cephalopods (including ammonites and nautiloids), gastropods, ostracodes (Picard et al. 1969; Solien 1979; Solien et al. 1979; James 1980), lobster-like crustaceans (Schram 1971), crinoids (sea lilies), echinoids (sea urchins), ophiuroids (brittle stars), and trace fossils, including resting traces of asteroids (sea stars) (Picard et al. 1969; Solien 1979; Solien et al. 1979; James 1980). Vertebrates and their relatives are represented by conodonts (extinct eel-like chordates) (Solien 1979; Solien et al. 1979; James 1980), sharks (Schroeder 1968; Mutter et al. 2005), bony fish (Stephenson 1965), the ichthyosaur *Cymbospondylus* (Massare and Callaway 1994), and bone fragments (Kummel 1954). Shells are sometimes represented by injected siltstone casts (Boyd et al. 1999; Boyd and Newell 2002).

Thaynes Limestone fossils are often discussed in terms of recovery from the catastrophic Permian–Triassic extinction event (Schubert and Bottjer 1995). Aspects of interest include shell beds with only one overwhelmingly dominant species (Boyer et al. 2004), gastropods of unusually small size (Fraiser et al. 2005), and distinctive microbially-created wrinkle structures (Pruss et al. 2004). These are used as evidence for a slow recovery; there is evidence, however, that a slow recovery was not a universal phenomenon for Thaynes Limestone time (Brayard et al. 2010).

Park Collections

Complicating any accounting of specimens from FOBU is the vague locality information provided by many early workers, and nearly a century of commercial collecting from sites now near or within FOBU. A complete accounting of private holdings of material from what is now FOBU is not feasible. For a NPS-related example of the wide dissemination of fossils from Fossil Basin, a member of the Billings or Rockefeller families associated with the house and grounds now designated Marsh-

Billings-Rockefeller National Historical Park (MABI) in Vermont purchased a "Priscacara" *liops* (MABI 3940) during the late 19th or early 20th centuries. The specimen probably derives from the 18-inch layer, and could well come from a site now within FOBU (Tweet et al. 2010). Private collections include not only specimens held loose, but fossils used as decorative stone in furnishings, a practice noted by Hesse (1939). Commercial collecting goes back at least to the 1870s, when the Bell brothers were operating at Twin Creek (Peale 1879). Specimens of abundant fish such as *Diplomystus* and *Priscacara* could be purchased inexpensively: the former could be had for as little as $1.50, and the latter for $2.00 during the late 1930s (Hesse 1939). Early on, the fossils were primarily sold at the Green River Village train station, where passengers were shown various fossils according to their train tickets: tourist car passengers saw the cheapest specimens ($0.50 to $1 *Knightia* specimens and more expensive "sunfish"), Pullman car passengers were shown specimens priced at $25 to $35, and stateroom passengers were shown "museum pieces" priced around $100 (Stucker 1966).

Many institutions have substantial collections of Fossil Basin fossils. Hesse (1939) reported that few museums of his time lacked fossil fish from the Green River Formation of southwestern Wyoming. It is known that fish from Fossil are at the National Museum of Natural History (USNM) (Veatch 1907) and Yale (YPM) (Thorpe 1938). Per Li et al. (1997), Cope's (1877) *Phareodus testis* was named from Twin Creek (USNM 4014), but as noted there is some question regarding whether or not he'd found Twin Creek by that time. The bird *Limnofregata azygosternon* is based on USNM 22753, found at the former Haddenham site, which is now part of FOBU (Olson 1977). Researchers from Princeton collected fossil vertebrates from the Wasatch Formation (=Knight Formation of previous usage) from the Fossil Butte area. As mentioned earlier, the holotype specimens of *Copelemur australotutus* (USNM 22261) (Beard 1988) and *Hexacodus uintensis* (PU 16175) (Gazin 1952) come from what is now FOBU. The list of institutions in the "Data Sets" section at the end of this document is compiled from institutions mentioned in publications as having Fossil Basin specimens; thus, not all of the institutions necessarily have fossils from within FOBU, but the wide net should ensure that significant collections that are from FOBU have not been overlooked.

FOBU's own collections are extensive. For example, a publically-available file on the FOBU website (http://www.nps.gov/fobu/naturescience/upload/Fossil%20Lake%20Quarry%20data%201998-2007.xls) provides a list of over 2,500 fossils of Green River Formation plants, insects, and fish documented from the monument's research quarry between 1998 and 2007. Many of these fossils were collected. FOBU also has examples of material collected from elsewhere in the basin, such as the holotype of the fruit *Lagokarpos lacustris* (McMurran and Manchester 2010), and casts of the bat *Onychonyctris* (Simmons et al. 2008), several other mammals including a pantolestid, *Apatemys*, *Hyopsodus*, and *Hyracotherium*, the lizard *Saniwa*, and an undescribed snake (A. Aase, pers. comm., October 2011). FOBU obtained the Green River Formation paleobotanical collection of William Tidwell (Brigham Young University) in the early part of the first decade of the 2000s.

Cultural Resource Contexts

FOBU has the potential for fossil-cultural resource associations, but none have been recognized yet. It is possible that archeological investigations may find fossil materials used as tools, as they have been found at archeological sites in the vicinity. Chert projectile points were found at Skull Point (McGuire 1977), and microbial chert was used to make some of the artifacts found at the Taliaferro Site (Smith and Creasman 1988). In a more general sense of cultural context, the bountiful fossils of Fossil Butte are part of the culture of Kemmerer and the surrounding area of southwestern Wyoming, and are a source of local pride. Fossil Lake fossils have long been visible to the public at large; for example, a Fossil Lake gar was shown at the 1893 Chicago World's Fair (Stucker 1966). FOBU also has historical artifacts from the early days of fossil collecting in Fossil Basin. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- FOBU reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 95 localities, all of which were reported to be in good condition in 2010. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring, with FOBU as one of the case studies (Santucci et al. 2009).

- The geology and paleontology of Fossil Lake are being brought to light not only by FOBU, but also by the various commercial fossil concerns in the vicinity. FOBU wants to maintain "good working relationships

with commercial collectors focused on shared information so both parties benefit" (A. Aase, pers. comm., October 2011).

- Because of the abundance and publicity of fossils at FOBU, fossil resources are at increased risk for fossil theft and vandalism. These conditions also make FOBU an ideal location for studying visitor attitudes to fossils. For example, Karen Hockett from Virginia Polytechnic Institute and State University wrote her doctoral dissertation on visitor attitudes to fossil resource protection at FOBU, and co-wrote a report for the monument (Hockett and Roggenbuck 2003; Hockett 2008). Further research in this direction could be valuable, not only for FOBU but for other parks and monuments with extensive fossil resources.

- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented. Usually, it is recommended that fossils be left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts; however, both the Wasatch Formation and the Green River Formation erode quickly. Thus, scientifically important fossils should be collected (A. Aase, pers. comm., October 2011). When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Marsh-Billings-Rockefeller National Historical Park
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Mesalands Dinosaur Museum

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National Museum of Natural History
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Princeton Museum of Natural History
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Royal Ontario Museum
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Golden Spike National Historic Site

Golden Spike National Historic Site (GOSP), Utah, was established to commemorate the completion of the first transcontinental railroad in the United States, an event that took place May 10, 1869, at Promontory, Utah (GOSP Figure 1). GOSP was designated as a historic site April 2, 1957, and authorized as a unit of the National Park Service July 30, 1965. Its boundaries have remained the same, but the area under federal ownership has changed twice, on July 30, 1965 and September 8, 1980. GOSP encompasses 1,106.93 ha (2,735.28 acres), 891.60 ha (2,203.20 acres) of which are under federal administration (GOSP Figure 2).

Geologic Background

GOSP is located north of Great Salt Lake in eastern Box Elder County, within the Bonneville Basin, which forms the northeastern Great Basin. The site is situated between the Promontory Mountains on the southeast and the North Promontory Mountains on the northwest. Brigham City is about 48 km (30 miles) to the east-southeast, and Ogden is about 60 km (37 miles) to the southeast. The Geologic Resources Division of the National Park Service (NPS) coordinated a geologic resources inventory workshop for GOSP during June 1999 (NPS 1999), leading to a geologic report for the site in 2006 (Thornberry-Ehrlich 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002). A paleontological resource inventory was undertaken during 2007 and 2008 (Aase undated). Fossils from GOSP have been mentioned in NPS (1999), Rhode (2000), Madsen et al. (2001), Koch and Santucci (2002), Thornberry-Ehrlich (2006), Santucci and Kirkland (2010), and Aase (undated).



GOSP Figure 1. Much of GOSP's geology is hidden by cover, but there are some areas of outcrops (VLS).

The geologic history exposed at GOSP is limited to a short stretch of the Late Paleozoic (see the appendix for a geologic time scale), and the past few tens of thousands of years. Bedrock is represented by the lower portion of the Oquirrh Formation, deposited roughly 320 Ma (million years ago) by a continental sea. Its variable rock types reflect fluctuating water depth (Bissell 1959) and changing sediment sources (Jordan 1979). Deposition of the Oquirrh Formation occurred within a topographic low called the Oquirrh Basin, between relative highs on the east, west, and southeast (Davis et al. 1994). The basin seems to be divisible into northern and southern sub-basins with distinct stratigraphy; GOSP would be in the northern sub-basin (D. Clark, Utah Geological Survey geologist, pers. comm., January 2012). Over 5 km (3 miles) of sediment accumulated in this basin. Many aspects of the Oquirrh Formation and Oquirrh Basin are poorly known, such as the sources of the sediment, the depositional history, and depositional environments, and nomenclature for the rocks has been slow to develop (Hintze and Kowallis 2009). Depth increased in the basin over time (Chamberlain and Clark 1973), possibly to over 330 m (1,000 ft) in places (Hintze and Kowallis 2009). The changes in depth were not distributed evenly (Jordan 1979), and it appears that the margins of the basin underwent rapid uplift near the Pennsylvanian–Permian boundary (Jordan and Douglass 1980). Later, during the Cretaceous, the rocks of the Oquirrh Basin were transported approximately 65 km (40 miles) to the southeast by thrust faulting (Hintze and Kowallis 2009).

Deposition is next recorded at GOSP for the late Pleistocene. Before that, the surroundings were altered by Basin and Range extension, a prolonged period of faulting that began after 20 Ma and gave the Great Basin its distinctive topography (Willis et al. 2010). Large faults are present in the vicinity of GOSP, including one at the western base of the North Promontory Mountains (McCalpin et al. 1987). To the northwest of GOSP, Hansel Valley, bounded on the east by the North Promontory Mountains, has been a site of damaging earthquakes in the past few hundred years. These include the largest historic earthquakes in the state (McCalpin et al. 1987; Robinson and McCalpin 1987).

In addition to faulting and earthquakes, northern Utah has been affected by volcanism during the past few million years. Sporadic volcanic events are recorded as ash beds in Great Salt Lake cores (Davis 1998), with five named ashfalls between 6.5 and 0.6 Ma (Davis and Moutoux 1998) and at least three events during the Lake Bonneville cycle (see below) (Oviatt and Nash 1989). A number of igneous flows from 8.6 to 0.44 Ma have been found in Curlew Valley, the large valley to the west of Hansel Valley. Some of these igneous events may be responsible for some of the lake ash beds (Felger et al. 2008; Felger et al. in prep).

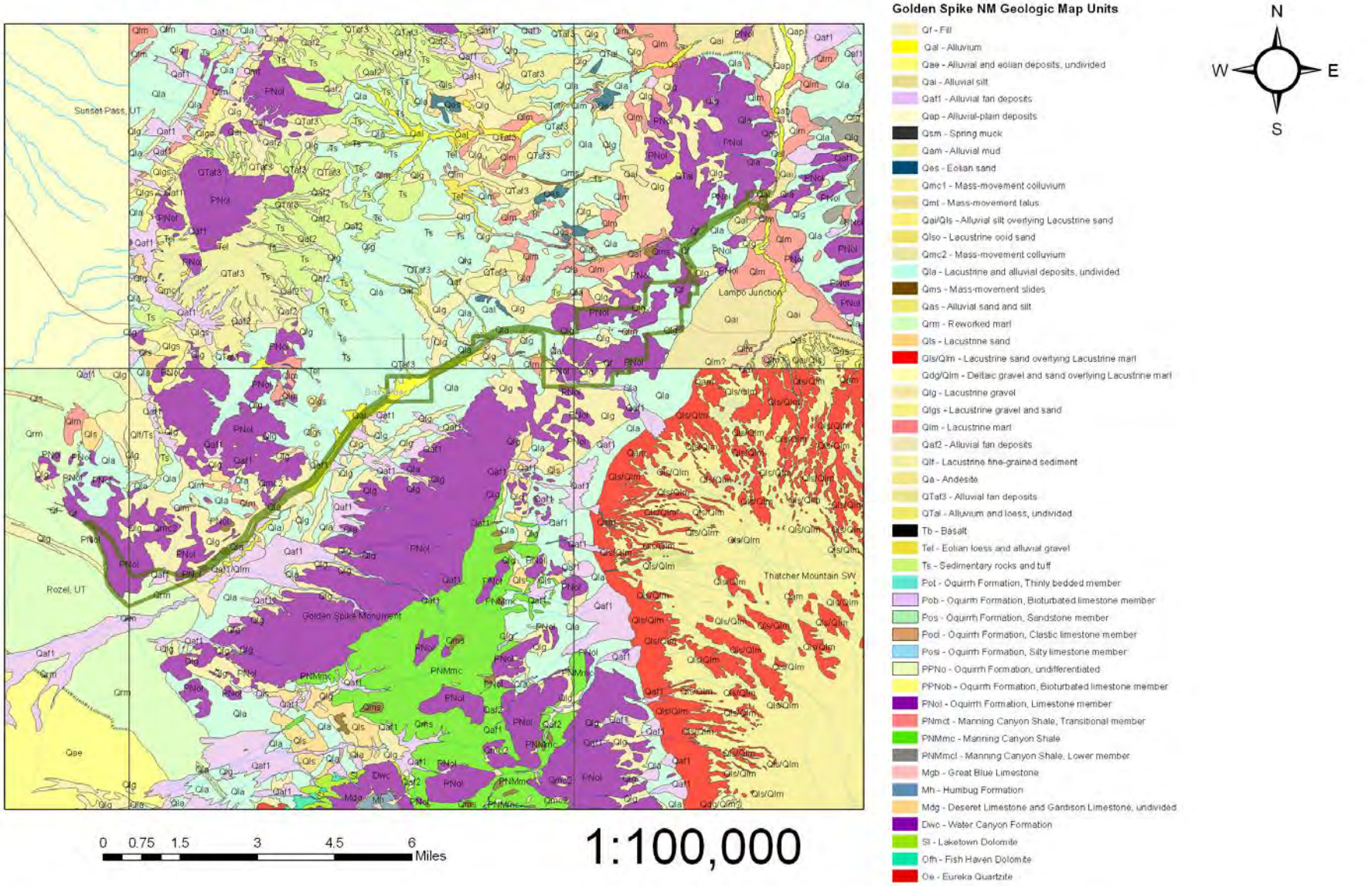
Much of the Bonneville Basin, including GOSP, was occupied by a lake system during portions of the Pleistocene. Marsh and shallow lacustrine environments appeared in the north part of the basin around 5 Ma. An open lake was present in the north beginning about 800,000 to 600,000 years ago, but was only established in the south part of the basin around 300,000 years ago (Kowalewska and Cohen 1998). Deposits from at least four lake cycles can be distinguished in Hansel Valley (Robinson and McCalpin 1987). The four most recent cycles before Great Salt Lake, from oldest to youngest, are the Pokes Point, Little Valley, Cutler Dam, and Lake Bonneville cycles (Balch et al. 2005). The largest iteration, Lake Bonneville, occupied 51,700 km^2 (20,000 $miles^2$) at its greatest extent (Rhode et al. 2005), reaching its maximum elevation of about 1,550 m (5,090 ft) above sea level around 19,000 years ago (Oviatt 1997). The lake dropped during a catastrophic flooding event around 17,400 years ago, and then dropped to a level near that of the present Great Salt Lake some 3,000 years later as the climate changed. This has been followed by over 14,000 years of more modest fluctuations. Great Salt Lake succeeded Lake Bonneville about 11,500 years ago (Willis et al. 2010).

Cold climate conditions were present during the Pleistocene, with the northern part of the Bonneville Basin relatively drier than the southern part (Madsen et al. 2001). Cold montane steppe vegetation dominated by sagebrush prevailed near the end of the Pleistocene. Limber pine woodlands were widespread at low elevations just before the Pleistocene–Holocene boundary. Warming and drying conditions caused the retreat and disappearance of the Pleistocene flora, replaced by desert scrub (Rhode and Madsen 1995; Madsen 1996). The Early Holocene of the northern Great Basin was cool and moist. This climate regime was replaced by a hot and dry regime during the middle Holocene (Grayson 2002), which then moderated before turning warmer and drier again in the past 2,000 years (Rhode 2000). Humans were present in the Bonneville Basin perhaps before 12,300 years ago, and had reached the east side of Lake Bonneville over 11,000 years ago (Rhode et al. 2005).

The only bedrock unit exposed at GOSP is a limestone member from the base of the Oquirrh Formation (Pennsylvanian). Surficial units include various Quaternary lacustrine deposits from Lake Bonneville and younger alluvial sediments (see GOSP Figure 3 for a geological map) (Miller et al. 1991; Miller and Schneyer 1994; Miller 2000; Miller and Crittenden 2000, 2004; Thornberry-Ehrlich 2006). Fossils of marine invertebrates are known from the Oquirrh Formation within the historic site, and Holocene-age packrat middens have been discussed in print as well (GOSP Table 1).



GOSP Figure 2. Geography of GOSP (NPS).



GOSP Figure 3. Schematic geological map of GOSP. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1043034.

Formation	Age	Fossils Within GOSP	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Packrat middens	Primarily lacustrine and alluvial deposits
Oquirrh Formation	Pennsylvanian	Sponges, solitary and colonial corals, bryozoans, brachiopods, nautiloids, gastropods, trilobites, crinoids, echinoids, shark teeth, and fish bones	Open marine to a variety of coastal marine settings

GOSP Table 1. Concise tabulation of GOSP stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of GOSP present opportunities for education, interpretation, and continued or future scientific research in the monument.

Oquirrh Formation (Pennsylvanian)

The Oquirrh Formation (or Group; the stratigraphic nomenclature for this unit is complicated) is a limestone, sandstone, quartzite, and siltstone unit deposited across parts of northwestern Utah and southern Idaho. It is noted for its lithological variability (Jordan 1981), although this characteristic can be overstated: in the Promontory Mountains, individual limestone beds can be followed over kilometer-scale distances (Crittenden 1988). In and around GOSP, the Oquirrh Formation is informally divided into several members. Thornberry-Ehrlich (2006) includes five members; from the published geologic maps of the area, three are commonly used. From oldest to youngest, these are a limestone member, a bioturbated (disrupted by biological activity) limestone member, and a thinly-bedded member (Jordan et al. 1988; Miller et al. 1991; Miller and Schneyer 1994). Together, these three members can be over 3,000 m (10,000 ft) thick (Jordan et al. 1988). The limestone member at the base of the formation is the only unit exposed within GOSP (Thornberry-Ehrlich 2006). It is a fossiliferous unit mostly composed of beds of gray limestone, with minor brown sandstone and some chert (Jordan et al. 1988; Miller et al. 1991; Miller and Schneyer 1994). Deposition of the limestone member occurred during the Pennsylvanian, but the three cited maps differ on the specific subdivision of the Pennsylvanian.

As at GOSP, the lowest part of the Oquirrh Formation is often composed of limestone, and several names have been applied depending on location. In much of Utah the basal limestone is known as the West Canyon Limestone, in the northern Oquirrh Mountains it is the Lake Point Limestone, and in southern Idaho it is the Bannock Peak Limestone (Davis et al. 1994). Other names used include the Hall Canyon Member (obsolete) (Bissell 1959) and the Bridal Veil Limestone Member in the Wasatch Range (Baker and Crittenden 1961). Together, these units are interpreted as having been deposited on carbonate platforms during the latest Mississippian (latest Chesterian stage) to the early Middle Pennsylvanian (earliest Atokan stage) (Davis et al. 1994), approximately 320 to 310 Ma. Dating is done via fossil content, particularly fusulinid foraminifera (foraminifera are amoeba-like protists that form "shells," and fusulinids are a large extinct type) and conodonts (eel-like extinct vertebrates). Olson (1960) found the oldest Oquirrh Formation rocks of the Promontory Mountains to be at least Morrowan in age, which dates to approximately 318 to 312 Ma. Depositional environments include lagoons, bays, and other areas of restricted circulation, to open marine settings with depths of 10 to 200 m (30 to 660 ft) (Davis et al. 1994). Because of the widespread nature of these conditions, the lower limestone at GOSP probably has a similar history.

The lower limestone of the Oquirrh Formation is fossiliferous within GOSP. Arvid Aase (Fossil Butte National Monument) documented 61 fossiliferous sites during 2007 and 2008 (Aase undated). Sites included *in situ* specimens and loose fossiliferous stones. His report includes the following:

> Regional exposures of the Oquirrh Formation are commonly nonfossiliferous with locally abundant invertebrates and rare shark teeth. Exposures in GOSP follow this regional trend. Most limestone beds are nonfossiliferous or have widely separated, poorly preserved fossils. Locally fossils are abundant or individual limestone beds may contain more fossils than adjacent beds. In these cases the fossils tend to be better preserved.
>
> Inventory of the park's exposed rock produced 61 localities. Each locality is documented on a Paleontological Locality Form which includes GPS location, brief verbal description of the rock and what fossils were present. The invertebrate fossils are most commonly preserved by recrystallization of the original aragonite into calcite [different forms of calcium carbonate]. These fossils are generally exposed as white patterns worn flush with the gray rock. Locally fossils on bedding surfaces may be silicified. Silica is more resistant to erosion therefore the fossils stand out in relief and more delicate fossils can be seen such as fenestrate (lacy) bryozoans [moss animals], trilobite

sclerites [hard parts] and brachiopod [lamp shells] spines. Because of their porosity bryozoan mounds are preferentially silicified adjacent to recrystallized brachiopods, crinoids [sea lilies] and corals. Phosphatic fossils, represented by two shark teeth, one small patch of broken fish bones, and lingulid brachiopod shells, preserve as black to blue bone or shell.

Nearly all fossils exposed in GOSP are in massive limestones and therefore not susceptible to theft and are extremely slow to weather. Additionally, many of the fossils are recrystallized and therefore less readily identified by the general public as a fossil. Therefore monitoring is not necessary for fossil localities in GOSP except #13. Locality 13 has brachiopods in thin, easily accessible sandy beds in a railroad cut. This locality should be monitored annually by using photographs to determine if the beds have been disturbed. (Aase undated)

By far the most common fossils reported by Aase are crinoids (present at 43 locations) and brachiopods (present at 41 sites). Other fossil types, in order of decreasing abundance, include solitary corals (present at 14 sites; these are typically the extinct forms known as rugose or horn corals), bryozoans of various types (13 sites), colonial corals (5 sites), gastropods (possibly 5 sites), shark teeth (2 sites), trilobites (2 sites), echinoids (sea urchins; 1 site), fish bones (1 site), nautiloids (1 site), and sponges (1 site). All of the localities were in limestone. The great majority (52) were in micritic limestone, a very fine-grained limestone formed from mud-sized calcium carbonate particles.

In general, the other basal limestone units of the Oquirrh Formation have similar fossil assemblages to what is known from GOSP. Fossils of these units are dominated by marine invertebrates. The assemblage includes oncolites (layered spherical structures) and other microbial structures, foraminifera, sponges, corals, bryozoans, brachiopods, bivalves, cephalopods, gastropods, scaphopods (tusk shells), ostracodes (seed shrimp), trilobites, crinoids, echinoids, invertebrate trace fossils (burrows, fecal pellets, and general bioturbation), conodonts, and fish teeth (Beus 1968; Tooker and Roberts 1970; Smith 1983; Davis et al. 1989, 1994). Fossils recovered from the Oquirrh Formation of the Promontory Mountains include foraminifera, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, and crinoids (Olson 1960).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary sediments of GOSP include a variety of sediments, from gravel-sized to clay-sized, deposited primarily as a result of Lake Bonneville or as more recent alluvium (Thornberry-Ehrlich 2006). Typical Quaternary fossils of the Great Basin include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as packrat middens). GOSP has already yielded Holocene-age packrat middens, and because of its Lake Bonneville history it may have other paleoecologically-useful fossils as well. Summaries of Quaternary fossil sites in the vicinity in Utah include Nelson and Madsen (1980, 1987), Jefferson et al. (1994), and Gillette and Miller (1999).

The Quaternary fossil record of GOSP consists to date of two packrat middens described by David Rhode (Desert Research Institute–Reno) (Rhode 2000; pers. comm., November 2010). One dates to 3,670 ± 50 radiocarbon years before present (4,100 to 3,870 calendar years before present) and the other dates to 3,110 ± 50 radiocarbon years before present (3,420 to 3,210 calendar years before present); both are from an elevation of 1,495 m (4,900 ft). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate. Most of the plant taxa from the middens were shrubs, forbs, succulents, and grasses, with Utah juniper as the only identified tree taxon; today the juniper is only found at higher elevations (Rhode 2000). The NPS geological resources inventory workshop for GOSP also noted the presence of packrat middens from an undisclosed cave locality within GOSP (NPS 1999), which may be a reference to Rhode's middens. Santucci et al. (2001) provided an overview and cited selected examples of National Park Service fossils found in caves.

The middens at GOSP are part of a much larger set of samples (Rhode 2000); Madsen (1996) reported the existence of over 100 midden sites on the northern margin of the Bonneville Basin, ranging in age from over 50,000 years old to a few hundred years old. Only a few have been published; aside from Homestead Cave (Smith and Betancourt 2003), described below, published midden sites within 100 km (60 miles) of GOSP are limited to several sites on Antelope Island, over 64 km (40 miles) to the southwest of GOSP (Rhode 2000), and Oneida Narrows in Idaho, about 80 km (50 miles) northeast of GOSP (Smith and Betancourt 2003).

The Lake Bonneville sediments at GOSP are likely candidates for future fossil discoveries. GOSP is within the Bonneville shoreline level (locally just above 1,580 m [5,200 ft]) (Miller et al. 1991; Miller and Schneyer 1994), and the Provo shoreline (locally about 1,480 m [4,840 ft]) passes through parts of northern GOSP (Miller et al. 1991), indicating lacustrine deposition over much of GOSP between approximately 22,000 to 14,000 years ago, at least (Oviatt 1997; Murchison and Mulvey 2000; Godsey et al. 2005; Willis et al. 2010). Fossils associated with the Bonneville Basin lake system include diatoms (a type of algae with cell walls of silica) (Balch et al. 2005), a variety of palynomorphs (organic microfossils, in this case including freshwater algae, dinoflagellate plankton, pollen, and fungal spores; Davis and Moutoux 1998), charcoal, leaf hairs (Balch et al. 2005), wood (Spencer et al. 1984), roots

(Sprinkle 1993), other plant fragments (Oviatt et al. 2005), sponges, bivalves, gastropods, brine shrimp (body fragments, eggs, and pellets), ostracodes, brine fly larval claws (Balch et al. 2005), grasshoppers (Shaver 2000), invertebrate borings (Spencer et al. 1984) and other bioturbation (Robinson and McCalpin 1987), and fish (Smith et al. 1968; Murchison 1989). The most common fossils are bivalves, gastropods, and ostracodes (Santucci and Kirkland 2010). The Bonneville Basin hosts one of the most continuous records of the Neogene in North America (Balch et al. 2005). Well sites on Great Salt Lake have provided cores with pollen records going back at least 13.5 million years, and sedimentation back to 38 Ma (Davis and Moutoux 1998). Its fossil record is complemented by that of Bear Lake. At its closest, Bear Lake is about 90 km (55 miles) northeast of GOSP. Like the Lake Bonneville system, it too fluctuated in depth over the past few hundred thousand years (Laabs and Kaufman 2003; Kaufman et al. 2009). Its fossils include diatoms (Moser and Kimball 2009), pollen and other palynomorphs (Jiménez-Moreno et al. 2007; Doner 2009), bivalves, gastropods, ostracodes (Laabs and Kaufman 2003), mammoths (Gillette and Miller 1999; DeBlieux 2010), bighorn sheep, bison, and possibly musk oxen (Gillette and Miller 1999).

While lacustrine fossils can be encountered throughout the Bonneville Basin, there are a few specific reports of interest for GOSP. Olson (1960) reported ostracodes from the Promontory Mountains. McCoy (1987) recovered gastropods from the south end of the Promontory Mountains and near Corinne, about 31 km (19 miles) southeast of GOSP. Murchison (1989) found gastropods and fish bones dating to near the end of the Pleistocene at a site 17.5 km (10.9 miles) northwest of Corinne, which must be not far from GOSP. Robinson and McCalpin (1987) reported gastropods, ostracodes, and bioturbation in Hansel Valley lake sediments. The white marl includes gastropods and ostracodes, and can be bioturbated (Green 1988).

The Quaternary terrestrial vertebrate record of northern Utah and southeastern Idaho is not as comprehensive as the lacustrine fossils of the Bonneville Basin. There are numerous isolated terrestrial vertebrate finds associated with Lake Bonneville shorelines (Miller 1982; Nelson and Madsen 1980, 1987; Gillette and Miller 1999). They have yielded bones belonging to members of the extinct Pleistocene megafauna, such as ground sloths, sabre-toothed cats, short-faced bears, mastodons, mammoths, horses, camels, llamas, extinct mountain goats, and musk oxen (Heaton 1999). Searching Faunmap (http://www.ucmp.berkeley.edu/faunmap/index.html) shows about a dozen vertebrate sites within approximately 100 km (60 miles) of GOSP in northern Utah, with another handful in Idaho. By far the best-known vertebrate site in northern Utah is Homestead Cave.

Homestead Cave is about 50 km (30 miles) southwest of GOSP, on the west side of Great Salt Lake. This site has yielded over 200,000 elements belonging to insects, fish, reptiles, birds, and mammals. The great majority of specimens belong to mammals, with 183,798 mammal bones and teeth identified to genus or species level (Madsen et al. 2001). The Homestead Cave assemblage was produced by roosting owls and packrats between 11,270 ± 135 radiocarbon years before present (13,400 to 12,800 calendar years before present) and 1,020 ± 40 radiocarbon years before present (1,005 to 890 calendar years before present) (Madsen 2000). Other fossils from the site include hackberry seeds, a variety of other plant remains (such as wood, twigs, seeds, leaves, needles, flowers, and bark), bivalves, gastropods, ostracodes, and coprolites (fossil feces) from packrats, lagomorphs, artiodactyls, and possibly mice and carnivorans (Hunt et al. 2000). Some 14,000 fish specimens have been recovered (Broughton et al. 2006), representing 11 freshwater taxa (Broughton et al. 2000; Broughton 2000). Most are from the lowest strata in the cave, before the lake became inhospitable (Broughton et al. 2000). The bird record is diverse, including shore birds, waterfowl, raptors, perching birds, and a variety of others (Livingston 2000). 40 taxa of mammals have been identified from the cave (Grayson 2002). Most are small mammals such as rodents or rabbits, with large mammals such as bobcats, wolves, bears, deer, pronghorns, and bighorn sheep represented by their small bones, perhaps collected by packrats (Grayson 2000). There is a limited human component as well, including some bones (Shaver 2000). Similar sites include Cathedral Cave, just north of Homestead Cave and with a fossil record extending a few thousand years farther into the past (Madsen et al. 2000), and Hogup Cave, 45 km (28 miles) southwest of GOSP. Hogup Cave has a slightly shorter record (Grayson 1987) but a more extensive human record, including human coprolites with parasite eggs (Fry and Moore 1969).

Park Collections

Aase (undated) did not collect fossils during his survey, but recommended collecting portable fossiliferous specimens. The midden specimens described by Rhode (2000) are housed at the Desert Research Institute in Reno, NV (D. Rhode, pers. comm., August 2011).

Cultural Resource Contexts

GOSP has the potential for fossil-cultural resource associations, but none have been recognized yet. Two broad possibilities for the historic site include fossil materials used for tools and other human-modified objects by prehistoric inhabitants, and the possibility that someone working on the railroad within the historic site commented on fossils. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The recommendations of Thornberry-Ehrlich (2006) should be reviewed in regard to what has been accomplished and what should be implemented. These included: "perform a comprehensive study of paleontological resources at GOSP; compile an inventory of all paleontological specimens present at the site; determine the locations of paleontological specimens removed from the park as part of private collections; and draw visitor attention to the fossil resources at Golden Spike with graphics, brochures, and exhibits."

- Aase (undated) recommended monitoring fossil site GOSP-13, removing portable fossiliferous rocks from the vicinity of trails, and suggested providing interpretation of rocks and fossils at trail locations with non-portable exposed fossiliferous rocks. He noted GOSP-53 as a possible interpretive site in its locality form.

- GOSP reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning one locality, which was reported to be in good condition in 2010. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Historic site staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, historic site staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the historic site and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Data Sets

Golden Spike National Historic Site Paleontological Archives
5/1985–present. (hard copy data; reports; electronic data; photographs; maps; publications). Originated by Santucci, Vincent; status: Active.

Golden Spike National Historic Site Collections
1957–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by GOSP staff; status: Active.

Golden Spike National Historic Site Digital Geologic Map
2007–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=1043034.

Desert Research Institute
1990s–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by DRI staff; status: Active.

Hovenweep National Monument

Hovenweep National Monument (HOVE), Colorado and Utah, was proclaimed March 2, 1923. It was established to protect structures built by the Ancestral Puebloans (also known as the Anasazi), including cliff dwellings, pueblos, and towers (HOVE Figure 1). Its borders were changed April 26, 1951, November 20, 1952, and April 6, 1956. Today HOVE includes six parcels of land near the southern part of the Colorado-Utah border, and encompasses 317.65 ha (784.93 acres), all of which are under federal administration. HOVE is part of the Southeast Utah Group of the National Park Service (NPS), headquartered in Moab, Utah. The Southeast Utah Group also includes Arches National Park (ARCH), Canyonlands National Park (CANY), and Natural Bridges National Monument (NABR); summaries for these parks and monuments are also included in this report.

Geologic Background

HOVE is divided into six administrative units, four of which are located in northwestern Montezuma County, Colorado, and two of which are located in southeastern San Juan County, Utah. From west to east, the four Colorado units are Holly, Horseshoe and Hackberry, Cutthroat Castle, and Goodman Point. All are either adjacent to or within Canyons of the Ancients National Monument, administered by the Bureau of Land Management. The two Utah units are, from west to east, Cajon and Square Tower. The Cajon Unit is located within the Navajo Nation Indian Reservation, while the Square Tower Unit is just north of it. The various HOVE units are southeast of the Abajo Mountains, northwest of the San Juan Basin, and range from west-northwest to north of Sleeping Ute Mountain, depending on the unit. The nearest population centers are Cortez, 14 km (9 miles) southeast of the Goodman Point Unit, and Blanding, 29 km (18 miles) northwest of the Cajon Unit.



HOVE Figure 1. Hovenweep Castle can be found in HOVE's Square Tower Unit (VLS).

Geologic scoping sessions, coordinated by the Geologic Resources Division of the NPS, were held for HOVE and the other Southeast Utah Group parks during May 1999 (NPS 1999), leading to a geologic report for HOVE (Thornberry-Ehrlich 2004). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002). Other references that describe the geology or paleontology of HOVE include Hackman (1955), Santucci (2000), Poole (2001), and Santucci and Kirkland (2010).

The geologic history exposed at HOVE includes rocks from the Upper Jurassic to lower Upper Cretaceous, and unconsolidated Quaternary deposits (see the appendix for a geologic time scale). As the Jurassic drew to a close, North America was moving north out of equatorial latitudes; the Four Corners region reached 22° N during this time (Loope 2008). Aridity decreased through the Late Jurassic (Lockley and Wright 2003). The more humid conditions supported a greater diversity of life and greater biomass, as illustrated by the fossil assemblage of the well-known Morrison Formation. The Morrison Formation is thought to have been a semi-arid seasonal savanna-like setting (Engelmann et al. 2004; Hasiotis 2004; Turner and Peterson 2004).

The Sevier Orogeny (mountain-building event) began during the Jurassic in response to the collision of the North American and Farallon plates along the western margin of North America (Morris et al. 2010), and persisted into the Paleogene (Hintze and Kowallis 2009). During this orogeny, thrust faulting in western Utah formed large mountains in central Utah (Morris et al. 2010). These uplifts provided sediment for the formations deposited during the Cretaceous (Dickinson and Gehrels 2008), and the weight of the mountains depressed the adjacent areas, forming the foreland basin (including HOVE) that would become a long-lived depositional basin (Morris et al. 2010). The Early Cretaceous is represented at HOVE by the Burro Canyon Formation, another fluvial unit (Poole 2001; Thornberry-Ehrlich 2004).

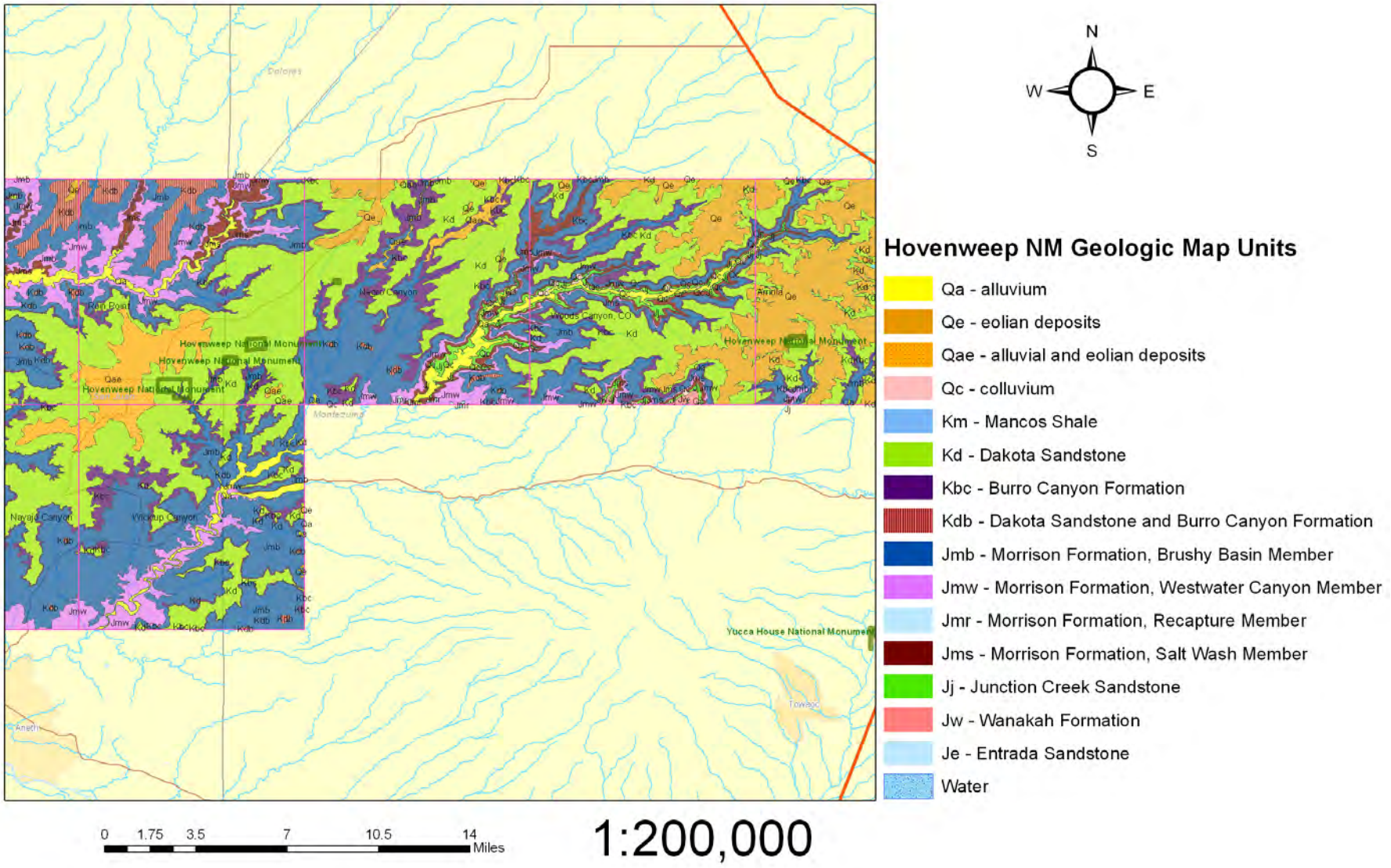
Terrestrial processes continued until about 100 Ma (million years ago), when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Sandstone signals the onset of several cycles when the Western Interior Seaway transgressed (advanced) and regressed (withdrew) over the land (Gustason 1989; Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS

units both within and outside of the Northern Colorado Plateau Network (NCPN). Only the earliest portion of the history of the Western Interior Seaway is represented at HOVE, as the Dakota Sandstone (Poole 2001). Other areas in the vicinity have fuller records of the Late Cretaceous, such as Mesa Verde National Park (MEVE; see the MEVE summary in the Southern Colorado Plateau paleontological resources report for more details), 21 km (13 miles) southeast of Goodman Point.

Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of the Laramide Orogeny (Cather 2003). This mountain building event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). Mushroom-shaped igneous laccoliths intruded during the early part of the Laramide Orogeny, giving rise to some of the area mountain ranges, such as Sleeping Ute Mountain and the La Plata Mountains, with La Plata samples dated to 67 Ma (Semken and McIntosh 1997). Extensive downcutting followed recent uplift. In addition to periods of uplift, the Colorado Plateau underwent local igneous activity during the late Eocene, Oligocene, and Miocene. In the vicinity of HOVE, more than 80 volcanic and intrusive features formed in the Navajo volcanic field of the Four Corners region between 28 and 19 Ma (Semken 2003).

Humans have lived in the HOVE area for over 10,000 years. The Ancestral Puebloans, who built the monument's best-known structures, began to live continuously at HOVE sites around 900 AD. Their population reached a peak of over 2,500 people in the late 1200s, but they departed for the Rio Grande Valley of New Mexico and the Little Colorado River Basin of Arizona shortly thereafter (Hovenweep National Monument undated).

Rock units exposed at HOVE, from oldest to youngest, include the Brushy Basin Member of the Morrison Formation (possibly) (Upper Jurassic), the Burro Canyon Formation (Lower Cretaceous), the Dakota Sandstone (Upper Cretaceous), and Quaternary eolian (wind-blown) deposits and other deposits (Pleistocene–Holocene). The Cajon, Cutthroat Castle, Holly, Horseshoe and Hackberry, and Square Tower units have outcrops of the Burro Canyon Formation and the Dakota Sandstone; outcrops of the Brushy Basin Member may be present in the canyons associated with the Cajon, Horseshoe and Hackberry, and Square Tower units. The Goodman Point Unit has outcrops of the Dakota Sandstone and surficial Quaternary eolian deposits (see HOVE Figure 2 for a geological map) (Poole 2001). Additionally, because all of the units are situated on canyons, all have loose Quaternary deposits from erosion as well. All of the geologic units are known to be fossiliferous, but none of the handful of fossils currently known from HOVE can be positively attributed to any of them at this time (HOVE Table 1).



HOVE Figure 2. Schematic geological map of HOVE. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1043050.

Formation	Age	Fossils Within HOVE	Deposition Environment
Quaternary sediments	Pleistocene–Holocene	None to date; packrat middens and isolated bones of large mammals are most likely	Eolian deposits and valley fill
Dakota Sandstone	Late Cretaceous	None to date; plant fossils, invertebrate fossils, and invertebrate traces are most likely	Terrestrial (especially fluvial) becoming shallow marine over time
Burro Canyon Formation	Early Cretaceous	None to date; fossils known elsewhere include plant fossils, freshwater invertebrates, and freshwater and terrestrial vertebrates	Fluvial, floodplain, and lacustrine settings
Morrison Formation	Late Jurassic	None to date; petrified wood, freshwater invertebrates, and large vertebrate bones are most likely	Fluvial, floodplain, and lacustrine settings

HOVE Table 1. Concise tabulation of HOVE stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of HOVE present opportunities for education, interpretation, and continued or future scientific research in the monument. To date, only three fossils have been reported from HOVE: two mollusk shells found on the surface at Square Tower (V. Webster, Southeast Utah Group museum curator, pers. comm., February 2010), and an unidentified bone found by a Utah Geological Survey geologist (Santucci and Kirkland 2010). In the absence of more detailed provenance information, the monument's stratigraphic units will be discussed under "Potential Sources of Paleontological Resources;" however, it is likely that these three fossils derive from at least one of the stratigraphic units. In particular, because of the presence of large marine mollusks in the upper Dakota Sandstone elsewhere, it may be the source for the mollusk shells.

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within HOVE. However, they are known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from one or more of them.

Morrison Formation (Upper Jurassic)

The Morrison Formation is a heterogeneous and variegated unit dated to the middle Late Jurassic, between about 155 and 148 Ma (Foster 2007). The formation is divisible into several members, although only the Brushy Basin Member is likely to have outcrops at HOVE. The Brushy Basin Member of the HOVE area is between 30 and 110 m (100 and 350 ft) thick, and is composed of gray, pale green, red-brown, and purple bentonitic mudstone (bentonite is altered volcanic ash) with some conglomeratic sandstone beds rich in chert pebbles (Poole 2001; Thornberry-Ehrlich 2004). It was mostly deposited on floodplains, with significant contributions of volcanic ash, and some meandering streams. The upper part of the member in the Four Corners area was deposited in a shallow basin called the T'oo'dichi Basin. Mud was deposited here on an open plain with a high water table. The setting alternated between a wetland and a dry plain (Foster 2007). The climate is thought to have been semi-arid and seasonal, like the African savanna (Engelmann et al. 2004; Hasiotis 2004), with conditions becoming more humid higher in the formation (Parrish et al. 2004).

The Morrison Formation is famous for its fossils, particularly its large dinosaurs. Such familiar genera as *Allosaurus*, "*Brontosaurus*" (*Apatosaurus*), *Brachiosaurus*, *Diplodocus*, and *Stegosaurus* are best known from skeletons found in Morrison Formation rocks, although the best finds have been to the north, east, and west of HOVE in Colorado, Utah, and Wyoming (Turner and Peterson 1999; Foster 2007). The nearest site in the Brushy Basin Member with vertebrate fossils identifiable at the genus level is Scheetz' Quarry 1 in the Uravan area, approximately 115 km (70 miles) north of Goodman Point. Fossils of baby *Dryosaurus*, a type of ornithopod (beaked bipedal herbivorous dinosaur) were recovered from this site, low in the Brushy Basin Member (Turner and Peterson 1999).

The Morrison Formation has one of the best terrestrial fossil records of the Mesozoic. There are several recent enumerations of its fossils: Chure et al. (2006) listed all taxa known to that time, Foster (2003, 2007) detailed the vertebrate assemblages, and Foster (2003) included quarry-level lists of vertebrates. Microbial, fungal, or photosynthetic organisms and traces represented in the Morrison Formation include stromatolites,

fungi (both body and traces), charophyte algae, bryophytes, horsetails, ferns, pteridosperms (seed ferns), cycads, bennettitales, ginkgoes, czekanowskiales, and conifers (Ash and Tidwell 1998; Chure et al. 2006), along with palynomorphs of rhodophytes (red algae), mosses, lycopsids (clubmosses), ferns, cycads, bennettitales, conifers, gnetales, and unknown plants (Dodson et al. 1980; Tschudy et al. 1980, 1981, 1988a, 1988b; Chure et al. 2006), as well as a variety of plant debris, root casts, indeterminate wood, and seeds (Chure et al. 2006). Invertebrates represented by body fossils include sponges, gastropods, unionid bivalves, conchostracans (clam shrimp), ostracodes (seed shrimp), and crayfish (Chure et al. 2006). Many other invertebrates are known from traces, including numerous insect groups not yet known from body fossils. Invertebrate trace fossils have been attributed to anthozoans (corals and anemones), brachiopods (lamp shells), gastropods, bivalves, nematodes, annelids, horseshoe crabs, mayflies, orthopterans (grasshoppers, crickets, and locusts), caddisflies, hemipterans (aphids, cicadas, and allies), flies, beetles, hymenopterans (ants, bees, and wasps), termites, decapod crustaceans, and echinoderms (Hasiotis 2004; Chure et al. 2006); however, some of the traces, such as those attributed to anthozoans, brachiopods, horseshoe crabs, and echinoderms are only known from marine-influenced units that are not found at HOVE (Hasiotis 2004). Vertebrates known from body fossils in the Morrison Formation include several varieties of early ray-finned fish like bowfins, lungfish, frogs, salamanders, turtles, sphenodontians (tuatara), lizards, possible snakes, the lizard-like aquatic reptile *Cteniogenys*, terrestrial crocodylomorphs, *Goniopholis* (closely related to the ancestry of modern crocodilians) and other extinct crocodylomorphs, long-tailed and short-tailed pterosaurs, multiple large (such as *Allosaurus*, *Ceratosaurus*, and *Torvosaurus*) and small (such as *Ornitholestes*) theropods, diverse sauropods (generalized, stocky like *Camarasaurus*, elongate like *Diplodocus*, or long-limbed and long-necked like *Brachiosaurus*), plated dinosaurs like *Stegosaurus*, armored dinosaurs, bipedal herbivorous dinosaurs (such as *Camptosaurus* and *Dryosaurus*), and triconodont, docodont, multituberculate, and symmetrodont mammals (Foster 2003, 2007; Chure et al. 2006). Tracks are known from most of these groups as well, as well as coprolites of herbivorous dinosaurs and mammal burrows (Chure et al. 2006). Finally, eggshells from the Morrison Formation have been attributed to turtles, crocodile relatives, and dinosaurs (Hirsch 1994; Bray and Hirsch 1998; Chure et al. 2006).

Burro Canyon Formation (Lower Cretaceous)

The Burro Canyon Formation of HOVE averages 45 m (150 ft) thick, and is mostly composed of light gray to light brown sandstone and conglomerate. There are also some purplish siltstone, shale, and mudstone beds; thin lenses of gray limestone and chert are present in the upper part of the unit (Poole 2001; Thornberry-Ehrlich 2004). The lowest part of the formation is identified as the Karla Kay Conglomerate Member (Ekren and Houser 1959). The Burro Canyon Formation is interpreted as fluvial at HOVE (Poole 2001; Thornberry-Ehrlich 2004). Sediments were deposited under a semi-arid climate by braided streams, probably reworking material from the Morrison Formation (Walters et al. 1987). The predominant paleoflow was to the north (Craig 1981).

The age of the Burro Canyon Formation is uncertain. It is usually thought of as having been deposited between the Barremian stage of the Early Cretaceous (130 to 125 Ma) and the onset of Dakota Sandstone sedimentation around 100 Ma, although the base of the formation has been reported to intertongue with the Brushy Basin Member of the Morrison Formation in places, which would imply a Late Jurassic age for part of the formation (Ekren and Houser 1959; Dickinson and Gehrels 2008). However, the two units can be difficult to distinguish. Mudstones of the Burro Canyon Formation weather as fragments, while Brushy Basin Member mudstones form a hard frothy crust (Ekren and Houser 1959). The upper contact with the Dakota Sandstone is disconformable (Poole 2001), but the Dakota Sandstone and the Burro Canyon Formation can also be difficult to distinguish; one difference is the relative abundance of plants, which are common in the Dakota Sandstone and rare in the Burro Canyon Formation (Simmons 1957). The Burro Canyon Formation and Cedar Mountain Formation of Utah are lateral equivalents. Generally, the Burro Canyon Formation is present east of the Colorado River, and the Cedar Mountain Formation is present west of it (Dickinson and Gehrels 2008). The two intertongue in places (Dayvault and Hatch 2005). The Cedar Mountain Formation had more volcanic contributions, while the Burro Canyon Formation had higher-energy deposition (Craig 1981).

The fossils of the Burro Canyon Formation are not well known at this time. The current fossil assemblage from the Burro Canyon Formation includes organic debris, lake-dwelling algae, palynomorphs (organic microfossils) including conifer and angiosperm pollen (Tschudy et al. 1984), fern leaves (Simmons 1957), the tree fern *Tempskya* (Dayvault and Hatch 2005), cycadeoids (Tidwell et al. 1976), conifers (Simmons 1957), early flowers (Fishell 1981; Armstrong 2010), plant fragments (Stokes 1952), wood fragments (Tschudy et al. 1984), freshwater bivalves (Reeside 1957), gastropods, freshwater ostracodes, ganoid fish scales (Simmons 1957), and dinosaur bones (Trujillo et al. 2004), including a theropod femur and claw (Armstrong 2010). Additionally, a site in Utah near the border with Colorado (the Hotel Mesa Quarry) has yielded fossils of sharks, bony fish, lungfish, various crocodile relatives, theropods, the sauropod *Brontomerus*, and ornithopods (beaked bipedal dinosaurs) (Taylor et al. 2011). It has been attributed to both the Burro Canyon Formation (Kirkland and Madsen 2007) and the Ruby Ranch Member of the Cedar Mountain Formation (Taylor et al. 2011); by geography, it is within the Burro Canyon Formation (Taylor et al. 2011). The Karla Kay Conglomerate has yielded a root cast and freshwater bivalves (O'Sullivan 1962).

Dakota Sandstone (Upper Cretaceous)

The Dakota Sandstone (also Formation or Group) is a heterogeneous rock unit with outcrops as far from Utah as Minnesota; because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in areas on the western side of the seaway may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). The Dakota Sandstone of HOVE averages 40 m (125 ft) thick, and is composed of yellow-gray to tan carbon-rich sandstone beds in the lower section and yellow-brown to gray sandstone in the upper section. Some dark mudstone and coal beds are also present. The lower section is interpreted as fluvial and the upper section is interpreted as beach sand (Poole 2001; Thornberry-Ehrlich 2004). The Dakota Sandstone of the southern Colorado Plateau is often divided into three informal members, with a conglomeratic base, a heterogeneous coaly middle, and an upper marine sandstone. This division reflects environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma (see Titus et al. 2005 for discussion of the age).

The fossils of the Dakota Sandstone are diverse, befitting a formation including rocks from a variety of depositional settings present over a wide geographic range. The best-known terrestrial fossils from the Dakota Sandstone are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Sandstone (Meek and Hayden 1862); the study of these fossils reached an apex with Lesquereux (1892), a monograph that described 437 species of angiosperms from the Dakota Sandstone of Kansas. Later researchers are not entirely confident of this number, but still recognize the Dakota Sandstone for its wealth of early angiosperms (Landis and Dilcher 2000). Ferns, clubmosses, *Equisetum* (Skog and Dilcher 1994), cycads, conifers, and monocotyledonous angiosperms have also been recognized (Baxter 1954). The floral assemblage of Utah outcrops includes coal (Molenaar and Cobban 1991), palynomorphs (moss and fern spores, and gymnosperm and angiosperm pollen; Nichols 1997), the tree fern *Tempskya* (Ash and Read 1976), *Equisetum*, possible gymnosperms, and angiosperms (Tidwell et al. 1976). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks come from older Dakota Group rocks, and include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Sandstone rocks in Utah (including from Bryce Canyon National Park) reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossil organisms include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Specimens of the tree fern *Tempskya* have been collected from Dakota Sandstone rocks adjacent to MEVE. They are the youngest record of this genus and probably represent a new species (Scott et al. 2001). Mineable coal beds are known in the Dakota Sandstone west of Towaoc, over 23 km (14 miles) south of Goodman Point (Griffitts 2001). Dakota Sandstone marine fossils include algae, foraminifera (another type of "shell"-forming protist; Tibert et al. 2003), dinoflagellate cysts (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Sandstone may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera, corals, bryozoans (moss animals), brachiopods, and echinoderms (Peterson 1969), and reworked petrified wood (Titus et al. 2005).

Quaternary sediments (Pleistocene–Holocene)

The Quaternary deposits of HOVE include areas of eolian loess, silt, and sand that are large enough to be mapped; valley fill and other loose material are also present in the canyons, with a greater range of sediment sizes (Poole 2001; Thornberry-Ehrlich 2004). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Thornberry-Ehrlich (2004) also noted the possibility of finding reworked fossils in Quaternary deposits. Summaries of Quaternary fossil sites in southern Utah, northeastern Arizona, and northwestern New Mexico include Lance (1960), Lindsay and Tessman (1974), Nations and Landye (1984), Agenbroad and Mead (1989), Mead and Agenbroad (1992), O'Neill (1992), Harris (1993, 2008), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), Morgan and Lucas (2003, 2005), Lucas and Morgan (2005), and Coats et al. (2008). To date, fossils are not known in the Quaternary deposits of HOVE.

Paleoecological materials are better known in the vicinity of HOVE than large vertebrate fossils. Two published paleoclimate sites are located within MEVE: Wetherill Mesa in southwest

MEVE (Martin and Byers 1965) and MEVE prehistoric reservoir sites (Wright et al. 2001; Holloway and Wright 2006). Pollen sites within about 100 km (60 miles) of HOVE include: Sagehen Flats in the Dolores River Valley, 17 km (11 miles) east-northeast of Goodman Point (now drowned by the McPhee Reservoir) (Petersen 1985); Beef Pasture, 51 km (32 miles) east-northeast of Goodman Point in the La Plata Mountains (Petersen 1985); and Twin Lakes, 56 km (35 miles) east-northeast of Goodman Point in the La Plata Mountains (Petersen and Mehringer 1976). Packrat (*Neotoma* spp.) middens, collections of plant matter and other material made by packrats and solidified into durable "time capsules" as a side effect of their urine, are present at MEVE. MEVE collections include 29 specimens dated to 100 ± 50 years ago (G. Cox, MEVE archivist, pers. comm., February 2009), suggesting that older middens may also be present. Several localities with packrat middens are known west of HOVE (Anderson et al. 2000), including: Cottonwood Cave, 35 km (22 miles) west of Cajon (Betancourt 1990); Falling Arch, 42 km (26 miles) west-northwest of Cajon (Betancourt 1990); Fishmouth Cave, 44 km (27 miles) west-northwest of Cajon (Betancourt 1984); The Loop, 50 km (31 miles) west-southwest of Cajon (Betancourt 1990); Allen Canyon Cave, 66 km (41 miles) northwest of Square Tower (Betancourt 1984); and White Canyon, within Natural Bridges National Monument (NABR), 81 km (50 miles) northwest of Cajon (Mead et al. 1987). There is the possibility of finding fossils such as packrat middens in alcoves and rock shelters within HOVE. The contact between the Burro Canyon Formation and Dakota Sandstone is riddled with spring and seeps (Thornberry-Ehrlich 2004), which indicate that this horizon could be a good level to inspect. Santucci et al. (2001) presented an overview and cited selected examples of National Park Service fossils found in caves.

Quaternary macrofossils are much less well known. The most significant report of macrofossils in the vicinity of HOVE is a partial skull and vertebral column of a muskox (described as *Symbos cavifrons*, also known as *Bootherium bombifrons*) found in 1982 at Grass Mesa, roughly 20 km (14 miles) northeast of Goodman Point. The site is now drowned by the McPhee Reservoir. The specimen was dated to 15,970 ± 155 radiocarbon years before present (between 19,430–18,830 calendar years before present per Calib 6.0 [http://calib.qub.ac.uk/calib/]; note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years) (McDonald et al. 1987). The most extensive finds are from deposits in the Navajo Indian Irrigation Project of San Juan County, New Mexico, an area over 100 km (60 miles) southeast of Goodman Point. Over 5,400 bones from the last few thousand years have been found at archeological sites there, and belonged to canids, weasels, rodents (including cricetids, geomyids, and sciurids), rabbits, horses, cervids (including deer and elk), and bovids (including antelope, sheep, and cattle) (Lyman 1983).

Park Collections

There are two fossils from HOVE in the collections of the Southeast Utah Group: HOVE 18724 and 18725. Both are mollusk shells collected on the surface (not *in situ*) from the Square Tower Group. They were collected by Franklin Wallace on September 3, 1964. HOVE 18724 is small, and 18725 is large (V. Webster, pers. comm., February 2010).

Cultural Resource Contexts

HOVE is known for its archeology, so it would not be surprising to find paleontological resources associated with or as cultural resources. For example, it is possible that the two shells in HOVE collections were brought in by earlier inhabitants. Tools and other artifacts made of petrified wood and fossiliferous stone are extremely common in the Southwest, and would not be unexpected at HOVE either. Stokes (1972) interpreted three petroglyphs in southeastern Utah as representing proboscideans (possibly mastodons), but such identifications are extremely controversial. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for ARCH, CANY, HOVE, and NABR during May 1999 (NPS 1999), and produced a geologic resource evaluation report for HOVE in 2004 (Thornberry-Ehrlich 2004). The participants of the scoping meetings developed a series of recommendations for future work including continuation of paleontological resource inventories, increased protection, staff training, and the completion of a paleontological resource management plan.

- The geological provenance of the mollusks and bone should be established.

- The monument should consider future field inventories for paleontological resources to more fully document *in situ* occurrences of fossils. The monument may consider a formal site documentation and condition assessment for significant fossil localities. Monitoring of significant sites should be undertaken at least once a year in the future. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Hovenweep National Monument Paleontological Archives
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Hovenweep National Monument Collections
1923–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by HOVE staff; status: Active.

Hovenweep National Monument Digital Geologic Map
2007–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=1043050.

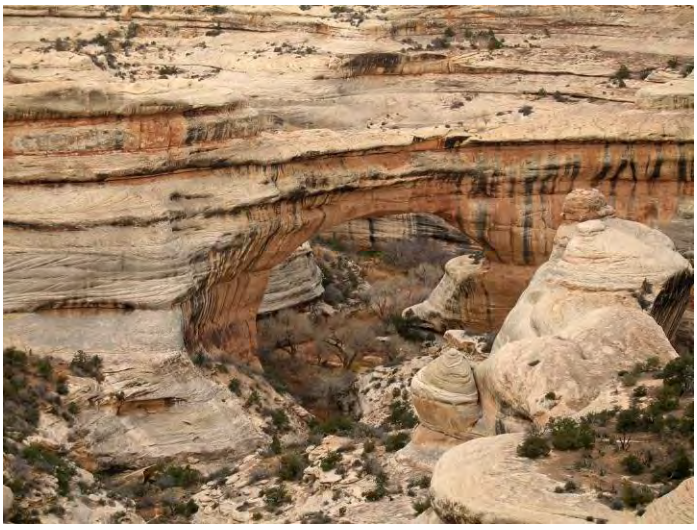
Southeast Utah Group Museum
1923–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by SEUG staff; status: Active.

Natural Bridges National Monument

Natural Bridges National Monument (NABR), Utah, was proclaimed April 16, 1908. It is named for its three natural sandstone bridges, which are among the largest in the world (NABR Figure 1). Ancestral Puebloan rock art and ruins are also present within the monument. The boundaries of NABR changed September 25, 1909, February 11, 1916, and August 14, 1962. The monument currently encompasses 3,090.38 ha (7,636.49 acres), all of which are under federal administration (NABR Figure 2). NABR is part of the Southeast Utah Group of the National Park Service (NPS), headquartered in Moab, Utah. The Southeast Utah Group also includes Arches National Park (ARCH), Canyonlands National Park (CANY), and Hovenweep National Monument (HOVE); summaries for these parks and monuments are also included in this report.

Geologic Background

NABR is located in west-central San Juan County, Utah. It is midway between the two "arms" of Glen Canyon National Recreation Area (GLCA), and is approximately 28 km (17 miles) from GLCA at its nearest. Manti–La Sal National Forest is less than 5 km (3 miles) northeast of NABR, and CANY is about 42 km (26 miles) northeast of NABR at its nearest. NABR is geologically part of the Monument Uplift (or Upwarp) on the Colorado Plateau. The monument is geographically dominated by Cedar Mesa, a plateau to the south, and by White Canyon. White Canyon's main branch runs through western and northern NABR, and its large fork, Anderson Canyon, passes through southern NABR; the two join near Kachina Bridge, in western NABR. The namesake natural bridges (Kachina Bridge, Owachomo Bridge, and Sipapu Bridge [NABR Figure 1]) illustrate three stages of natural bridge formation, and represent former meander bends in the canyons' rock walls that have been cut through by flowing water. These natural bridges were formed in the Cedar Mesa Sandstone (Huntoon et al. 2010).



NABR Figure 1. Sipapu Bridge (VLS).

Geologic scoping sessions, coordinated by the Geologic Resources Division of the NPS, were held for NABR and the other Southeast Utah Group parks during May 1999 (NPS 1999), leading to a geologic report for NABR (Thornberry-Ehrlich 2004). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002). The geology and paleontology of NABR has been described in a number of references, including Wilson and Pierson (1958), Baars et al. (1975a, 1975b), Mead et al. (1987), Agenbroad and Mead (1987), Betancourt (1990), Spaulding (1992), Duffy (1998), NPS (1999), Anderson et al. (2000), Huntoon et al. (2004, 2010), Koch and Santucci (2002), Thornberry-Ehrlich (2004), Mountney (2006), and Santucci and Kirkland (2010).

The geologic record exposed at NABR covers the Early Permian to the Late Triassic, and the Quaternary (see the appendix for a geologic time scale). The late Paleozoic, including the Permian and preceding Pennsylvanian, was a time of tectonic and structural activity in the American West (Hintze and Kowallis 2009; Huntoon et al. 2010). The stresses led to the appearance of several features that influenced deposition at NABR: the Ancestral Rockies, including the Uncompahgre Uplift east and northeast of NABR; the Monument Uplift, at the time a broad low upwarp; and the Paradox Basin between the two uplifted features. The oldest rocks exposed at NABR (the Cedar Mesa Sandstone) were deposited during this period of activity, shortly after 300 Ma (million years ago) (Huntoon et al. 2010). The area including NABR was equatorial at this time (Huntoon et al. 2002). Depositional settings changed from extensive dune seas or ergs (giant dune fields) to rivers and floodplains. The sedimentary record of the monument then paused due to a period of erosion or nondeposition (Huntoon et al. 2010), leaving a hiatus of approximately 30 million years at the monument.

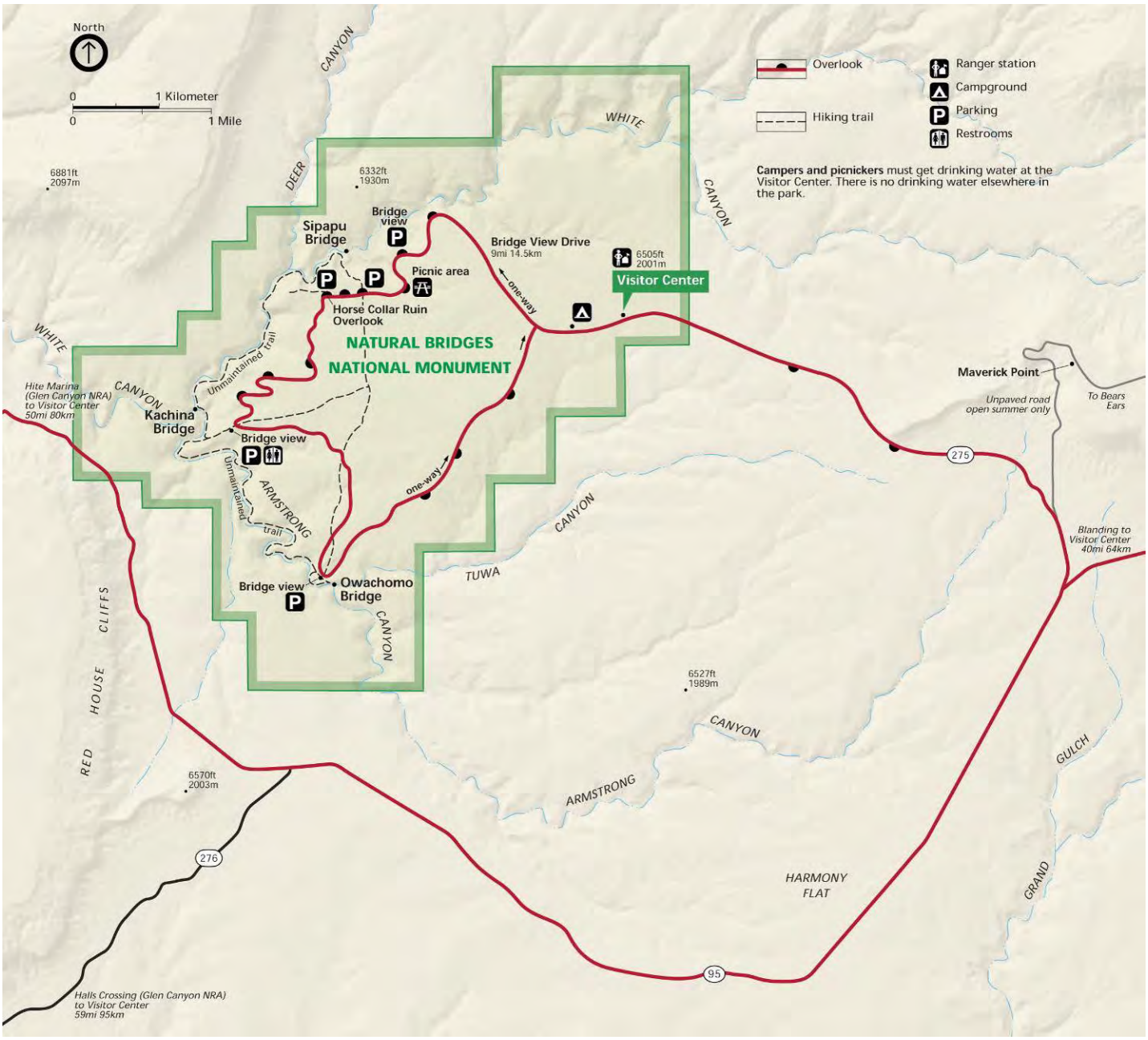
Triassic rocks at NABR are principally represented by the Moenkopi Formation of the Early–Middle Triassic, with limited outcrops of the Chinle Formation of the Late Triassic. Southern Utah during the Early and Middle Triassic was a broad coastal plain that was occasionally flooded by a sea to the west (Stewart et al. 1972a; Blakey et al. 1993; Hintze and Kowallis 2009). The top of the Moenkopi Formation is separated from the Chinle Formation by millions of years, with intervening erosion (Pipiringos and O'Sullivan 1978). The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time (Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed

substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003). Chinle Formation rocks show that the area gradually became drier over time as the supercontinent Pangaea moved north. During its deposition, the Four Corners area of the Southwest moved from near the equator to arid-zone latitudes. What had been a setting of rivers, lakes, and deltas turned arid (Tanner 2003). These changes are not seen at NABR, but are well-developed in the rocks of nearby CANY and GLCA.

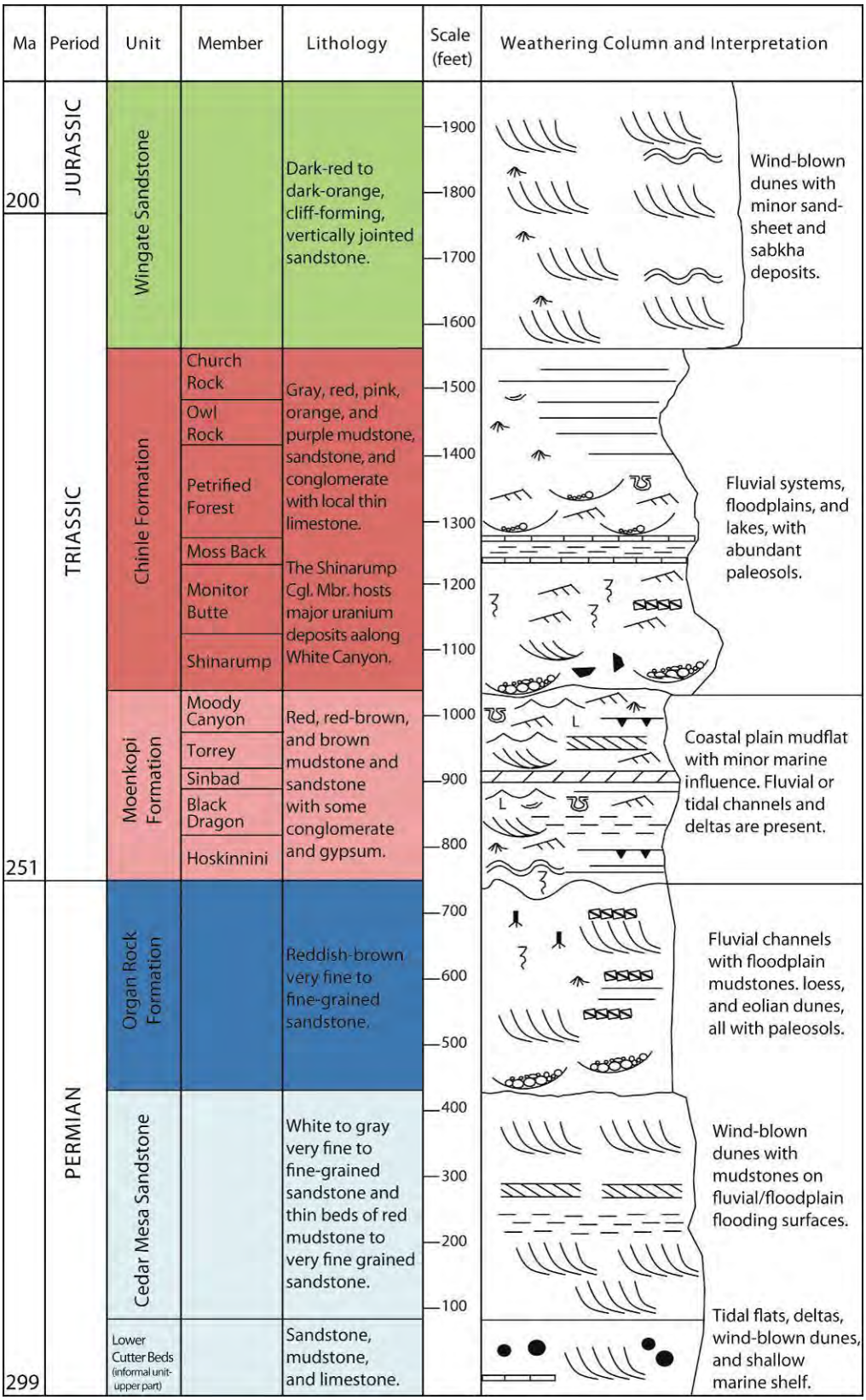
Deposition from the Permian to the Cretaceous left thousands of meters or feet of sedimentary rocks at what would become NABR, but most of the rocks were eroded to expose the Cedar Mesa Sandstone that makes up much of the monument's surface area. The erosion of these overlying rocks began with the Late Cretaceous–Eocene Laramide Orogeny (mountain-building event) (Huntoon et al. 2010). The Circle Cliffs uplift and the Kaiparowits and Henry Mountains structural basins are examples of local Laramide structures (Anderson et al. 2010), and the Monument Uplift was reactivated at this time as well (Huntoon et al. 2010). In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent local igneous activity during the late Eocene, Oligocene, and Miocene. Igneous rocks emplaced at depth formed the nearby Abajo, Henry, and La Sal mountain ranges between about 31 and 23 Ma (Nelson et al. 1992; Nelson 1998; Nelson and Davidson 1998), and between 28 and 19 Ma more than 80 volcanic and intrusive features formed in the Navajo volcanic field to the southeast in the Four Corners area (Semken 2003). Downcutting and erosion followed the uplifts. An episode of rapid downcutting began after 6 Ma, when the Colorado River began cutting through the Grand Canyon (Luchitta 1989; Pederson 2008) and in turn lowered base levels for its drainage system, including in the NABR area. The namesake natural bridges of NABR were probably mostly carved during a period of greater precipitation than the present, perhaps between approximately 30,000 and 12,000 years ago (Huntoon et al. 2010).

The long post-Triassic depositional hiatus at NABR ended during the Pleistocene, preserving evidence of the biota that lived there during the Ice Age. The megafaunal extinctions on the Colorado Plateau claimed such animals as the mammoth, the ground sloth *Nothrotheriops shastensis*, the shrub ox *Euceratherium collinum*, and the mountain goat *Oreamnos harringtoni*. Dated dung deposits from the Colorado Plateau indicate that they went extinct between 11,800 and 11,000 radiocarbon years ago (Mead and Agenbroad 1992). Note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years. Dates here were calibrated using Calib 6.0 (http://calib.qub.ac.uk/calib/) where appropriate. The Colorado Plateau was cooler and more moist during the last glacial period, becoming more arid afterward. It has been most arid during the past 1,500 years (Elias et al. 1992). It had been thought that both megafauna and human cultural sites more than 6,000 years old were rare on the Colorado Plateau, but this is now known to be inaccurate (Agenbroad 1997). The human record at NABR includes at least the past 1,000 years, with the Ancestral Puebloans (also known as the Anasazi) (Huntoon et al. 2010).

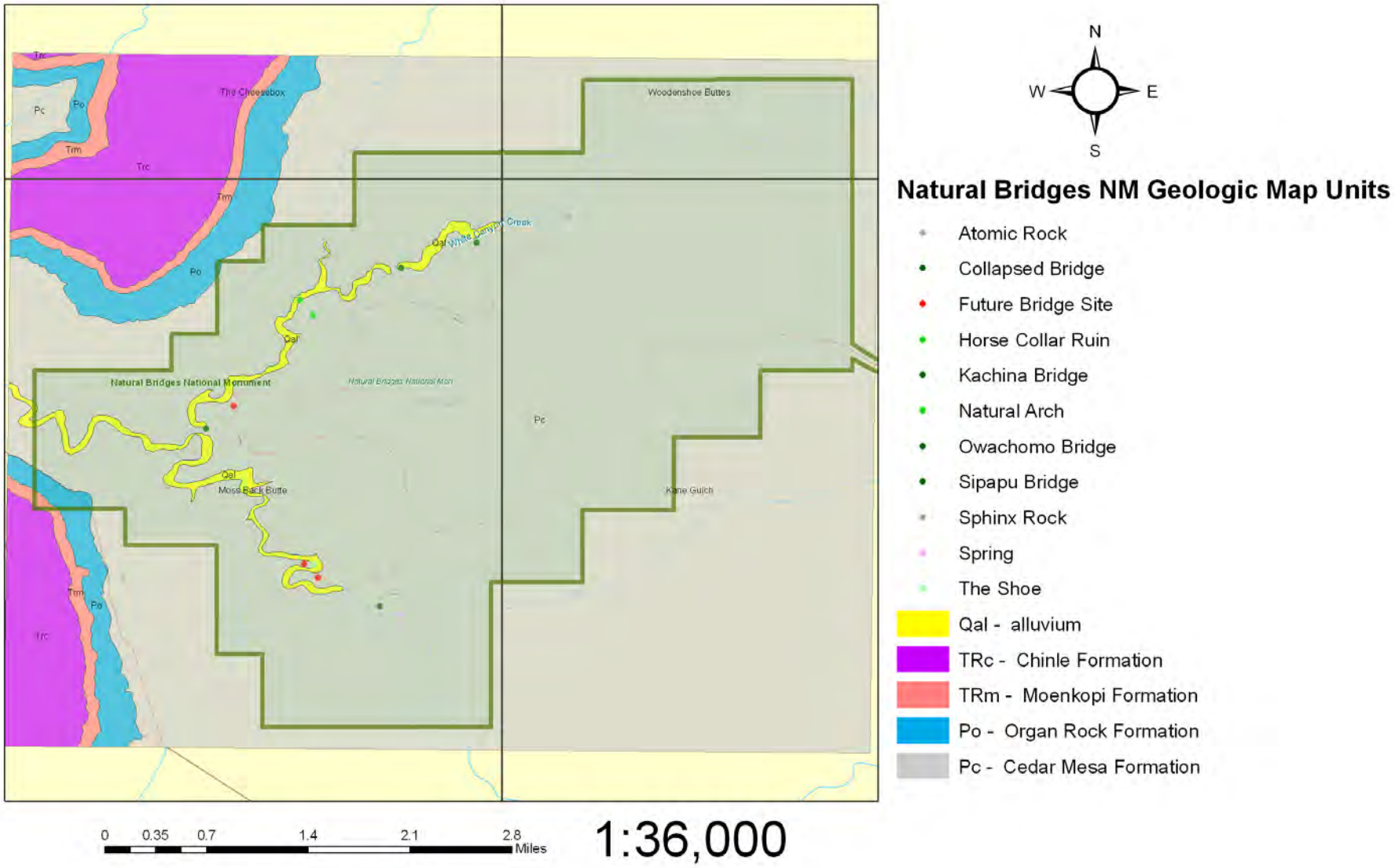
Rock units exposed at NABR include, from oldest to youngest: the Cedar Mesa Sandstone and Organ Rock Formation of the Cutler Group (Lower Permian); the Moenkopi Formation (Lower–Middle Triassic); the Chinle Formation (Upper Triassic); and a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene). The Cedar Mesa Sandstone makes up the vast majority of the monument's outcrops, with progressively smaller exposures of the Organ Rock Formation, Moenkopi Formation, and Chinle Formation on the rising elevation of the southern corner of NABR's western border. Quaternary deposits are most extensive in Armstrong and White canyons and their tributary canyons (see NABR Figure 3 for a stratigraphic column and NABR Figure 4 for a geological map) (Huntoon et al. 2010). Paleontological resources at NABR are known from the Cedar Mesa Sandstone (root casts and other terrestrial trace fossils), the Chinle Formation (reworked fragments of petrified wood), and Quaternary deposits (plant matter, pollen, bones, dung, and packrat middens) (NABR Table 1). Additionally, fossil collections from NABR are maintained by several institutions.



NABR Figure 2. Geography of NABR (NPS).



NABR Figure 3. Stratigraphic column for NABR (from Huntoon et al. [2010], used with permission of the Utah Geological Association).



NABR Figure 4. Schematic geological map of NABR. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1038959.

Formation	Age	Fossils Within NABR	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Extinct mountain goat bones and dung, and packrat middens	Alluvial and fluvial deposits
Chinle Formation	Late Triassic	Reworked petrified wood	Fluvial, floodplain, lacustrine, and eolian settings, becoming drier over time
Moenkopi Formation	Early–Middle Triassic	None to date; invertebrates, invertebrate traces, and vertebrate traces are most likely	Coastal plain
Organ Rock Formation	Early Permian	None to date; root traces and invertebrate traces are most likely	Fluvial and floodplain settings
Cedar Mesa Sandstone	Early Permian	Root casts, invertebrate burrows	Coastal eolian sand dunes with occasional marine incursions

NABR Table 1. Concise tabulation of NABR stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of NABR present opportunities for education, interpretation, and continued or future scientific research in the monument. There is a recent report of possible tracks in a canyon, currently unverified (C. Hayes, HOVE/NABR Superintendent, pers. comm., November 2009); given the geology of the monument, these would most likely be in the Cedar Mesa Sandstone.

Cutler Group: Cedar Mesa Sandstone (Lower Permian)

The Cedar Mesa Sandstone is a coastal sandstone unit with both marine and eolian (wind-blown) influences, near a fluctuating shoreline (Baars 2010), although some older references interpreted it as more subaqueous (Kirkland 1963; Chamberlain and Baer 1973; Vaughn 1973). On the nearby Cedar Mesa, this unit is approximately 130 m (425 ft) thick. At NABR, the Cedar Mesa Sandstone is the rock unit that makes up the namesake natural bridges, and is composed of white sandstone with thin interbeds of red mudstone sandwiched by red sandstone. The white sandstone is interpreted as large eolian dunes, with some interdune and sand sheet deposits. The red mudstone and sandstone deposits are horizontal beds up to 3 m (10 ft) thick that probably formed during flooding events that deposited continuous surfaces. Isolated mudstone deposits may represent ponds. Gypsum and limestone have been reported from the Cedar Mesa Sandstone to the south, interpreted as a salt flat (sabkha) on a marine shore, but only lenses of limestone are present at NABR, found in the red mudstone layers. The formation grades out of underlying "lower Cutler beds" and grades into the overlying Organ Rock Formation in the NABR area (Huntoon et al. 2010). To the east, it transitions into undivided Cutler Formation rocks, similar to the lower Cutler beds. Fossiliferous limestones are also included in this transition (Billingsley et al. 2002).

The Cedar Mesa Sandstone was deposited during the Early Permian (Wolfcampian stage) (Blakey and Knepp 1989; Condon 1997), sometime between approximately 299 and 280 Ma. Multiple episodes of ergs are separated by erosional surfaces thought to be from marine transgressions (Baars 2010). Sand was blown by strong north-northwest winds (Parrish and Peterson 1998; Peterson 1988; Condon 1997). The climate was arid; NABR is thought to have been near the center of the Cedar Mesa Sandstone's erg (Huntoon et al. 2010).

Fossils are not common in the Cedar Mesa Sandstone, but there are fossils in the unit at NABR. One outcrop at the monument, discovered in 1997, preserves large root casts over 3 m (10 ft) long, averaging 2.5 to 5 cm (1 to 2 in) thick. They are interpreted as an example of an "island" of vegetation among terrestrial dunes, similar to modern coastal sabkhas. Associated with these are smaller root traces that are likely modern root calcifications (Duffy 1998). Huntoon et al. (2010) noted the presence of a red mudstone on the Sipapu Bridge Trail with root casts and burrows. Vertical and horizontal burrows are present in some of the monument's interdunal sandstones, albeit poorly preserved (Santucci and Kirkland 2010). Photographs of probable tree roots in red mudstone, trace fossils in sandstone, and trace fossils in red silty sandstone from the Sipapu Bridge Trail are provided in Huntoon et al. (2000a).

A fair number of fossils have been reported from Cedar Mesa Sandstone outcrops in the vicinity of CANY and GLCA. At CANY, there are foraminifera (amoeba-like protists that form "shells") and rare crinoids (sea lilies) in the sandstone (Baars 2010), and fossiliferous limestone is present in the transition with the undivided Cutler Group (Billingsley et al. 2002). Langford et al. (2007) found bryozoans (moss animals), brachiopods (lamp shells), and crinoids from marine

transgressive surfaces. Several track sites are known from CANY as well (Loope 1984; Santucci and Kirkland 2010). Just east of CANY, at Indian Creek, a few vertebrate fossils are known from the Cedar Mesa Sandstone or Cedar Mesa Sandstone/undivided Cutler Group transitional zone in one or two sites. Sumida et al. (1999a) described a site in the Cedar Mesa Sandstone that, when first located, included a "log jam" of petrified conifer wood, as well as bones of the temnospondyl amphibian *Eryops* and pelycosaur ("mammal-like reptile") *Sphenacodon*, but the site was later vandalized (Sumida et al. 1999a). Stanesco et al. (2002) later published a brief description of fossils from Indian Creek, in the Cedar Mesa Sandstone–undivided Cutler Group transitional zone. These fossils include root casts and conifer logs in a paleo-arroyo cut into eolian rocks, and disseminated plants, vertebrate bone fragments, and teeth from an overlying shallow lake limestone (Stanesco et al. 2002). Three published track sites have been found in the Cedar Mesa Sandstone at adjoining GLCA, with two more sites just outside of its boundaries. Synapsids (the group including pelycosaurs, other mammal relatives, and true mammals) were the primary track makers. Lizard-like tracks are also known from one site (Lockley et al. 1998). Other fossils known from the Cedar Mesa Sandstone elsewhere include ghost shrimp burrows and other invertebrate trace fossils from near the top of the unit (Chamberlain and Baer 1973), microbial structures and rare marine fossils in the limestone and gypsum beds, rare sand-sized marine fossil fragments in the sandstone layers (Huntoon et al. 2010), and rare casts of marine invertebrates (Kirkland 1963).

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location (Stewart et al. 1972b) and investigator. These differences have limited relevance for NABR, because Chinle Formation outcrops at the monument are limited to a small portion of the lower part of the formation in the southern corner of the monument's western boundary (Huntoon et al. 2010). In the White Canyon area, O'Sullivan and MacLachlan (1975) identified the members of the Chinle Formation, from oldest to youngest, as the Shinarump Conglomerate, a mudstone member, the Moss Back Member, a claystone member, a limy member, a siltstone member, and the Hite Bed, whereas Huntoon et al. (2010) used the Shinarump Conglomerate, Moss Back, Monitor Butte, Petrified Forest, Owl Rock, and Church Rock Members. For the purposes of this document, the Chinle Formation is a colorful unit primarily composed of mudstone, sandstone, conglomerate, and thin limestone, deposited in streams, lakes, and on floodplains (Huntoon et al. 2010). More detailed information on the members of the Chinle Formation in southeastern Utah can be found in the CANY summary of this report.

The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) because this division is known to be in the overlying Wingate Sandstone (Lucas et al. 1997, 2006). The Moenkopi Formation–Chinle Formation disconformity is marked by paleovalleys cut into the Moenkopi Formation and filled by the Shinarump Conglomerate (Huntoon et al. 2010). As before, drainage moved northwest to a shoreline, at this time in western Utah. The climate is interpreted as monsoonal, with pronounced wet and dry seasons (Huntoon et al. 2010). Toward the top of the Chinle Formation, the climate became much drier as Pangaea drifted north and moved the area out of favorable latitudes (Tanner 2003).

The various members of the Chinle Formation have similar types of fossils, but differ in the genera and species present and which are most commonly found. The Chinle Formation flora includes algae, pollen, spores, leaves of ferns, cycads, and conifers, and petrified wood. Chinle Formation invertebrates include bivalves, gastropods, ostracodes, conchostracans (clam shrimp), crayfish, and trace fossils. The vertebrate fossil assemblage of the Chinle Formation includes freshwater sharks, early bony fish, lungfish, coelacanths, metoposaur amphibians, non-mammalian therapsids (another type of "mammal-like reptiles"), sphenodontians (relatives of the modern tuatara), procolophonids and trilophosaurs (extinct early reptiles), aetosaurs and revueltosaurs (armored herbivorous crocodile relatives), crocodile-like phytosaurs, shuvosaurids (toothless dinosaur-like crocodile relatives), "rauisuchians" (large dinosaur-like terrestrial carnivores related to crocodiles), sphenosuchians (small dinosaur-like crocodile relatives), early dinosaurs and dinosaur relatives, footprints, and coprolites, with lungfish, metoposaurs, aetosaurs, and phytosaurs most common (Stewart et al. 1972b; Parrish 1989; Lucas 1993; Heckert et al. 2005). At present, the Chinle Formation is known to have contributed petrified wood to the loose sediment in Armstrong Canyon and White Canyon at NABR (NPS undated). Because of the very small outcrop area, it is unlikely that the Chinle Formation will have a great deal of *in situ* fossils in the monument. The most likely potential *in situ* fossils are pieces of petrified wood, which may also be somewhat radioactive because uranium concentrates in Chinle Formation fossil wood (Johnson and Thoradson 1966; Baars et al. 1975b). Plant fossils have been found at a number of locations in the lower Chinle Formation of the White Canyon and its vicinity, in part because of this property (Finnell et al. 1963; Thaden et al. 1964; Johnson and Thoradson 1966; Baars et al. 1975b; Dubiel 1982, 1983a, 1983b).

Quaternary sediments (Pleistocene–Holocene)

Quaternary sedimentary deposits at NABR are mostly made up of rock fragments of clay to gravel sizes associated with flowing water. These recent sediments have received little study (Thornberry-Ehrlich 2004). They are best developed along White and Armstrong canyons, and their larger tributary canyons (Huntoon et al. 2010). Typical Quaternary fossils of the

Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). NABR has both types, and there are numerous sites with mammoth, bison, shrub ox, musk ox, sloth, mountain goat, camel, or tapir bones, or rock art of mammoth or bison within a 100 km (60 mile) radius of the monument (Agenbroad 1997; Elias 1997), as well as paleoecological sites (Anderson et al. 2000). Summaries of Quaternary fossil sites in southern Utah and the vicinity include Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008).

To date, the Quaternary fossil record of NABR derives from one location, a large flat-floored rock shelter in a Cedar Mesa Sandstone canyon wall. This site was described in detail by Mead et al. (1987). It has yielded bones and dung of the extinct mountain goat *Oreamnos harringtoni* and numerous packrat middens of various ages with plant fossils and rat droppings. *Oreamnos* bones are represented by two metapodials. *Oreamnos* dung, with its distinctive cuboid to rectangular shape and unusually great weight, is more abundant. Analysis of the dung showed that the goats ingested birch (perhaps in spring), bark, limber pine (*Pinus flexilis*), Douglas fir (*Pseudotsuga menziesii*), grass, and sedge, with the pine, fir, bark, and grasses as the most abundant components. Packrat middens, collections of plant material and food waste constructed by packrats (*Neotoma* spp.) and cemented by their viscous urine, are found both strongly hardened and unconsolidated. Skeletal remains are also mentioned as being present in the NABR middens, but are undescribed. At least two layers rich in *Oreamnos* dung and three primary layers of packrat middens are present in the rock shelter, in a roughly 100 cm (40 in) thick stratigraphic column that took over 30,000 years to form. The oldest dated material in the column is *Oreamnos* dung, produced more than 39,800 radiocarbon years before present; this represents the oldest dated record of *O. harringtoni*.

The youngest date at the NABR rock shelter is on a packrat midden, dated to 9,660 ± 160 radiocarbon years before present (11,405–10,510 calendar years before present). Plant macrofossils include several species that no longer grow nearby, including Engelmann spruce (*Picea engelmannii*), limber pine, and Douglas fir (Mead et al. 1987). NABR's rock shelter fossils are also discussed in several other documents (Agenbroad and Mead 1987; Betancourt 1990; Mead and Agenbroad 1992; Spaulding 1992; Mead et al. 1993; Mead and Lawler 1995; Anderson et al. 2000; Santucci et al. 2001; Santucci and Kirkland 2010; the online U.S. Geological Survey/National Oceanic and Atmospheric Administration Packrat Midden Database [http://esp.cr.usgs.gov/data/midden/]), but little new information has been published since the 1987 description.

There is the possibility of finding additional alcove and rock shelter fossils within NABR, particularly packrat middens. Santucci et al. (2001) presented an overview and cited selected examples of National Park Service fossils found in caves, including at NABR, ARCH, and GLCA. Quaternary specimens from NABR are at Northern Arizona University (NAU; Flagstaff) and the Museum of Northern Arizona (MNA; Flagstaff), per the NABR collections database (V. Webster, Southeast Utah Group museum curator, pers. comm., November 2009) (see "Park Collections" below). As noted in the Chinle Formation section, there are also fragments of Chinle Formation petrified wood in Armstrong Canyon and White Canyon (NPS undated).

The most diverse fossil sites within 100 km (60 miles) of NABR are Bechan Cave within GLCA, and Cowboy Cave, over 80 km (50 miles) north of NABR. Bechan Cave is the best-known and most diverse of several Quaternary alcove and rock shelter sites at GLCA. It is located near the confluence of the Colorado River and Escalante River. Containing 300 cubic meters (10,600 cubic feet) of fossil dung among other fossil resources, Bechan Cave has been described in a number of publications. Its fossils include: fungal spores in droppings (Davis 1987; Davis and Shafer 2006); pollen (Davis 1990); charcoal, wood (Agenbroad and Mead 1989), seeds and plant fragments (Davis et al. 1984); insects (Elias et al. 1992); dung assigned to packrats, cottontail rabbits, Shasta ground sloths (*Nothrotheriops shastensis*), mammoths, possibly horses, the extinct shrub ox *Euceratherium collinum*, bighorn sheep, and extinct mountain goats; hair from shrews, packrats and other rodents, bats, coyotes, bear, ground and mylodont sloths (*Glossotherium* and *Nothrotheriops*), mammoths, horses, bison, bighorn sheep, and deer; bones from toads, snakes, grouse-sized birds, a variety of rodents, rabbits, and *Euceratherium*; and packrat middens covering the last 12,000 years (Agenbroad and Mead 1987, 1990; Mead and Agenbroad 1992; Santucci et al. 2001; Santucci and Kirkland 2010). Human artifacts are also known from younger levels of the cave (Agenbroad et al. 1989). Fourteen plant samples and seven dung samples from mammoths and shrub oxen have been dated from Bechan Cave, yielding a range that spans from 13,505 ± 580 radiocarbon years before present (17,835–14,460 calendar years before present) to 11,630 ± 150 radiocarbon years before present (13,800–13,215 calendar years before present) (Mead and Agenbroad 1992).

Cowboy Cave and its contents were extensively described by Jennings (1980). It is notable for having both extensive Pleistocene material and four distinct Holocene cultural layers above, spanning the last 13,000 years at least (humans present only since approximately 9,000 years ago). In the Pleistocene layer, various types of mammal dung and hair were found with packrat middens and a handful of bones, mostly bison, but also two juvenile mammoth tusks and a possible elk bone (Barnett and Coulam 1980; Hansen 1980; Jennings 1980; Spaulding and Peterson 1980). The cultural levels show human use of stone, bones, hair, hides, shells, plant parts, and other local organic materials to make tools, ornaments, bags, and other artifacts (Hewitt 1980; Hull 1980a, 1980b; Janetski 1980; Lucius 1980;

Weder 1980). Prehistoric dung at the cave mostly came from bison, but also includes feces from Shasta ground sloths, elephants, horses, elk, elk or camel (Mead and Agenbroad 1992), rabbits, rodents (Hansen 1980), and humans (Hogan 1980).

In addition to these fossils, invertebrates may also be found. Holocene-age gastropods and ostracodes (seed shrimp) are known from Cedar Mesa, southeast of NABR (Salkin 1975).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within NABR. However, they are known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from one or more of them.

Cutler Group: Organ Rock Formation (Lower Permian)

The Organ Rock Formation (or Shale or Tongue) is a red bed sandstone and shale unit. It is only about 90 m (300 ft) thick at NABR, thickening farther from the Monument Uplift (Huntoon et al. 2010), and appears banded because of the alternation of resistant and nonresistant beds (Thornberry-Ehrlich 2004). It is primarily composed of sandstone in the area. The lower contact is transitional with the Cedar Mesa Sandstone. In some areas it is capped by the White Rim Sandstone, but this unit was apparently eroded from the NABR area during the later Permian or Early Triassic. Instead, the Organ Rock Formation is disconformably overlain by the Moenkopi Formation at NABR and its immediate vicinity (Huntoon et al. 2010). Like the Cedar Mesa Sandstone, the Organ Rock Formation grades into undivided Cutler Formation rocks to the east (Baars 2010). It is thought to be equivalent to the Hermit Formation of the Grand Canyon area (Blakey and Knepp 1989). The only large exposures of the unit at NABR are near the western border's southern end (Huntoon et al. 2010).

Deposition of the Organ Rock sediments occurred during the Early Permian (Leonardian stage), sometime between approximately 280 and 271 Ma (Blakey and Knepp 1989; Stanesco et al. 2000). Plant fossils indicate that the climate was semi-arid with long growing seasons (Vaughn 1964); precipitation was greater than in the environments of the Cedar Mesa Sandstone (Huntoon et al. 2010). Its depositional environments ranged from marginal marine to continental lowlands (Dubiel et al. 1989; Huntoon et al. 2010). Rocks from fluvial, floodplain, eolian, marginal marine mudflat, and tidal flat depositional environments are known, with fluvial and floodplain deposits common at NABR. Paleosols (ancient soil horizons) are found throughout this formation (Huntoon et al. 2010).

Fossils are not yet known from the Organ Rock Formation at NABR, although root casts are known just west of the monument (Thornberry-Ehrlich 2004). Root casts and burrows are perhaps the most likely fossils to be found in this unit at the monument, as they are known particularly from floodplain deposits (Huntoon et al. 2010). Vertebrates and plants have also been found in the Organ Rock Formation, although again not yet at NABR. The small vertebrate assemblage is best known from the Monument Valley area. Fossils are largely restricted to stream channel deposits found about 11 m (35 ft) above the base of the unit. Fossiliferous deposits are typically in conglomerate or associated sandstone, or sandstone lenses. These fossiliferous locations stand out from surrounding red bed mudstones, shales, and sandstones (Vaughn 1964). The vertebrates include sharks, palaeoniscid fish, at least three types of labyrinthodont amphibians including *Eryops*, the reptiliomorphs (early reptile-like tetrapods) *Seymouria sanjuanensis*, *Diadectes*, and *Tseajaia campi*, and the pelycosaurs *Ophiacodon*, *Ctenospondylus*, *Dimetrodon*, and *Sphenacodon* (Vaughn 1973; Sumida et al. 1999a, 1999b, 1999c). Fish are rare; scales are only known from coprolites (fossil feces). Additionally, tetrapod tracks are also known, from a greater stratigraphic range than body fossils (Vaughn 1964). Plant fossils include specimens of *Callipteris*, *Protoblechnum*, *Supaia*, *Taeniopteris*, *Walchia*, and *Yakia*, representing pteridosperms (seed ferns) and possible ferns and conifers (Vaughn 1964; Mamay and Breed 1970). These are similar to the floral assemblage of the Hermit Formation (Mamay and Breed 1970).

Moenkopi Formation (Lower–Middle Triassic)

The Moenkopi Formation is a heterogeneous unit present in several western states. At NABR, its exposures are essentially limited to the southern corner of the western border. It is 90 to 120 m (300 to 400 ft) thick on White Canyon outside of the monument, forming a slope between the Organ Rock Formation and Chinle Formation. Like the Organ Rock Formation, it is a red bed unit, and the two are hard to distinguish if there are no intervening beds; the Moenkopi Formation has somewhat coarser beds. Despite the similarities, there is a disconformity representing some tens of millions of years of missing time between the two. The Moenkopi Formation's upper contact with the Chinle Formation is also disconformable, with a gap in time of similar length. The rocks of the Moenkopi Formation in the NABR area are predominantly mudstones and sandstones, red to brown in color, with localized conglomerates and gypsum (Huntoon et al. 2010).

The Moenkopi Formation dates to the Early and early Middle Triassic (Heckert et al. 2005), between approximately 250 and 240 Ma. Its rocks formed in diverse settings of a coastal plain mudflat, with streams, deltas, tidal channels, salt flats, and

floodplains represented. The area drained to a sea to the west and northwest, under a climate interpreted as warm and monsoonal (Huntoon et al. 2010). It has been divided into several members, depending on location (Stewart et al. 1972a). In southeastern Utah, it was only divided relatively recently, aside from the Sinbad Limestone (Blakey 1974). These members include, from oldest to youngest, the Hoskinnini, Black Dragon, Sinbad, Torrey, and Moody Canyon members (Huntoon et al. 2010). The Black Dragon Member and Sinbad Limestone are interpreted as coastal and marine units, the Torrey Member is interpreted as a large delta, and the Moody Canyon Member is interpreted as a restricted shallow water to sabkha setting (Blakey 1974). Because the Sinbad Limestone disappears before reaching the NABR area (O'Sullivan and MacLachlan 1975), the Black Dragon and Torrey members are essentially indistinguishable (Blakey 1974). Because of this, and because of the geographically limited exposures of Moenkopi Formation outcrops at NABR, the members will only be described briefly. Additional information on these members of the Moenkopi Formation can be found in the CANY or Capitol Reef National Park (CARE) summaries of this document. Fossils are not yet known from the Moenkopi Formation at NABR, but its members are known to be fossiliferous elsewhere.

The Hoskinnini Member is a cliff-forming red bed unit (Finnell et al. 1963), composed of silty sandstone and sandy siltstone. It was transferred from the underlying Cutler Group (Stewart 1959), and is occasionally classified as a formation on its own, with the age mooted as possibly at least partially Permian (Blakey 1974; Baars 2010). The Hoskinnini Member lacks diagnostic fossils (Finnell et al. 1963) but does have stromatolites (layered sedimentary structures formed by microorganisms) (J. Kirkland, pers. comm., July 2009) and bioturbation (erasing of sedimentary structures by extensive organism activity) (Huntoon and Dubiel 2000).

The Black Dragon Member is a heterogeneous unit of mudstone to sandstone, limestone, dolomite, and gypsum (Blakey 1974). It is slightly younger than the Hoskinnini Member but occupies a similar stratigraphic position, and grades into the Hoskinnini Member in the White Canyon area (Huntoon et al. 1994). Fossils are not currently well-known in the Black Dragon Member. Some body fossils are known, such as bivalves, gastropods, a possible ammonite, ostracodes, and a conodont (extinct eel-like chordate) fragment. These body fossils are best known from the San Rafael Swell (Blakey 1974). Trace fossils are known as well (Blakey 1974).

The Torrey Member is mostly composed of sandstone and siltstone, but many other rock types are present in minor quantities. Often the base is siltstone, followed by ledgey sandstone in the middle, and then an upper sandstone/siltstone section (Blakey 1974). It grades into the overlying Moody Canyon Member (Stewart et al. 1972a). Fossils from the Torrey Member include horsetails (*Equisetum*), brachiopods, bivalves, ostracodes, fish scales (Blakey 1974), fish bones (Mickelson et al. 2006), amphibian bones (Blakey 1974), a skull of the labyrinthodont amphibian *Parotosuchus* (from Hite Crossing) (Lucas and Schoch 2002), invertebrate traces (including marks thought to come from millipedes and horseshoe crabs), fish swim traces, and tetrapod tracks (Mickelson et al. 2006). Tracks have been found as sandstone casts in mudstone (Mickelson et al. 2006).

The Moody Canyon Member is a heterogeneous unit with red to brown siltstones and mudstones, sandstones, and an upper section of dolomite and gypsum. It is notable for its continuous regular beds (Blakey 1974). Fossils are rare in this unit, with only reptile tracks (*Chirotherium*) from an outcrop in the Orange Cliffs (Blakey 1974) reported to date.

Additionally, there are also a few records of Moenkopi Formation fossils in southeastern Utah where the member is not specified. Burrows are found near the base of the formation on Cedar Mesa (Huntoon et al. 2010). Fish scales and a possible coelacanth fragment were reported from the upper part of the formation in the Deer Flat area of White Canyon (Finnell et al. 1963). Trace fossils such as burrows are known from CANY and the vicinity (Baars 2010), such as burrows and trails on the low broad tidal flats of the Moenkopi Formation between Moab and Hite (Rigby et al. 1971). Carbonaceous debris was reported in the Moenkopi Formation at Taylor Canyon (Stewart et al. 1972b). Finally, track sites have been found in Moenkopi Formation rocks at GLCA (Mickelson et al. 2006). Types of traces from these sites include horseshoe crab tracks, swim traces, and lizard-like tracks (Lockley et al. 1998).

Park Collections

The NABR collections database includes 78 catalog numbers for paleontological materials, all but one of which pertain to Quaternary specimens. The lone non-Quaternary specimen is NABR 100, a fossil fern or conifer of unknown provenance. It is also the only specimen currently in an NPS facility, being reposited in Southeast Utah Group collections (V. Webster, pers. comm., November 2009).

The Quaternary specimens include conifers (19 catalog numbers), dicots (7), unidentified plants (15), *Neotoma* dung, midden material, and other related specimens (8), *Marmota* (3), *Oreamnos* (5, probably including dung), unidentified mammals (5), mammal dung (2), unidentified animals (4), undifferentiated composite specimens (4), and bulk sediment (5). All of the plant, composite, and *Neotoma*-related specimens are currently

in the collections of Northern Arizona University (Flagstaff; NAU), including NABR 1328-1355 and 2539-2574. The *Marmota* material and most of the *Oreamnos*-related specimens and bulk sediment are at the Museum of Northern Arizona (MNA; Flagstaff), including NABR 1315-1327 (V. Webster, pers. comm., September 2009).

Cultural Resource Contexts

NABR has archeological sites and a record of human presence that is over 1,000 years old (Huntoon et al. 2010), so it would not be surprising to find paleontological resources with or as cultural resources. Stokes (1972) interpreted three petroglyphs in southeastern Utah as representing proboscideans (possibly mastodons), but such identifications are extremely controversial. As there are many examples of rock art at NABR, depictions of extinct megafauna (or art interpreted as such depictions) may eventually be found at the monument as well. Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The monument should consider future field inventories for paleontological resources to more fully document *in situ* occurrences of fossils. The monument may consider a formal site documentation and condition assessment for significant fossil localities. Monitoring of significant sites should be undertaken at least once a year in the future. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.

- Within the Northern Colorado Plateau Network (NCPN), the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and increase understanding of this important fossil-bearing unit. A similar USGS–NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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1964–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by USNM staff; status: Active.

Southeast Utah Group Museum
1908–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by SEUG staff; status: Active.

Pipe Spring National Monument

Pipe Spring National Monument (PISP), Arizona, preserves historic and cultural resources associated with Pipe Spring, a desert water source that has attracted people for thousands of years (PISP Figure 1). It is best known for the historic cattle ranch associated with Winsor Castle, both Mormon pioneer sites. PISP was proclaimed May 31, 1923. It encompasses 16 ha (40 acres), all of which are under federal administration.

Geologic Background

PISP is located in northeastern Mohave County, northern Arizona. It is surrounded by the Kaibab Indian Reservation. Few large population centers are found in the area; it is about 21 km (13 miles) west-southwest of Fredonia, Arizona, and 80 km (50 miles) southeast of St. George, Utah. PISP is located on Winsor Point on the southeast tip of Winsor Mountain, which itself is part of the Moccasin Terrace, on the southern escarpment of the Vermillion Cliffs of northern Arizona and southern Utah. The monument is associated with the large Sevier Fault, which roughly bisects it in a north-south direction; areas west of the fault are associated with the Uinkaret Plateau, and areas east of it are associated with the Kanab Plateau. About 1.6 km (1 mile) north of the monument the fault forks (Billingsley et al. 2004). Just west of PISP are two folds. First is a syncline, or concave-up fold. Just beyond the syncline, following the axis of Winsor Mountain, is the Moccasin Monocline, a step-like fold that descends to the east (Billingsley et al. 2004).



PISP Figure 1. Much of PISP is covered by Quaternary alluvium, with heights of Kayenta Formation and Navajo Sandstone rocks (NPS photo).

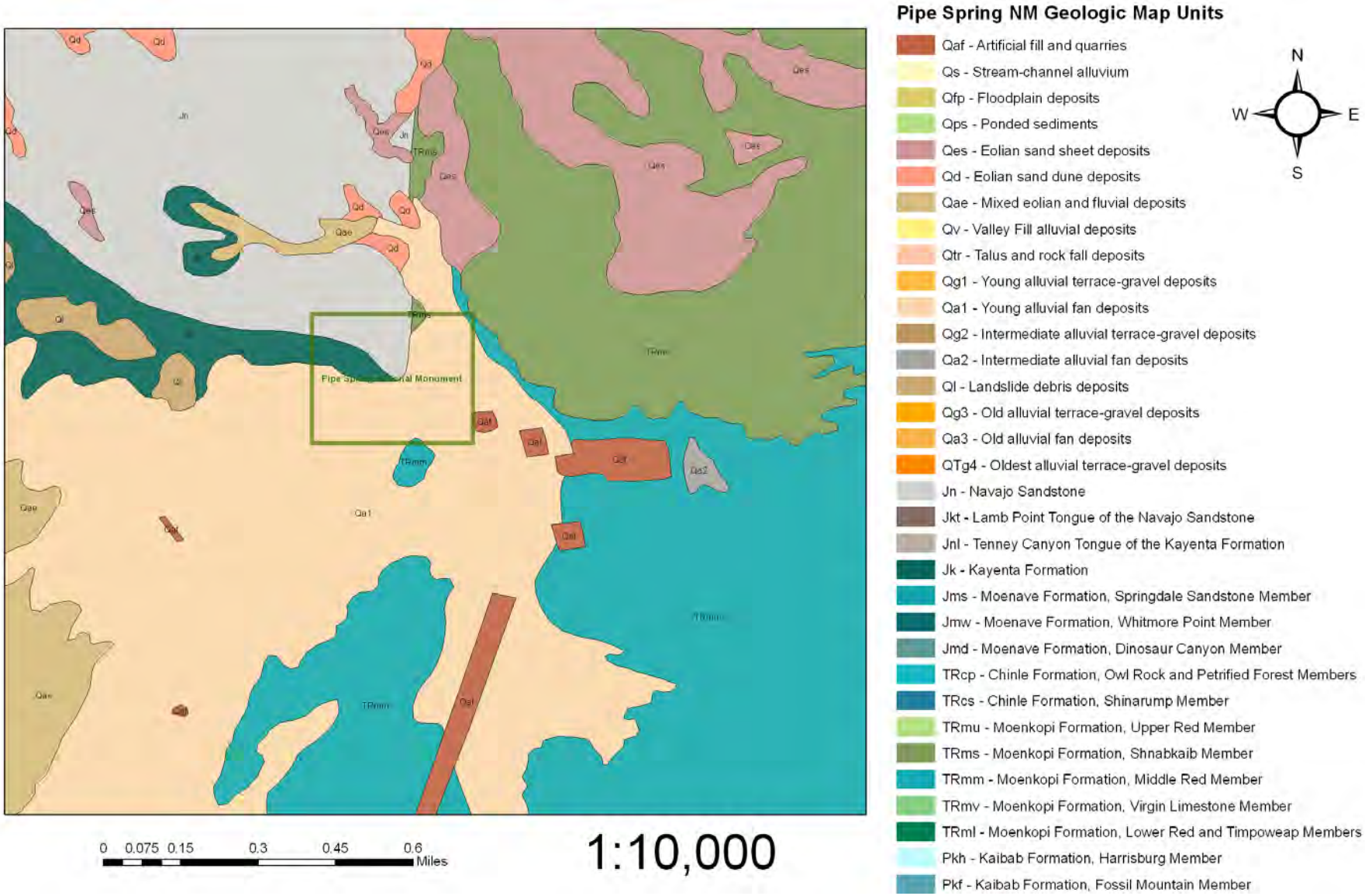
The geology of PISP has been described in several documents, including Truini (1999, 2002), National Park Service (NPS) (2001), Billingsley et al. (2004, 2008), Truini et al. (2004), Sharrow (2009), and Graham (2010). PISP fossils have been described by Stokes (1988), Santucci (1991), Cuffey et al. (1997, 1998), Santucci et al. (1998), Koch and Santucci (2002), and Hunt et al. (2005). A geologic scoping session coordinated by the Geologic Resources Division of the NPS was held for PISP during June 2001 (NPS 2001), leading to a geologic report for the monument (Graham 2010). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).

The geologic record exposed at PISP covers the Lower Triassic, Lower Jurassic, and the Quaternary (see the appendix for a geologic time scale). Triassic rocks at PISP are represented by the Moenkopi Formation of the Early–Middle Triassic, approximately 245 Ma (million years ago). During the Early and Middle Triassic, northern Arizona and southern Utah formed a broad coastal plain that was occasionally flooded by a sea to the west (Stewart et al. 1972; Blakey et al. 1993; Hintze and Kowallis 2009). The Moenkopi Formation is usually succeeded by the Chinle Formation, but the Chinle Formation is not seen at PISP because of faulting and erosion. Instead, the next highest formation exposed at the monument is the Kayenta Formation, deposited primarily by fluvial processes on a broad, gentle floodplain separated from the sea to the west (Blakey 1994; Peterson 1994). About 190 Ma, this setting was gradually replaced by the vast erg that became the Navajo Sandstone (Harshbarger et al. 1957; Dickinson and Gehrels 2003, 2009; Rahl et al. 2003; Lucas et al. 2005), the largest erg known from North America (Blakey 2008; Blakey and Ranney 2008). This Sahara-like desert was found throughout the Four Corners area into Nevada and Wyoming (Hintze and Kowallis 2009). At this time, the latitude of the area was only about 10° N, near the western edge of the supercontinent Pangaea (Loope et al. 2004a). Strong winds prevailed from the north to northwest (Irmis 2005), with cross-equatorial winds in the winter, and monsoonal rain in the summer (Loope et al. 2004a, 2004b). The youngest period of Navajo Sandstone deposition is not preserved at PISP, and post-Navajo Sandstone deposition is only represented by unconsolidated sediment from the Quaternary. However, the effects of several geologic events are represented at PISP, even if deposition is not.

The Moccasin Monocline formed over ancient fault zones during the Laramide Orogeny (Billingsley et al. 2004, 2008), a mountain-building event that took place from the Late Cretaceous to the Eocene. This orogeny was responsible for much of the present form of the Rocky Mountains (Aubrey 1991). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). In the PISP area, faulting on the Sevier Fault system has occurred since the

Late Miocene. Movement was extensive during the Pliocene and Pleistocene, with up to 395 m (1,300 ft) of vertical displacement on the Sevier Fault at PISP (downward movement on the west; Billingsley et al. [2004]). The vertical movement resulted in rocks of substantially different ages (Early Triassic and Early Jurassic) being put into lateral contact. Igneous activity also flared up in southwestern Utah and northwestern Arizona during the Pliocene and Pleistocene. A series of igneous events occurred in southwestern Utah beginning around 2.4 Ma as part of the Western Grand Canyon basaltic field (Biek et al. 2010). Volcanism took place south and southwest of PISP during the Quaternary, as part of the Uinkaret volcanic field (Billingsley et al. 2008). People have been present in the PISP area for at least the past 8,000 years (Truini 1999).

Bedrock units exposed at PISP include, from oldest to youngest, the Moenkopi Formation (Lower Triassic), and the Kayenta Formation and Navajo Sandstone of the Glen Canyon Group (Lower Jurassic). Surficial units include Quaternary alluvial fan sediments (see PISP Figure 2 for a geological map) (Billingsley et al. 2004; Graham 2010). *In situ* trace fossils are known from the Navajo Sandstone in PISP, there are tracks in a loose boulder potentially from the Kayenta Formation, and fragments of reworked petrified wood can be found in the Quaternary sediments (PISP Table 1). There are also fossils in PISP museum collections.



PISP Figure 2. Schematic geological map of PISP. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1043912.

Formation	Age	Fossils within PISP	Depositional Environment
Quaternary sediments	Holocene	Reworked petrified wood	Primarily alluvial fan deposits
Navajo Sandstone	Early Jurassic	Dinosaur tracks	Desert with large eolian sand dunes
Kayenta Formation	Early Jurassic	Amphibian tracks	Fluvial, floodplain, and eolian settings
Moenkopi Formation	Early Triassic	None to date; microbial structures and marine invertebrates are most likely	Coastal plain, tidal flat, salt flat, and restricted marine settings

PISP Table 1. Concise tabulation of PISP stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of PISP present opportunities for education, interpretation, and continued or future scientific research in the monument.

Glen Canyon Group: Kayenta Formation (Lower Jurassic)

The Kayenta Formation of PISP and the immediate vicinity is composed mostly of dark red and light red-brown beds of lime-rich mudstone, siltstone, and sandstone. It is a slope-forming unit. The thickness of the formation ranges from 80 to 120 m (270 to 400 ft). In some locations in the area, it is divided by the Lamb Point Tongue of the Navajo Sandstone into a main body and overlying Tenney (sometimes misspelled Tenny) Canyon Tongue, but this is not the case at PISP (Billingsley et al. 2004). Sharrow (2009) interpreted the Navajo Sandstone capping Winsor Point as the Lamb Point Tongue, but later agreed with Billingsley et al.'s interpretation that this is the lowermost portion of the main body of Navajo Sandstone (Graham 2010). The upper part of the Kayenta Formation at PISP is sometimes described as a sandy facies (Truini 1999; Truini et al. 2004). The upper contact with the Navajo Sandstone is gradational but sharp (Billingsley et al. 2004). The Kayenta Formation makes up the lower slopes of Winsor Point west of the Sevier Fault in PISP, and can be found in the northwestern quarter of the monument (Billingsley et al. 2004). The upper Kayenta Formation intertongues with the lower Navajo Sandstone because the dunes of the latter overrode the streams of the former, advancing from the northeast (Riggs and Blakey 1983). The age of the Kayenta Formation is now thought to be middle Early Jurassic, between approximately 197 and 190 Ma (Lucas et al. 2005). The Kayenta Formation was formed by shifting freshwater braided and meandering streams (Colbert 1974); floodplain deposits are also known (Lucas et al. 2005).

The Kayenta Formation includes two major facies: a sandstone-rich facies near its type locality in northeastern Arizona, and a siltstone-rich facies to the west. These are known respectively as the typical (or sandy) facies and silty facies (Harshbarger et al. 1957). PISP is entirely within the silty facies (Blakey 1989), expressed as interbedded siltstone, mudstone, and sandstone. It is thought that the silty facies represent the sluggish lower part of Kayenta Formation rivers and streams (Harshbarger et al. 1957), where low-energy drainage fed lush floodplains (Jenkins and Shubin 1998). The silty facies is much more fossiliferous than the sandy facies, which is best known for its vertebrate tracks (Hamblin and Foster 2000; Lucas and Tanner 2007). Examples of sandy-facies fossils can be found at Canyonlands National Park (CANY) (Santucci et al. 1998; Santucci and Kirkland 2010), Capitol Reef National Park (CARE) (Santucci et al. 1998), and Glen Canyon National Recreation Area (GLCA) (Lockley et al. 1998; Santucci and Kirkland 2010), extending through the formation proper into the Kayenta Formation–Navajo Sandstone transition zone.

There is one report of potential fossils from the Kayenta Formation within PISP: Andrea Bornemeier reported the presence of amphibian tracks on a fallen boulder that most likely fell from a nearby Kayenta Formation outcrop. It was first located by one of the authors (VLS) (PISP Figures 3, 4, and 5) (A. Bornemeier, PISP chief of interpretation and resource management, pers. comm., June 2011).



PISP Figure 3. Kayenta Formation block with possible tracks (photo courtesy Andrea Bornemeier, PISP).



PISP Figure 4. Close-up on part of possible tracks (photo courtesy Andrea Bornemeier, PISP).



PISP Figure 5. Second close-up on part of possible tracks (photo courtesy Andrea Bornemeier, PISP).

Most vertebrate body fossils from the Kayenta Formation are from the silty facies of the formation, and were discovered in the Gold Spring Wash area of northeastern Arizona, over 160 km (100 miles) southeast of PISP in the vicinity of Cameron and Tuba City (Lucas et al. 2005). Per Lucas et al. (2005), taxa represented in the formation by body fossils include sharks, bony fish, the frog *Prosalirus*, the caecilian (limbless amphibian) *Eocaecilia*, the turtle *Kayentachelys*, a sphenodontian reptile (tuatara), numerous crocodylomorphs (crocodiles and their closest relatives) including *Calsoyasuchus*, *Eopneumatosuchus, Kayentasuchus*, *Protosuchus*, and unnamed taxa, the pterosaur (flying reptile) *Rhamphinion*, theropod dinosaurs (bipedal and generally carnivorous) *Dilophosaurus*, "Megapnosaurus" *kayentakatae*, and the undescribed "Shake N Bake" theropod, a prosauropod dinosaur (early long-necked herbivores) now named *Sarahsaurus* (Rowe et al. 2011), armored dinosaurs *Scelidosaurus* and *Scutellosaurus*, an unnamed heterodontosaurid dinosaur, the tritylodontid synapsids ("mammal-like reptiles") *Dinnebiton*, *Kayentatherium*, and *Oligokyphus*, possible early mammal *Dinnetherium*, and a haramyid mammal (there are discrepancies between the table and text in Lucas et al. [2005]; the text seems more accurate). Also reported are lungfish (Jenkins and Shubin 1998), salamanders (Curtis and Padian 1999), lizards (Jenkins et al. 1983), the theropod *Kayentavenator* (Gay 2010), and morganucodontid mammals (Jenkins et al. 1983). Sandy facies fossils include some of the crocodylomorphs. Silty facies fossils are more diverse, and include petrified wood, amphibians, turtles, lizards, abundant sphenodontians, *Eopneumatosuchus*, pterosaurs, *Dilophosaurus* and other theropods, theropod tracks, prosauropods, *Scutellosaurus*, and haramyid and morganucodontid mammals. Fossils in both facies include coprolites (fossil feces), fish scales, crocodylomorphs, and tritylodontids such as *Kayentatherium* (Clark and Fastovsky 1986). Frogs are the rarest component of the assemblage and are only found in the silty facies (Jenkins and Shubin 1998).

Kayenta Formation vertebrate trace fossils include coprolites (Luttrell and Morales 1993), therapsid ("mammal-like reptile") tracks, lizard-like tracks (Hamblin and Foster 2000), and dinosaur tracks such as *Grallator* and *Eubrontes* (Lucas and Tanner 2007). Kayenta Formation fossils that are not vertebrates or vertebrate traces include microbial limestone (Luttrell and Morales 1993), petrified wood, which is locally abundant in the silty facies (Jenkins and Shubin 1998), plant impressions (Rigby et al. 1971), freshwater bivalves and gastropods (Harshbarger et al. 1957), ostracodes (seed shrimp) (Lucas and Tanner 2007), and invertebrate trails and burrows (Luttrell and Morales 1993).

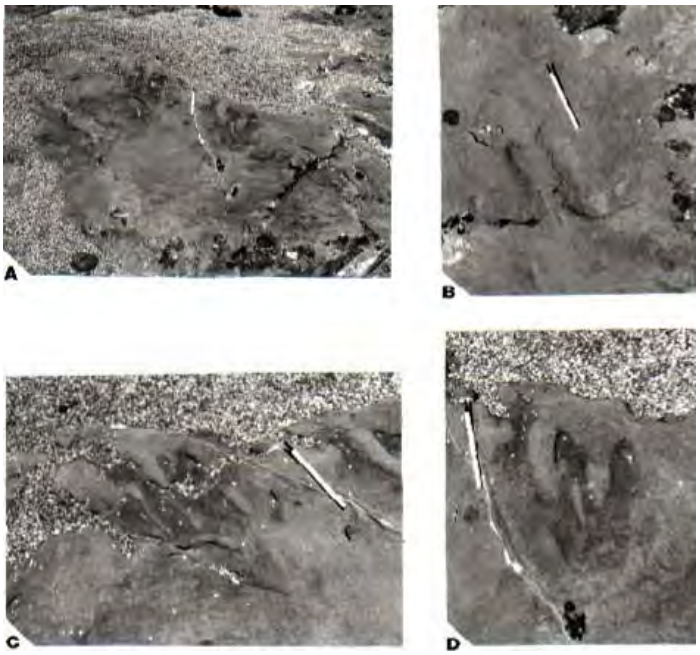
Glen Canyon Group: Navajo Sandstone (Lower Jurassic)

The Navajo Sandstone is exposed in the northwest quarter of PISP, where it caps Winsor Point (Billingsley et al. 2004). It is a cliff-forming unit composed of white, light red, and yellow-gray quartz-rich sandstone. Dark red sandstone and siltstone beds are present near the base, and purple-gray lime-rich sandstone lenses representing freshwater deposits are found interspersed. It can be very thick in the area, on the order of 460 to 610 m (1,500 to 2,000 ft) (Billingsley et al. 2004). While dating of Glen Canyon Group units has proven difficult to establish, the bulk of the Navajo Sandstone probably dates to the middle and late Early Jurassic (Lucas et al. 2005), between approximately 190 and 175 Ma. Formerly, the formation was thought to intertongue with Middle Jurassic units, but this was based on misinterpretations of similar eolian sandstones above it (Peterson and Pipiringos 1979).

The Navajo Sandstone represents the largest erg known from North America (Blakey 2008), and Kocurek (2003) identifies it as the largest known on Earth. As discussed in the "Kayenta Formation" section, the upper Kayenta Formation intertongues with the lower Navajo Sandstone as the dunes of the latter overrode the streams of the former. Recent research on geochronology of detrital zircon grains in the Navajo Sandstone suggests that most of the sand was eroded from the ancestral Appalachian Mountains, transported westward by a continental-scale river system to the western shore of Jurassic North America, and was then blown southward and incorporated into the Navajo–Nugget–Aztec dune field (Dickinson and Gehrels

2003, 2009; Rahl et al. 2003). Preservation of this tremendous thickness of wind-blown sand was made possible because of basin subsidence associated with Early Jurassic compressional deformation near the west margin of North America. This deformation caused the continental interior to flex downward, creating accommodation space for sediment accumulation (Allen et al. 2000).The Navajo Sandstone formed in a tropical arid setting with eolian, playa (ephemeral lake), and sabkha (salt flat) deposits (Mahaney et al. 2004). Wetter restricted interdunes (wetlands or ponds in depressions in the dune field) are also known, and are characterized by mudstone with planar bedding, dolomitic limestone, and sandstone (Irmis 2005). Interdune deposits are best known from the vicinity of Moab in Utah, over 340 km (210 miles) to the northeast (Wilkens 2008), but are also known from the Navajo Sandstone of ZION (Doelling et al. 2002) and Glen Canyon National Recreation Area. These deposits range from fluvial sandstones to lake limestones. During periods of increased precipitation, vegetation spread (Wilkens 2008). Fossils are mostly associated with interdunes (Loope et al. 2004b). As time went on during the deposition of the Navajo Sandstone, the climate became drier (Irmis 2005), although there were several periods of increased precipitation with durations on the order of tens of thousands of years (Milan et al. 2008).

The Navajo Sandstone is not extensively fossiliferous, but does preserve fossils within PISP in the form of tetrapod tracks. The first report was made by Stokes (1988), who published a photograph of a natural track cast which was on display at the monument. Roger J. Cuffey, a paleontology professor at Penn State, was following up on the report in 1995 when he was shown additional *in situ* tracks by PISP personnel. This find consisted of three tridactyl (three-toed) tracks found on a foot trail on the mesa behind the visitor center, about 2 m (7 ft) above the base of the Navajo Sandstone. These tracks are fairly large, measuring 30 cm (12 in) in length along the middle toe and 30 cm (12 in) in width across the two toes flanking it. Two of the tracks are essentially side-by-side, with a third 2.7 m (8 ft) ahead and on approximately the same line. Cuffey et al. (1997, 1998) found them to be most like the common Early Jurassic theropod tracks known as *Eubrontes*. Hunt et al. (2005) noted that these three tracks are unusual in that the two lateral toes are almost as long as the middle toe, instead of being substantially shorter. They also described the Stokes track as resembling Late Triassic tracks assigned to *Pseudotetrasauropus*, a tetradactyl (four-toed) track often preserved as a tridactyl underprint (Hunt et al. 2005; see Lockley et al. [2006] for further discussion of the nomenclature).



PISP Figure 6. Dinosaur tracks in the Navajo Sandstone at PISP (Cuffey et al. 1998). Per Cuffey et al., "A, left and right footprints together. B, isolated footprint. C, left footprint. D, right footprint."

To date, fossil organisms and traces represented in the Navajo Sandstone over its entire depositional area include stromatolites (layered microbial structures), root traces (Irmis 2005), pollen, spores, horsetails, ferns (Loope et al. 2004b), cycadeoids, conifers (Wilkens 2008), petrified wood (Irmis 2005), possible fungal filaments (Mahaney et al. 2004), worm burrows (Stokes 1978), insect burrows, scorpion tracks (Loope et al. 2004b), unionid bivalves (Wilkens 2008), ostracodes, conchostracans (clam shrimp) (Harshbarger et al. 1957), a crocodylomorph, the small theropod dinosaur *Segisaurus* (Irmis 2005), at least two types of prosauropods including *Seitaad* (Sertich and Loewen 2010), a tritylodontid synapsid, and vertebrate trace fossils (Irmis 2005). The assemblage may be a persistent remnant of the Kayenta Formation assemblage (Winkler et al. 1991). Sandstone dune deposits have mostly yielded vertebrates, while limestone interdune beds have mostly yielded plants and invertebrates (Stokes 1991), although arthropod tracks (*Octopodichnus*) are known from the dunes (Hunt and Lucas 2006). Interdune fossils include stromatolites, root casts, horsetails, ostracodes, and insect traces, with locally common petrified wood (Irmis 2005).

Vertebrate trace fossils are much more common than body fossils in the Navajo Sandstone, and have been extensively described. More than 60 dinosaur track sites are known in the Navajo Sandstone across its depositional area (Loope 2006; Milan et al. 2008). The various vertebrate trace fossils have gone under different names and interpretations, depending on the author. Recent works (Irmis 2005; Hunt and Lucas 2006;

Loope 2006) seem to have settled on seven or eight primary ichnogenera: *Brasilichnium* (synapsid, but see Milan et al. 2008 for an interpretation as a sauropodomorph [prosauropods and sauropods, informally known as brontosaurs]), *Batrachopus* (crocodylomorph), *Grallator* (small theropod), *Anchisauripus* (medium theropod), *Eubrontes* (large theropod), *Navahopus* (small ?sauropodomorph), *Otozoum* (larger sauropodomorph), and *Anomoepus* (ornithischian [bird-hipped] dinosaur) (Irmis 2005). A few oddities are also known, including lizard-like tracks (*Lacertipus*), a track once thought to be pterosaurian in origin ("*Pteraichnus*"), bird-like *Trisauropodiscus* tracks (Irmis 2005), undescribed large ornithopod tracks (Santucci and Kirkland 2010), and large vertebrate burrows thought to have been made by mammals and/or mammal relatives (Lucas et al. 2006; Riese et al. 2011). So far, large traces (*Anchisauripus*, *Eubrontes*, and *Otozoum*) are only known from interdunes, with smaller traces (*Batrachopus*, *Brasilichnium*, and *Grallator*) the only tracks found in dune sands (Loope 2006), although this differentiation may be due to preservational bias (Irmis 2005). In general, the vertebrate trace fossils are consistent with the known body fossils.

Quaternary sediments (Holocene)

Quaternary sediments mantle most of the northeast, southeast, and southwest quarters of PISP. They are dominated by alluvial fan sediments shed from the local highlands, and are thus mostly made up of material from Mesozoic formations (Billingsley et al. 2004). The sediments range in size from clay and silt to sand and gravel, and are thought to have been deposited during the Holocene. Deposits are 1.5 to 12 m (5 to 40 ft) thick (Billingsley et al. 2004). Although no Holocene fossils are known from these surficial deposits at PISP, the presence of reworked petrified wood justifies including the deposits with the Kayenta Formation and Navajo Sandstone as fossiliferous units at the monument. Surficial fragments of petrified wood are well-known in the PISP area, and the monument has an informal collection (A. Bornemeier, pers. comm., June 2011). The fossils derive from the Shinarump Member of the Chinle Formation, a Late Triassic-age unit present nearby (Billingsley et al. 2004, 2008).

The most fossiliferous types of Quaternary deposits on the Colorado Plateau are cave deposits and lake deposits, neither of which are present at PISP. Based on the type of deposits present in the monument, the most likely Quaternary fossils are isolated remains of mammals, particularly resistant materials such as teeth. The Holocene age excludes most of the extinct Pleistocene megafauna, such as sloths, mastodons, mammoths, horses, and camels, but bison are a possibility. The presence of a water source at PISP over at least the past 8,000 years (Truini 1999) improves the potential for finding Holocene-age fossils. Additionally, crevices, rockshelters, and alcoves in the cliff-forming Navajo Sandstone may harbor fossil packrat middens; these collections of urine-cemented plant fragments and other organic materials are important records of Quaternary environmental conditions in the Southwest. Fossil middens are abundant in Grand Canyon National Park (GRCA; see summaries in Cole [1990a, 1990b], Anderson et al. [2000], Thompson and Anderson [2000], and Kenworthy et al. [2004]). PISP has modern midden specimens in its collections, as well as typical terrestrial and freshwater Quaternary bivalves, gastropods and ostracodes (J. Leasor, PISP museum curator, pers. comm., June 2011). Otherwise, the fossil potential of PISP's Quaternary sediments is low, outside of the reworked petrified wood.

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the Moenkopi Formation within PISP. However, it is known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from it.

Moenkopi Formation (Lower Triassic)

The Moenkopi Formation is a heterogeneous coastal plain and shallow marine unit present in several western states. It has been divided into several members depending on location. At PISP, the formation's divisions follow the nomenclature established in adjacent areas of southwestern Utah. Two of the middle members are present, the middle red member and the overlying Shnabkaib Member (see McKee [1954], Stewart et al. [1972], or the summary for ZION in this document for descriptions of the other members of this area). The middle red member is present in two small areas of PISP, in the northeastern corner and along the southern boundary. The Shnabkaib Member is present in a small area east of the Sevier Fault on the east side of Winsor Point (Billingsley et al. 2004). The Moenkopi Formation dates to the Early and early Middle Triassic (Heckert et al. 2005); the members at PISP date to the Early Triassic (Blakey et al. 1993), between about 250 and 245 Ma.

The Moenkopi Formation of southwestern Utah and adjacent areas of Arizona and Nevada consists of a basal conglomeratic member under three pairs of alternating limestone/evaporate units and red-bed units, representing parts of three transgression-regression cycles. The middle red member and the Shnabkaib Member are the regressive phase of the second cycle and the transgressive phase of the third cycle (Stewart et al. 1972; Biek et al. 2010). The middle red member of PISP and the immediate area is composed of red-brown slope-forming gypsum-rich siltstone and sandstone, with some beds of white dolomite, green siltstone, and gray-green to red mudstone. It grades out of the underlying Virgin Limestone Member (not exposed at PISP) and grades into the overlying Shnabkaib Member, with a thickness of 85 to 90 m (280 to 300 ft) (Billingsley et al. 2004). It was deposited in tidal flat and coastal

plain settings (Biek et al. 2010) during a regression (Stewart et al. 1972). The Shnabkaib Member is a heterogeneous slope-forming unit. Most of it is composed of white or gray dolomite and interbedded gray lime- and silt-rich gypsum, with some red mudstone, siltstone, and sandstone beds. The red beds are more prominent near the top and bottom of the member, where it grades into red members. It is 25 to 35 m (90 to 110 ft) thick (Billingsley et al. 2004). The Shnabkaib Member includes tidal flat, sabkha, and restricted marine depositional settings, and has been compared to the Trucial Coast of the Persian Gulf and lagoons in Ojo de Liebre, Baja California (Lambert 1984). This member represents another transgression, although not as extensive as previous events documented in the Moenkopi Formation (Stewart et al. 1972).

The Moenkopi Formation is not known to be fossiliferous within PISP, and prospects for finding fossils in it at the monument are poor because of the small exposures and the sparsely fossiliferous nature of the middle red and Shnabkaib members. Elsewhere, the middle red member has produced unspecified fossils (Lundberg and Lohrengel 2005), and the Shnabkaib Member has produced marine fossils including stromatolites, algal filaments, brachiopods (lamp shells), bivalves, ammonites (nautilus-like cephalopods), gastropods, crinoids (sea lilies), and invertebrate trace fossils. Microbial structures are common in this member, but fossils are otherwise rare (Lambert 1984). Bivalves, gastropods, and crinoids may be found in Shnabkaib Member carbonate beds (Hintze 1986; Jenson 1986). In general, Moenkopi Formation red members may yield stromatolites, microbial-coated grains, burrows and bioturbation (general disruption of sediments by organisms) (Reif and Slatt 1979), and bits of wood and bone (Gregory and Williams 1947). Red members may also include tetrapod tracks, as found at ZION (Mickelson et al. 2006). Limestone beds interbedded with the red beds may contain skeletal fragments of brachiopods, bivalves, gastropods, and ostracodes (Reif and Slatt 1979). Invertebrate fossils are known from the Moenkopi Formation in the Kanab Valley near Fredonia, including brachiopods, bivalves, ammonites, other cephalopods, gastropods, crinoids, and trace fossils (Gilbert 1875; Walcott 1880; McKee 1954).

Park Collections

PISP collections include 48 catalog numbers for paleontological materials. Thirty-two of the numbers are for materials from modern packrat middens collected on the monument in 1995 by Jim Mead, then based out of Northern Arizona University in Flagstaff. These are mostly plant fragments, but also include insect fragments, a lizard maxilla (upper jaw bone), and packrat dung. Another ten are assigned to Quaternary invertebrates: five for gastropods, three for bivalves, and two for ostracodes. The remainder are assigned to dinosaur tracks and marine invertebrates (bivalves, ammonites, and crinoids) (J. Leasor, pers. comm., June 2011). The dinosaur track casts were probably collected near but outside of the monument. Additionally, there is an informal collection of petrified wood in the maintenance area (A. Bornemeier, pers. comm., June 2011).

Cultural Resource Contexts

PISP has the potential for fossil-cultural resource associations, but none have been recognized yet. The springs have attracted people for at least the past 8,000 years (Truini 1999), some of whom left lithic artifacts, charcoal, and petroglyphs within PISP boundaries (McKoy 2000). The use of petrified wood for tools is common in the Southwest, and finding such tools at PISP would not be surprising. There is also an anecdote that members of the Civilian Conservation Corps stationed at PISP collected petrified wood from off the monument, presumably from the Arizona Strip, during the 1930s (A. Bornemeier, pers. comm., June 2011).

Given the presence of petroglyphs and dinosaur tracks at PISP, it is possible that petroglyphs of tracks may be found. There are at least two cases that have been interpreted as such in the vicinity: at the South Gate of ZION, 43 km (27 miles) northwest of PISP; and at Flag Point, Grand Staircase–Escalante National Monument, 46 km (29 miles) northeast of PISP (Stokes and Bruhn 1960; Lockley et al. 2006b; these associations are not universally accepted). Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The recommendations of Cuffey et al. (1998) should be reviewed in regard to what has been accomplished and what may still be implemented. They recommended protecting the three exposed Navajo Sandstone tracks and marking them with a plaque, installing an exhibit about the tracks in the visitor center, correcting incorrect terminology on the extant trailside plaque, and searching for more tracks near the known tracksite.

Of these, the terminology has been corrected (A. Bornemeier, pers. comm., July 2011). A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- The possible Kayenta Formation tracks should be evaluated by a specialist.
- The Navajo Sandstone tracks require additional study, because of their unusual proportions.
- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.
- Considering the relatively small size of the monument, staff should work with adjacent land owners and encourage studies on these lands.
- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. Zion National Park, which currently provides assistance to PISP for archeology, or the Geologic Resources Division can coordinate additional documentation/research of such material.
- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.
- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.
- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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1923–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by PISP staff; status: Active.

Pipe Spring National Monument Digital Geologic Map
2007–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=1043912.

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Timpanogos Cave National Monument

Timpanogos Cave National Monument (TICA), Utah, was established to protect a network of limestone caves noted for their abundant speleothems (cave formations, formed by minerals leaving solution in water) (TICA Figure 1). The network consists of three large caves (Hansen Cave, Middle Cave, and Timpanogos Cave) and tunnels and smaller passages in the Wasatch Mountains. TICA was proclaimed October 14, 1922, and was transferred from the U.S. Forest Service to the National Park Service August 10, 1933. It encompasses 101 ha (250 acres), all of which are under federal administration.

Geologic Background

TICA is located in the northeastern corner of Utah County in north-central Utah. It is on the north side of Mount Timpanogos, one of the peaks of the Wasatch Range. Most of the monument's land is on the south face of American Fork Canyon, carved by the American Fork River, 357 m (1,171 ft) below the entrance of Timpanogos Cave. Uinta National Forest surrounds TICA, with the Lone Peak Wilderness unit just north of the monument. To the west of the Wasatch Range is the Wasatch Front, where most of Utah's population is concentrated. Central Provo is about 23 km (14 miles) south of TICA, and central Salt Lake City is about 40 km (25 miles) north-northwest of TICA. The geology of TICA has been discussed in a number of documents, usually in reference to the monument's cave system. Documents dealing with more general geologic aspects of TICA, or its paleontology, include: Bullock (1942, 1944); Baker and Crittenden (1961); White and Van Gundy (1971, 1974); Hall (1981); Herron (1998); George (1999); National Park Service (NPS) (1999); Santucci et al. (2001); Koch and Santucci (2002); Thornberry-Ehrlich (2006); Mayo et al. (2009, 2010); Santucci and Kirkland (2010); Willis and Wills (2010); and Constenius et al. (2011). The Geologic Resources Division of the NPS coordinated a geologic resources inventory scoping session for TICA during May 1999 (NPS 1999), and produced a geologic resources evaluation report for the monument in 2006 (Thornberry-Ehrlich 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002).



TICA Figure 1. Heart-shaped speleothem at TICA (VLS).

The geologic history exposed at TICA includes three major episodes of deposition followed by prolonged tectonic activity. Deposition occurred during the late Neoproterozoic, the Cambrian, and the late Devonian to Mississippian (see the appendix for a geologic time scale). The oldest outcropping rocks at TICA, the quartzite (metamorphosed sandstone) and conglomerate of the Mutual Formation, were deposited at approximately 600 Ma (million years ago). These rocks were laid down on an older erosional surface (Mayo et al. 2010), as part of a braided fluvial system (Ehlers et al. 1997). Continental rifting preceded and possibly followed deposition (Christie-Blick 1997). Deposition was briefly interrupted by uplift and erosion, resuming at some time during the early Cambrian. The TICA area at this time was a shallow marine setting, with ongoing subsidence that allowed the deposition of thousands of meters of sediment over a few tens of millions of years. Concurrently, the sea became deeper, and sediment type changed in response. Thus, the oldest unit, the Tintic Quartzite, is made up of sand, while the overlying Ophir Formation has shale and limestone, and the Maxfield Limestone at the top of the Cambrian section is dominated by limestone, corresponding to the deepest water (Mayo et al. 2010).

There follows a gap in preserved deposition at TICA on the order of 150 million years in duration (Mayo et al. 2010). During this time, much of the Maxfield Limestone and Ophir Formation were eroded from the area (Baker and Crittenden 1961). Utah remained below sea level, and several transgression (marine advance)-regression (marine retreat) cycles occurred during the Middle Devonian and Mississippian in western Utah (Greenhalgh 1980), reflected in the formations present at TICA. Approximately 40 to 50 million years of shallow marine deposition are represented in the Devonian and Mississippian outcrops at TICA, including the limestone beds (the Deseret Limestone) that today host the monument's caves (Mayo et al. 2010).

Although there has been little deposition to record geologic events at TICA since the end of the Mississippian, nearby faults and folds and its present mountain canyon setting show that the area has been geologically active. The western margin of North America became a zone of renewed convergence during the Late Jurassic, which led to orogenies and igneous activity (Hintze and Kowallis 2009). First was the Sevier Orogeny, which lasted from the Late Jurassic into the Paleogene (Hintze

and Kowallis 2009), and typically produced folds and large, relatively thin thrust sheets of faulted rocks. The bedrock of TICA is part of one such thrust sheet (Mayo et al. 2010), which was active from the Late Cretaceous to the Late Eocene on the Charleston-Nebo fault (Constenius 1998; Willis and Wills 2010). Another mountain-building event, the Laramide Orogeny, began during the Late Cretaceous while the Sevier Orogeny was active and continued into the Eocene as well (Hintze and Kowallis 2009).

Compression was replaced by extension around 40 Ma, when the Sevier and Laramide orogenies ended. Initially, extension was accomplished by low-angle normal faults, such as the nearby Deer Creek detachment fault (Constenius 1998; Willis and Willis 2010). This style of faulting led the topographic highs created by thrust faulting to partially "relax" and slide back to the west (Constenius 1998; Vogel et al. 1998). The early phase of extension overlapped a period of igneous activity to the north of TICA. Intrusions of the Wasatch igneous belt pushed into the Wasatch Range between approximately 36 and 30 Ma, and the Keetley volcanic field was active between approximately 36.6 and 33.5 Ma (Vogel et al. 1998, 2001). Later, around 17 Ma, subduction of part of the East Pacific Rise beneath western North America accelerated extension, leading to high-angle normal faults. The resulting period of extension created the basin and range topography of the Great Basin. More locally, this extension led to uplift on the nearby Wasatch fault zone, which is responsible for much of the uplift of the Wasatch Range (Constenius 1998; Willis and Willis 2010). TICA is near the roughly perpendicular intersection of the Wasatch fault zone and the Uinta fault and fold system (White and Van Gundy 1971, 1974). The uplift of the Wasatch Range caused rivers to cut down through the newly uplifted rocks, leading to deep canyons like American Fork Canyon (Mayo et al. 2010). The TICA caves were partially filled with fine-grained fluvial sediments between about 780,000 and 350,000 years ago, when the American Fork River was at the elevation of the caves (Mayo et al. 2009). There has been 15 to 26 m (49 to 85 ft) of vertical movement on the Wasatch fault zone in American Fork Canyon over approximately the past 15,000 years, with 7 to 8 m (23 to 26 ft) of movement since about 5,300 years ago. Several large-magnitude seismic events have occurred on the fault zone at American Fork Canyon during the Holocene, most recently at 500 ± 200 years ago (Machette 1992; Machette et al. 1992). Faulting is also thought to be at least partially responsible for the configuration of the caves, by providing locations for preferential dissolution by water. The caves have existed since at least the period when the river was at their elevation (Mayo et al. 2010).

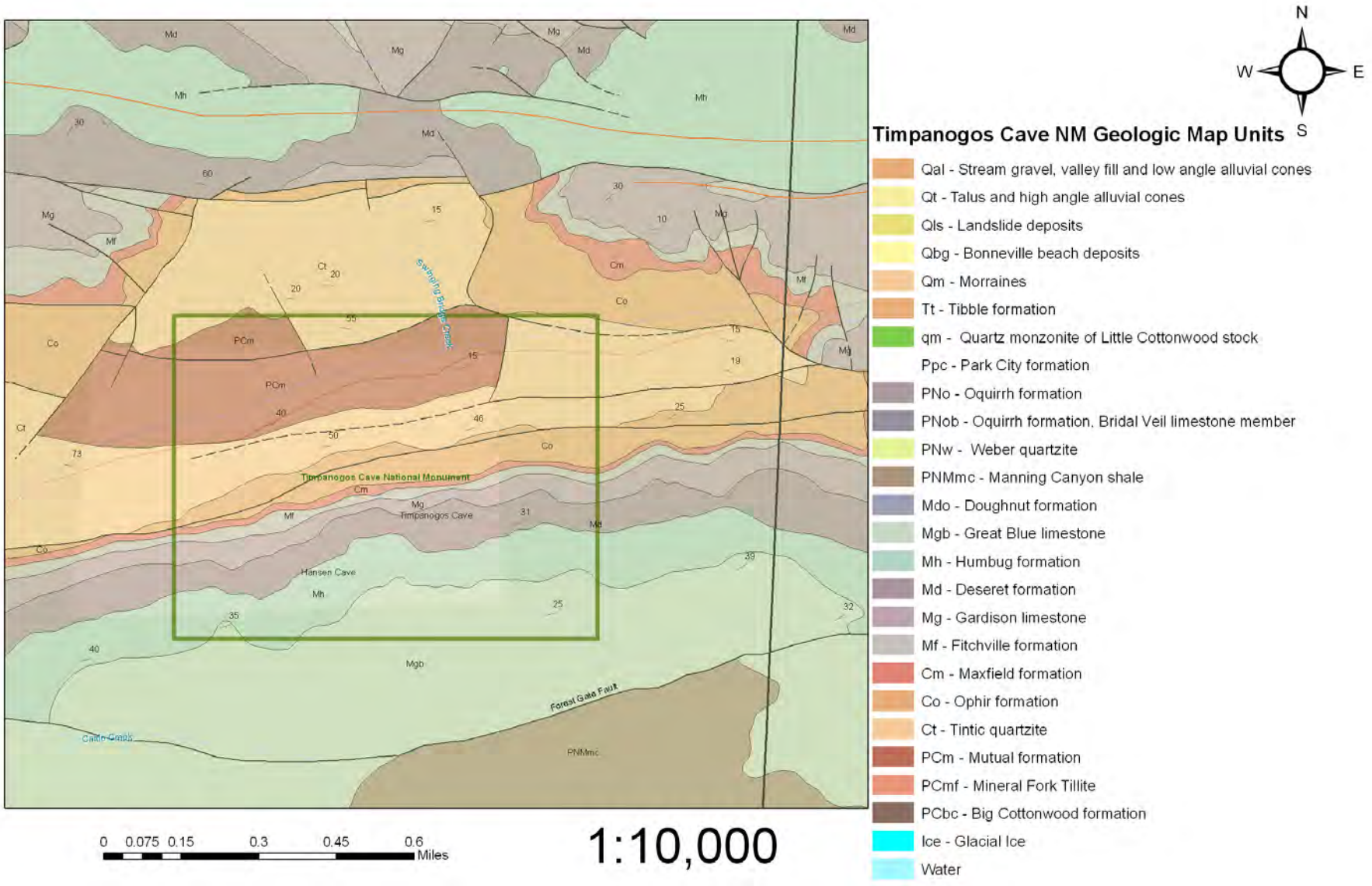
North-central Utah has also been influenced by recent glaciations and the former Lake Bonneville. Two Pleistocene glaciations have been identified in the Wasatch Range (Madsen and Currey 1979; Osborn and Bavis 2001). One, the Bull Lake glaciation, occurred about 150,000 to 100,000 years ago, and the other, the Pinedale glaciation, occurred between 26,000 and 12,300 years ago (Osborn and Bevis 2001; Munroe and Laabs 2009). Glaciers were present on Mount Timpanogos to as low as 1,900 m (6,230 ft) in elevation during the last glacial maximum at the end of the Pleistocene (Anderson and Anderson 1981), and ice was present above Tibble Fork, 6.4 km (4.0 miles) northeast of TICA on American Fork Canyon (Baker and Crittenden 1961). The canyon has a V-shaped cross section at TICA, indicating that ice did not reach it for any significant length of time (White and Van Gundy 1974). Since the end of glaciation, rock glaciers have been active on Mount Timpanogos during at least two periods over the Holocene (Anderson and Anderson 1981; Burke and Birkeland 1983).

Much of the area immediately west of the Wasatch Range was occupied by Lake Bonneville during portions of the Pleistocene. Marsh and shallow lacustrine environments appeared in the north part of the Great Salt Lake basin around 5 Ma. An open lake was present in the north part of the basin beginning about 800,000 to 600,000 years ago, but was only established in the south part of the basin around 300,000 years ago (Kowalewska and Cohen 1998). The largest iteration, Lake Bonneville, occupied 51,700 km^2 (20,000 $miles^2$) at its greatest extent (Rhode et al. 2005), reaching its maximum elevation of about 1,550 m (5,090 ft) above sea level around 19,000 years ago (Oviatt 1997). At this elevation it was well below TICA but present at the mouth of American Fork Canyon (Baker and Crittenden 1961). The lake dropped during a catastrophic flooding event shortly after, then lowered to the modern Great Salt Lake level over a few thousand years (Nelson and Madsen 1987). Great Salt Lake and a few other lakes are remnants of Lake Bonneville (Rhode et al. 2005). Sporadic volcanic events are recorded in Great Salt Lake cores (Davis 1998). Humans were present in the Bonneville basin perhaps before 12,300 years ago, and had reached the east side of Lake Bonneville over 11,000 years ago (Rhode et al. 2005).

Rock units exposed at TICA include, from oldest to youngest, the Mutual Formation (Neoproterozoic), Tintic Quartzite (lower–middle? Cambrian), Ophir Formation (lower?–middle Cambrian), Maxfield Limestone (middle Cambrian), Fitchville Formation (Upper Devonian–Lower Mississippian), Gardison Limestone (Lower Mississippian), Deseret Limestone (Lower–Middle Mississippian), Humbug Formation (Middle Mississippian), Great Blue Limestone (Middle–Upper Mississippian), and Quaternary-age sediments (Pleistocene–Holocene) (see TICA Figure 2 for a stratigraphic column and TICA Figure 3 for a geological map) (Baker and Crittenden 1961; Thornberry-Ehrlich 2006; Mayo et al. 2010). The Deseret Limestone is known to be fossiliferous within TICA, and Quaternary fossils have been found in the caves as well. All of the other units are potentially fossiliferous within TICA (TICA Table 1).

Age	Name	MYA	Maximum Thickness (feet)	Description
Tert.	Tibble Fork Formation	<30	0-2,500	Fluvial pebble to boulder conglomerate, sandy to shaly tuffaceous sediments of fluvial origin with thin interbeds of limestone.
Pennsylvanian	Oquirrh Group	>256	4,000+	Thin to thick bedded, tan weathering sandstone and quartzite near the top and medium to dark gray limestone(Bridal Vale Falls Limestone) at the bottom. Crops out on the slopes of Mount Timpanogos.
Mississippian	Manning Canyon Shale		1,600	Black shale interbedded with fine-grained and gritty sandstone and gray to black limestone. Exposures are generally poor and it forms a broad bench north and west of Mount Timpanogos.
	Great Blue Limestone		2,800	Calcitic thin-bedded, dark-gray to black limestone and shaly limestone. Weathers to light-gray to pinkish tan flake or thin slabs.
	Humbug Formation		800+	Alternating dark- to light gray, fine- to coarse-grained dolomite with tawny weathering, fine- to medium-grained limey sandstone.
	Deseret Limestone		500+	Massive, cliff and cave forming, dark- to light-gray, fine- to coarse-grained dolomite. Contains the Timpanogos Cave system.
	Gardison Limestone	>342	600+	Ten to thirty feet of dark-gray, coarse-grained limestone at the base that is overlain by 80 to 100 feet of thin-bedded, blue-gray limestone with silty partings. The silty partings make the middle unit banded. About 400 feet of dark gray, massive limestone and dolomite containing masses of white chert 2 to 4 inches thick form the upper unit.
	Fitchville Dolomite	<517	500+	Basal beds are 2 to 10 feet thick consisting of cross-bedded dolomitic sandstone. Two thick layers of massive- to medium-bedded, dark- to pale-gray dolomite above the base.
Cambrian	Maxfield Limestone		0-300	Thin-to thick-bedded, blue-gray, mottled magnesian limestone and dolomite. Oolitic to pisolitic beds at the base an a conspicuous layer of laminated white dolomite at the top.
	Ophir Formation		300	Brown calcareous sandstone (upper), massive gray limestone (middle), and olive-green shale (lower).
	Tintic Quartzite	543	1,300	Nearly pure, medium- to coarse-grained, white to pinkish quartzite. Lower beds are generally cross-bedded and pebbley.
Precambrian	Mutual Formation	543 620	0 -1,300	Rusty to red-purple quartzite, grit and conglomerate with minor shale. The coarser sandstones are tan, gray or greenish-gray. The finer sandstones are grayish-red. The conglomerates contain pebbles and boulders up to 5 feet in diameter.
	Mineral Fork Tillite	620 660	0 -200	Black massive conglomerate quartzite and sandstone with thin layers of bedded siltstone. The massive tillite contains rounded pebbles, cobbles, and boulders in a sand matrix cemented by glacial flour.
	Big Cottonwood Formation		1,300	Rusty- to buff-weathering, massive quartzite and gray to dull-purple shale.

TICA Figure 2. Stratigraphic column for TICA (from Mayo et al. [2010], used with permission of the Utah Geological Association).



TICA Figure 3. Schematic geological map of TICA. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=2164812.

Formation	Age	Fossils Within TICA	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	A partial bison skeleton and packrat middens with associated snake, bird, and mammal bones	Fluvial and alluvial deposits, local landslides, and cave fill
Great Blue Limestone	Middle–Late Mississippian	None to date; marine invertebrates are most likely, and bryozoans, brachiopods, and blastoids have been found elsewhere on Mount Timpanogos	Nearshore and shallower marine settings, possibly with coastal marshes
Humbug Formation	Middle Mississippian	None to date; marine invertebrates are most likely, and have been found elsewhere in American Fork Canyon	Shallow marine
Deseret Limestone	Early–Middle Mississippian	Solitary and colonial corals, crinoids, and rocks composed of fossil fragments	Open marine to shallow marine
Gardison Limestone	Early Mississippian	None to date; marine invertebrates are abundant elsewhere	Shallow marine becoming deeper over time
Fitchville Formation	Late Devonian–Early Mississippian	None to date; marine invertebrates are abundant elsewhere	Shallow marine to supratidal
Maxfield Limestone	Middle Cambrian	None to date; fossils are rare but present elsewhere in American Fork Canyon, and include algae, marine invertebrates, and invertebrate traces	Marine
Ophir Formation	Early? –Middle Cambrian	None to date; marine invertebrates are most likely	Marine
Tintic Quartzite	Early–Middle? Cambrian	None to date; fossils are rare and include trilobites and invertebrate traces	Shallow marine
Mutual Formation	Neoproterozoic	None to date; fossils are very rare in this formation in general	Fluvial, floodplain, and lacustrine settings

TICA Table 1. Concise tabulation of TICA stratigraphy, fossils, and depositional setting; details and references can be found in the text.

Sources of Paleontological Resources

The fossils of TICA present opportunities for education, interpretation, and continued or future scientific research in the monument.

Deseret Limestone (Lower–Middle Mississippian)

The Deseret Limestone of the TICA area is a cliff-forming formation that contains the monument's cave system. It is mostly composed of light to dark gray dolomite with chert, with a basal layer of black shale. The thickness is up to roughly 130 m (420 ft) (Baker and Crittenden 1961). It was extensively weathered, probably during the Pleistocene (White and Van Gundy 1971). It is often divided into members (Sandberg and Gutschick 1979, 1984), but it is not divisible at TICA (Baker and Crittenden 1961). The lowest part of the Deseret Limestone formerly was included in a formation called the Gardner Dolomite, along with the underlying Gardison Limestone and Fitchville Formation (see below) (Morris and Lovering 1961). The Deseret Limestone has also been known as the Pine Canyon Formation (Morris and Lovering 1961; Greenhalgh 1980).

The Deseret Limestone was deposited during the Early and Middle Mississippian (early Osagean to middle Meramecian stages per Sandberg and Gutschick [1979] and Sandberg et al. [1979], approximately 345 to 335 Ma). The upper contact with the Humbug Formation is conformable (Baker and Crittenden 1961) and can be gradational (Biek 2005). The Deseret Limestone completes the transgression-regression cycle begun

during the underlying Gardison Limestone, being deposited during the regressive part of the cycle (Greenhalgh 1980).

The Deseret Limestone is fossiliferous at TICA, having yielded colonial corals, solitary horn corals, crinoids (sea lilies), and coquinoid rocks (made up mostly of fossil fragments) at the monument (TICA Figures 4 and 5) (Santucci et al. 2001). Silty limestone beds in this formation opposite the mouth of Little Mill Canyon, 2.75 km (1.71 miles) up-canyon from TICA, are also very fossiliferous (Baker and Crittenden 1961). Beds 30 to 60 m (100 to 200 ft) above the phosphatic member in American Fork Canyon are extensively fossiliferous, with an assemblage including corals, bryozoans (moss animals), brachiopods (lamp shells), nautiloid cephalopods, and trilobites (Crittenden 1959).



TICA Figure 4. Coral from TICA (TICA 145) (photo courtesy Camille Pulham, TICA).

The fossil assemblage of the Deseret Limestone is similar to that of the Gardison Limestone (see below), but Deseret Limestone fossils are not as abundant, show more distortion, and have often been altered to silica. The Deseret Limestone also has more corals and brachiopods (Anderson 1974). Fossils are usually sparse (Baker and Crittenden 1961; Santucci et al. 2001; Biek 2005). Microorganisms and plants are represented by foraminifera (amoeba-like protists that form "shells"), radiolarians (another type of "shelled" protist), algal fossils (Sandberg and Gutschick 1979), woody plant material (Peterson et al. 2000), and lycopsids (clubmosses) (Tidwell and Jennings 1986). Invertebrates are represented by sponges (Sandberg and Gutschick 1979), corals, bryozoans, brachiopods, bivalves (Gilluly 1932), ammonite and nautiloid cephalopods (Peterson et al. 2000), gastropods (Gilluly 1932), ostracodes (seed shrimp) (DeCelles and Gutschick 1983), crinoids (Morris and Lovering 1961), and echinoids (sea urchins) (DeCelles and Gutschick 1983). Vertebrates and their relatives are represented by conodonts (extinct eel-like chordates) (Sandberg 1979; Sandberg and Gutschick 1979). Chert can include radiolarians, sponges, ostracodes, and echinoids (DeCelles and Gutschick 1983).



TICA Figure 5. Gastropod from TICA (TICA 4369) (photo courtesy Camille Pulham, TICA).

Quaternary sediments (Pleistocene–Holocene)

Quaternary deposits have not been completely mapped at TICA (Baker and Crittenden 1961; Constenius et al. 2011), but based on the monument's topography and geography, most Quaternary deposits probably consist of river sediments in American Fork Canyon, and various types of rockfalls and landslides. Quaternary deposits are also present in the caves, which have yielded probable Holocene fossils from several locations. Summaries of Quaternary fossil sites in Utah include Nelson and Madsen (1980, 1987), Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008). Typical Quaternary fossils of Utah include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). TICA's Quaternary fossils include both categories.

The best-known Quaternary fossils of TICA consist of bones found associated with packrat middens (TICA Figures 6 and 7). Several sites were located in the early to middle 1990s by Rod Horrocks during cave surveying and fill removal. They were excavated in 1998 and described by Christian George in 1999. Fossil sites were discovered in the Entrance Room of Hansen Cave, Hidden Mine Cave, and the Boneyard near the outside entrance to Timpanogos Cave. Most of the bones came from mammals, but a snake mandible and bird bones were recovered as well. Bones from marmots, ground squirrels, deer mice, packrats, voles, hares, and bighorn sheep were found at the Entrance Room. Snake, marmot, ground squirrel, deer mouse, packrat, vole, black bear, pine marten, mink, raccoon, and bighorn sheep bones were found at Hidden Mine Cave. The Boneyard yielded ground squirrel and packrat bones. The

specimens are thought to date to the Holocene (George 1999). The presence of packrat middens also implies ancient plant fossils, such as twigs and pollen, which could be useful for paleoecological reconstructions. Additional Quaternary finds from TICA include mammal bones at the Grotto rockshelter, found in 1939 by Tom Walker (although these may be of quite recent age, as the rockshelter was occupied by a mountain lion prior to 1880); packrat middens with bones in the Organ Pipe Room of Hansen Cave; and a partial bison skeleton found in gravel at the mouth of Swinging Bridge Cave (Santucci et al. 2001; Santucci and Kirkland 2010).



TICA Figure 6. Miscellaneous bone fragments from cave deposits (TICA 167) (photo courtesy Camille Pulham, TICA).

Younger cave bones are known from nearby American Fork Cave, 5 km (3 miles) up-canyon from TICA's headquarters. Here, mammal bones were found with a variety of human artifacts that indicate occupation sometime between 1,000 and 300 years ago. The animal assemblage includes gastropods, fish, birds, porcupines, woodchucks, prairie dogs, packrats (and packrat middens), bats, lynx, bears, martens, weasels, skunks, mountain sheep, and elk. Bone artifacts such as awls, beads and other ornaments, chisels, flakers, and gaming pieces were also found. Plants and plant-based materials recovered include a few corn kernels, wooden shafts, cords, and textiles (Hansen and Stokes 1941). If more precise ages were known for the TICA and American Fork Cave materials, they could make a paleoecological series for this area.



TICA Figure 7. A rodent skull from TICA (TICA 2887) (photo courtesy Camille Pulham, TICA).

There are a number of isolated Quaternary vertebrate finds associated with Lake Bonneville (Miller 1982; Nelson and Madsen 1980, 1987; Gillette and Miller 1999), but TICA's high elevation means its fossils are probably most comparable to the handful of other high-elevation late Quaternary vertebrate sites in the state, which have different assemblages than low-elevation sites (Nelson and Madsen 1987). The high-elevation sites include Blonquist Rockshelter, the Huntington sinkhole and Huntington Dam sites, Silver Creek Junction, and Stevens Creek (Smith et al. 1999).

Blonquist Rockshelter is from an undisclosed location in Summit County to the northeast. It has yielded thousands of fossils of plants, mollusks, insects, bony fish, amphibians, reptiles, birds, and mammals, with a fossil record extending from over 11,000 years ago to around 6,000 years ago. It is thought to be primarily a raptor deposit. Human artifacts are present in one layer, and packrat middens are also present (Nelson 1988; Nelson et al. 1989; Beasley 1993; Liggett and Nelson 1993).

The Huntington dam and sinkhole sites are associated with Huntington Reservoir, about 100 km (60 miles) south-southeast of TICA. The sinkhole site yielded some stems, woody fragments, and incomplete remains of two mastodons, which were the first mastodon bones found in Utah. Dates on the material suggest an early Holocene age (Miller 1987). Horse and bison bones have also been found at this site (Gillette and Miller 1999). The dam site has yielded a nearly complete skeleton of a Columbian mammoth with gut contents, and fragments of the short-faced bear *Arctodus*, along with pollen, leaves, wood, insects, dung, and a stone biface which may or may not be associated. The mammoth has been dated to near the Pleistocene–Holocene boundary. The relatively young age suggest that the high elevations may have had relict populations of mammals (Gillette and Madsen 1992, 1993).

The Silver Creek site is about 37 km (23 miles) northeast of TICA near Park City. It has the most diverse faunal assemblage of the high-elevation sites, with fossils of leopard frogs, mallards, teals, sage hens, water shrews, the sloth *Paramylodon*, pygmy rabbits, jackrabbits, ground squirrels, chipmunks, pocket gophers, deer mice, muskrats, heather voles, mountain voles, porcupines, the sabre-toothed cat *Smilodon*, lynx, dire wolves, coyotes, ermines, minks, badgers, Columbian mammoths, two species of horse, the camel *Camelops*, pronghorns, and *Bison* ?*latifrons*. The site is thought to represent flood deposits in a marshy area. While the age of the fossils is uncertain, they are thought to be at least 40,000 years old (Miller 1976).

The Stevens Creek site is over 150 km (95 miles) south-southeast of TICA near Ferron. It has yielded fossils including charcoal, gastropods, snakes, shrews, chipmunks, pocket gophers, voles, other rodents, and rabbits, and dates to the early Holocene (Smith et al. 1999).

As with vertebrate sites, paleoecological sites are not uncommon in the vicinity of TICA, but most are related to the Lake Bonneville/Great Salt Lake system. Well sites on Great Salt Lake have provided cores with pollen records going back at least 13.5 million years (Davis and Moutoux 1998). Packrat midden sites within about 100 km (60 miles) of TICA include several sites on Antelope Island, over 63 km (39 miles) to the northwest (Rhode 2000a), and Devils Gate, over 70 km (43 miles) to the west-southwest (Rhode 2000a, 2000b).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the following rock units within TICA. However, they are known to preserve fossils elsewhere, and future field investigations within the monument may recover fossils from one or more of them.

Mutual Formation (Neoproterozoic)

The Mutual Formation of the TICA area is composed of quartzite, conglomerate, shale, and grit. It is mostly red to purple-red in color, although the coarser quartzite is gray, tan, or greenish-gray, and finer-grained quartzite and shale are grayish-red. The formation is up to 400 m (1,300 ft) thick in the area, and is present in the northern part of the monument, including the base of American Fork Canyon (Baker and Crittenden 1961). Deposition is thought to have occurred between 610 and 580 Ma (Ehlers et al. 1997). The lower contact with the glacial rocks of the Mineral Fork Tillite is disconformable (Baker and Crittenden 1961), as is the upper contact with the Tintic Quartzite (Biek 2005). The Mutual Formation is interpreted as the deposits of a braided fluvial system, laid across a braidplain with local floodplain and lake deposits (Ehlers et al. 1997). Following deposition, there was some kind of tectonic activity, either a stand-alone period of continental rifting or a period of renewed extension from a rift that had begun before the deposition of the Mutual Formation (Christie-Blick 1997).

The Mutual Formation is not known for its fossils. Santucci and Kirkland (2010) reported trace fossils in the Mutual Formation of TICA, but this has not been corroborated.

Tintic Quartzite (lower–middle? Cambrian)

The Tintic Quartzite of the TICA area is nearly pure quartzite, usually white to pink in color when fresh but weathering to a rusty color. The lowest meter or so is often a conglomeratic layer, reflecting the disconformable contact with the Mutual Formation. The upper contact with the Ophir Formation is gradational or interbedded. The Tintic Quartzite is up to 400 m (1,300 ft) thick in the TICA area, and is the bedrock formation nearest the surface in much of northern TICA (Baker and Crittenden 1961). The gradational contact with the Ophir Formation may be on the order of 25 m (83 ft) thick (Perkins 1955).

The age of the Tintic Quartzite is not known precisely, in part because of the rarity of fossils. It is younger than 543 Ma (Christie-Blick 1997), and older than the Ophir Formation. Authors have described its age as early Cambrian (Perkins 1955; Yonkee et al. 2000; Biek 2005), middle Cambrian (Peterson and Clark 1974), and both (Ehlers et al. 1997). The formation is usually interpreted as shallow marine in origin. Hall (1981), using sections including one from TICA, interpreted the Tintic Quartzite as a shallow shelf unit, initially dominated by longshore currents, with the nearshore marine conditions becoming deeper over time (Hall 1981). The Tintic Quartzite has been compared to the roughly contemporaneous Tapeats Sandstone of the Grand Canyon region, which includes both shallow marine tidally-influenced deposits and braided fluvial deposits (Condon 1995).

Fossils are historically poorly known in the Tintic Quartzite (Ehlers et al. 1997), and have sometimes been declared absent (Gilluly 1932). To date, its fossil assemblage consists of trilobites (identified as *Olenellus*) and invertebrate trace fossils (Morris and Lovering 1961; Peterson and Clark 1974; Hall 1981). It is thought that turbulent water conditions may be responsible for the rarity of fossils (Hall 1981).

Ophir Formation (lower?–middle Cambrian)

The Ophir Formation of the TICA area includes three members. The lower member is a shale and siltstone unit, olive-green in color and 76 m (250 ft) thick (Baker and Crittenden 1961). In some areas nearby this member also has mica and quartzite (Bromfield and Patten 1981), indicating some metamorphism. The middle member is composed of pale gray limestone with layers of silty dolomite, and is 27 m (90 ft) thick. Finally, the

upper member is composed of intergrading limy sandstone and brown shale, and is 52 m (170 ft) thick (Baker and Crittenden 1961). It is present at TICA in a narrow band relatively low on the southern face of the American Fork Canyon (Baker and Crittenden 1961).

The Ophir Formation grades out of the underlying Tintic Quartzite and grades into the overlying Maxville Formation, although this upper contact is not always found due to erosion during the early and/or middle Paleozoic that removed the Maxville Formation and the upper to middle Ophir Formation in many areas (Baker and Crittenden 1961). The formation is typically regarded as largely to fully middle Cambrian in age, with possible lower Cambrian rocks at the base of the formation (Perkins 1955; Baker and Crittenden 1961; Biek 2005). It has been compared to the roughly contemporaneous Bright Angel Shale of the Grand Canyon area, which was deposited in a storm-influenced subtidal to intertidal environment (Condon 1995).

The Ophir Formation has a more abundant fossil assemblage than the Tintic Quartzite. To date, paleontological resources include oncolites (spherical layered structures formed by microorganisms) (Palmer 1970), the cyanobacteria fossil *Girvanella* (Morris and Lovering 1961; Rigo 1968), brachiopods, trilobites (Gilluly 1932; Rigo 1968), and invertebrate trace fossils (Baker and Crittenden 1961; Morris and Lovering 1961). Fossils are numerous but poorly preserved (Calkins and Butler 1943), and body fossils are often fragmentary (Baker and Crittenden 1961; Morris and Lovering 1961). Fossils appear to be most abundant in the lower member, followed by the middle member, with no explicit reference to fossils in the upper member. Trace fossils are abundant in the lower member, which also has fragments of brachiopods and trilobites (Baker and Crittenden 1961). The middle member has *Girvanella* and brachiopod and trilobite remains (Morris and Lovering 1961; Rigo 1968).

Maxfield Limestone (middle Cambrian)

The Maxfield Limestone (or Formation) of the TICA area is a dark gray to blue-gray limestone and dolomite unit. The base includes prominent oolitic and pisolitic layers; oolites and pisolites are round grains of concentrically-layered carbonates, with oolites smaller than 2 mm in diameter (a little larger than $1/16^{th}$ of an inch) and pisolites larger. The upper part is a distinct bed of white dolomite (Baker and Crittenden 1961). The formation can be up to 60 m (200 ft) thick, but is usually much less due to the Paleozoic erosion that also affected the Ophir Formation, as mentioned above. For example, in the 7.5-minute topographic quadrangle including TICA, only the lowest of several members is present (Baker and Crittenden 1961). The Maxfield Limestone is present on the south face of the American Fork Canyon at TICA (Baker and Crittenden 1961).

The Maxfield Limestone is dated to the middle Cambrian based on its fossils (Baker and Crittenden 1961). The upper contact with the Fitchville Formation is a disconformity (Biek 2005), with over 100 million years of geologic history missing between the two. The Maxfield Limestone is interpreted as a marine formation. It has been compared to the partially-correlative Muav Limestone of the Grand Canyon area, which ranges from outer shelf, to intertidal, to supratidal, to tidal flat depositional environments (Condon 1995).

Fossils are rare in the Maxfield Limestone (Anderson 1974). Its fossil assemblage includes *Girvanella* (Rigo 1968), brachiopods, trilobites (Calkins and Butler 1943), and possible invertebrate burrows (Anderson 1974). Fossils are known in this formation near TICA on the American Fork Canyon (Baker and Crittenden 1961).

Fitchville Formation (Upper Devonian–Lower Mississippian)

The Fitchville Formation (or Dolomite) of the TICA area is mostly made up of two thick dolomite layers, with the lower layer light to medium gray in color and the upper layer dark gray. The base of the formation is composed of dolomitic sandstone and sandy dolomite, weathering to a tan color. It is between 30 and 53 m (100 and 175 ft) thick (Baker and Crittenden 1961). Like the underlying Maxfield Limestone, it is present on the south face of the American Fork Canyon at TICA (Baker and Crittenden 1961). In some publications (Gilluly 1932; Calkins and Butler 1943), it is mistakenly identified as the Jefferson Dolomite (Baker and Crittenden 1961). It has also been included in a formation called the Gardner Dolomite (Perkins 1955), a now-abandoned name to which the overlying Gardison Limestone was also allocated (Morris and Lovering 1961; Greenhalgh 1980).

The Fitchville Formation dates from the Late Devonian to the Early Mississippian based on its fossils (late Famennian Age to Kinderhookian Age per Sandberg and Gutschick [1979], from approximately 365 to 350 Ma). At one time, it was thought to have a conformable contact with the overlying Gardison Limestone (Baker and Crittenden 1961), but studies of its fossils show that not only is the upper contact a disconformity (Sandberg and Gutschick 1979; Gosney 1982; Sando and Bamber 1984), but there is also an internal disconformity in the Fitchville Limestone at about the Devonian–Mississippian boundary (Sandberg and Poole 1977; Greenhalgh 1980; Sando and Bamber 1984). The Gardison Limestone occupies channels in the upper surface of the Fitchville Formation (Sandberg et al. 1979).

The Fitchville Formation is essentially a shallow marine unit, and was deposited during two marine transgression-regression cycles (Greenhalgh 1980). The lower Fitchville Formation was deposited during a Late Devonian regression in a narrow elongate marine basin, with water depths shallowing over time. The upper Fitchville Formation includes both a transgression and regression, eventually becoming shallow enough near the end of deposition to allow the widespread growth of

microorganisms in supratidal environments. This is reflected in the presence of stromatolitic limestone at the top of the formation (stromatolites are layered structures produced by microorganisms trapping sediment) (Greenhalgh 1980).

The fossil assemblage of the Fitchville Formation is dominated by invertebrates. Microorganisms are represented by stromatolites (Morris and Lovering 1961; Sandberg et al. 1979; Greenhalgh 1980). Invertebrates are represented by sponges, corals, bryozoans, brachiopods, gastropods, crinoids, echinoids, possible holothurians (sea cucumbers), and invertebrate trace fossils (Ausich 2003). Corals include rugose corals, such as horn corals (Gilluly 1932; Morris and Lovering 1961), and tabulate corals (Morris and Lovering 1961). Vertebrates are represented by conodonts, which have been extensively described (Sandberg and Poole 1977; Sandberg 1979; Sandberg and Gutschick 1979; Newman 1980; Gosney 1982). Some layers of the Fitchville Formation are made up mostly of fossil fragments (Morris and Lovering 1961), and the formation is often capped by a layer of stromatolitic limestone (the "curly limestone" of Morris and Lovering [1961] and Sandberg et al. [1979]). Fossils are sometimes altered by geologic processes (Anderson 1974), such as recrystallization of calcite to dolomite (Newman 1980). Fossils are known from this formation in and near the TICA area (Baker and Crittenden 1961).

Gardison Limestone (Lower Mississippian)

The Gardison Limestone of the TICA area includes three divisions. The basal 3 to 9 m (10 to 30 ft) is composed of dark gray dolomite, the middle 24 to 30 m (80 to 100 ft) is composed of highly fossiliferous blue-gray limestone with silty partings, and the upper 120 m (400 ft) or so are composed of dark gray limestone and dolomite with chert. Altogether, the formation is about 180 m (600 ft) thick. It is found in TICA on the south face of the American Fork Canyon, just below the Deseret Limestone that includes the monument's caves (Baker and Crittenden 1961). The Gardison Limestone was formerly identified with the Madison Limestone (Gilluly 1932; Calkins and Butler 1943; Baker and Crittenden 1961; Baker 1964), a formation from the north and east. It was also included in the Gardner Formation (Perkins 1955; Morris and Lovering 1961; Greenhalgh 1980).

The Gardison Limestone was deposited during the Early Mississippian (Early Osagean Age per Sandberg and Gutschick [1979] and Sandberg et al. [1979], approximately 350 Ma). The upper contact with the Deseret Limestone is conformable (Perkins 1955), with the upper 5 m (16 ft) or less of the Gardison Limestone a dark limestone that transitions into the basal member of the Deseret Limestone (Sandberg et al. 1979). The Gardison Limestone mostly represents shallow water deposition, from rapid accumulations of invertebrate fragments on a carbonate platform or shelf margin. The upper transitional beds represent deeper water, more like the conditions of the Deseret Limestone (Sandberg et al. 1979). Deposition occurred during part of a marine transgression (Greenhalgh 1980).

The Gardison Limestone is noted for its abundant fossils (Calkins and Butler 1943; Baker 1964; Anderson 1974; Sandberg et al. 1979), with some beds made up almost entirely of crinoid fragments (encrinite) (Sandberg et al. 1979). The formation's fossil assemblage includes corals, bryozoans, brachiopods (Greenhalgh 1980), bivalves (Perkins 1955), nautiloid cephalopods, gastropods, ostracodes, crinoids (Morris and Lovering 1961), and conodonts (Sandberg 1979; Sandberg and Gutschick 1979). In some cases, geologic processes have resulted in fossil dissolution, producing "ghosts" of fossils (Anderson 1974).

Humbug Formation (Middle Mississippian)

The Humbug Formation of the TICA area is composed of alternating beds of tawny-colored fine- to medium-grained lime-rich sandstone and limestone or dolomite. It is the formation stratigraphically and spatially above the cave-forming Deseret Limestone at TICA, and is up to 240 m (800 ft) thick (Baker and Crittenden 1961). Beds are not continuous in this formation, commonly interfingering and pinching out (Livingston 1955; Perkins 1955). The variation of rock type gives this formation a profile of ledges and slopes (Livingston 1955). The Uncle Joe Member of the underlying Deseret Limestone is sometimes confused with the Humbug Formation (Morris and Lovering 1961).

The age of the formation is not as well-constrained as for the adjacent formations, due to the relative scarcity of fossils. Based on the ages of the Deseret Limestone and overlying Great Blue Limestone, the Humbug Formation dates to the Middle Mississippian (middle or late Meramecian Age per Sandberg et al. [1979], approximately 335 Ma). The upper contact with the Great Blue Limestone is conformable to gradational (Biek 2005). The Humbug Formation is a shallow marine unit, with water depths thought to have been less than 30 m (100 ft). Deposition occurred on a carbonate bank, and strong currents dominated, with quiet intervals. Clastic sediment arrived from the west. The water temperature was probably near 20 °C (68 °F), and salinity was probably close to normal marine values (Livingston 1955).

Fossils are uncommon in the Humbug Formation (Sandberg et al. 1979), although it is fossiliferous in American Fork Canyon, and the limestone beds often include skeletal fragments (Livingston 1955). The fossil assemblage of the formation includes foraminifera, possible microbial structures (Livingston 1955), rugose and tabulate corals (Gilluly 1932), bryozoans, brachiopods (Gilluly 1932; Calkins and Butler 1943; Livingston 1955), gastropods, trilobites (Calkins and Butler 1943; Livingston 1955), blastoids (sea buds), crinoids, echinoids (Gilluly 1932), and possible invertebrate fecal pellets (Livingston 1955).

Great Blue Limestone (Middle–Upper Mississippian)

The Great Blue Limestone of the TICA area is mostly composed of thin-bedded shale-rich limestone, dark gray in color when unweathered but weathering to light gray or pinkish tan. At the base of the formation is a black shale layer. The formation is 850 m (2,800 ft) thick. It is found in southern TICA, in the highest elevations of the monument (Baker and Crittenden 1961). It has been divided into a number of members at other locations (Morris and Lovering 1961; Sandberg et al. 1979; Gordon et al. 2000), but this has not been done in the TICA area, where the formation is nearly homogeneous (Olsen 1955). Perhaps the most widely used of these members is the Long Trail Shale Member, a shaly unit in the middle of the formation that has often been distorted by tectonic activity, as it is less resistant to forces than other units (Perkins 1955).

The Great Blue Limestone dates to the Middle and Late Mississippian (latest Meramecian Age to Chesterian Age per Gordon et al. [2000], between approximately 335 and 320 Ma). Deposition occurred under an oscillating continental sea. Depositional environments included open marine conditions, local reefs or banks of mostly solitary corals and associated zones of skeletal fragment accumulations, and lagoons (Lindsay 1977). On the west slope of Mount Timpanogos, the formation is interpreted as nearshore and shallow water, perhaps even littoral, with the occasional beds of dark shale possibly representing coastal marshes. The alternation of environments may be responsible for the sparse fossil assemblage in this area (Olsen 1955). Sediments arrived from the west (Chamberlain 1981), from young highlands (Greenhalgh 1980).

The Great Blue Limestone has locally abundant fossils (Baker 1964), which have been used to examine depositional environments (Lindsay 1977). Microorganisms and plants are represented by foraminifera, structures from algae and other microorganisms (Lindsay 1977), and terrestrial plants, including *Lepidodendron*, lycopsids, and the horsetail *Calamites* (Chamberlain 1981). The invertebrate assemblage includes sponges, corals (Chamberlain 1981) (both colonial and solitary forms; Lindsay [1977]), hydrozoans (relatives of corals and jellyfish) (Zeller 1958), bryozoans, brachiopods, bivalves (Chamberlain 1981), ammonite and nautiloid cephalopods (Zeller 1958), gastropods, trilobites (Chamberlain 1981), ostracodes (Zeller 1958), blastoids (Olsen 1955), crinoids, echinoids (Zeller 1958), and trace fossils including burrows and pellets (Lindsay 1977). Finally, the vertebrate assemblage includes conodonts (Bissell and Barker 1977; Chamberlain 1981), sharks (Miller 1981), and fish spines (Chamberlain 1981). Bryozoans, brachiopods, and blastoids have been recovered from low in the formation on the west slope of Mount Timpanogos (Olsen 1955).

Park Collections

TICA has 196 cataloged paleontological objects, eight of which have been deaccessioned (C. Pulham, TICA chief of resource management, pers. comm., September 2010). Examples of invertebrates from the Deseret Limestone found in TICA are in monument collections. Such fossils include colonial corals, solitary horn corals, crinoid stems, and coquinoid rocks (Santucci et al. 2001). The bones collected by George (George 1999) are also in monument collections (C. Pulham, pers. comm., October 2010).

Cultural Resource Contexts

The potential exists for fossils to be found at TICA with some sort of cultural association. As discussed, American Fork Cave just up the canyon has recent bones in association with artifacts (Hansen and Stokes 1941), providing a good example for possibilities at TICA. Ancient humans are also associated with the Blonquist Rockshelter (Liggett and Nelson 1993) and Huntington Dam sites (Gillette and Madsen 1993). Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

Paleontological Resource Management, Preliminary Recommendations

- The NPS Geologic Resources Division coordinated a geologic resource inventory scoping session for TICA during May 1999 (NPS 1999), and produced a geologic resource evaluation report for the monument in 2006 (Thornberry-Ehrlich 2006). The participants of the scoping meetings developed a series of recommendations for future work, including: performing a comprehensive inventory of TICA paleontological resources; inventorying all paleontological specimens collected in the monument; determining the locations of paleontological specimens removed from TICA; interpreting TICA's fossil resources for visitors; and using fossils to examine paleoclimate. A Geologic Resource Monitoring

Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009).

- Further research should be conducted on the cave fossils, including dating and paleoecological studies. Comparison of TICA fossils to the younger bones from American Fork Cave is another potential line of investigation.

- Monument staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, monument staff should take advantage of them.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the monument and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Timpanogos Cave National Monument Digital Geologic Map
2006–present. (geologic map data; electronic files; GIS metadata; associated report). Originated by NPS-Geologic Resources Inventory staff; status: Active. Online. https://irma.nps.gov/App/Reference/Profile?Code=2164812.

Paleontological Assessment of Pleistocene and Holocene Deposits of Timpanogos Cave National Monument
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Zion National Park

Zion National Park (ZION), Utah, is renowned for its scenic geology (ZION Figure 1). Within the park's boundaries erosion has sculpted colorful sedimentary rock formations into canyons, cliffs, and mesas. ZION began as Mukuntuweap National Monument, which was proclaimed July 31, 1909. Its name was changed to Zion National Monument on March 18, 1918, and it was then redesignated Zion National Park on November 19, 1919. A separate Zion National Monument, which included the Kolob Canyons District, existed between January 22, 1937 and July 11, 1956, when it was incorporated into ZION. ZION's boundaries have changed several other times, including: June 13, 1930; June 3, 1941; February 20, 1960; October 21, 1976; and November 12, 1996. It currently encompasses 59,325.94 ha (146,597.60 acres), all but 1,426.21 ha (3,524.24 acres) of which are under federal administration (ZION Figure 2).

Geologic Background

ZION is located in southwestern Utah on the southern slopes of the Markagunt Plateau. Almost all of it is in Washington County, with small sections in southern Iron County and western Kane County. Most of the park's land forms a roughly rectangular parcel, with a northwestern extension that includes the Kolob Canyons District. ZION is in the western Colorado Plateau physiographic province, near the transition with the Basin and Range province on the west (Biek et al. 2010). The Hurricane fault zone, which passes through the western Kolob Canyons District, is the local boundary between the two provinces (Hatfield et al. 2010). The Hurricane Cliffs, produced by the fault, run through part of the district. ZION is dissected by several large canyons, the most prominent of which is Zion Canyon, cut by the North Fork of the Virgin River. Parunuweap Canyon is found in southeastern ZION, formed by the East Fork of the Virgin River. The Left Fork and Right Fork of North Creek drain western ZION. The Markagunt Plateau includes part of northern ZION. Cedar City is about 20 km (12 miles) north of the northern end of ZION, and St. George is centered about 42 km (26 miles) southwest of the park.



ZION Figure 1. A ZION landform composed of Navajo Sandstone, which forms many of the park's features (VLS).

A geologic scoping session coordinated by the Geologic Resources Division of the National Park Service (NPS) was held for ZION during April 1999 (NPS 1999), leading to a geologic report for the park (Graham 2006). The initial paleontological resource inventory and summary was prepared by Koch and Santucci (2002). A paleontological survey was prepared for the park by DeBlieux et al. (2005), with a follow-up dedicated to monitoring issued in 2008 (Markle 2008).

ZION has long been a popular location for geologic research, which has resulted in numerous publications that describe the geology of the park and vicinity. References that are useful for understanding the general geology of ZION include: Gregory (1947, 1950a); Gregory and Williams (1947); Hamilton (1984, 1995); Biek et al. (2004a, 2004b); Eves (2005); Graham (2006); and Biek et al. (2010). Geologic maps of ZION or its individual quadrangles include: Gregory (1950a); Hamilton (1978); Averitt (1967); Sable and Doelling (1993); Doelling (2002); Doelling et al. (2002); Willis and Hylland (2002); Willis et al. (2002); Graham (2006); Biek (2007a, 2007b); Biek et al. (2009, 2010); Hylland (2010), and Hayden (in prep.). References with particular bearing on the paleontology of ZION include: Hesse (1935); Stokes and Bruhn (1960); Hamilton (1979); Hevly (1979); Painter and Taylor (1998); Santucci et al. (1998); Smith and Santucci (1999, 2001); Santucci (2000); Koch and Santucci (2002); Smith et al. (2002); DeBlieux and Kirkland (2003); DeBlieux et al. (2003, 2005, 2006); Heckert et al. (2006); Mickelson et al. (2006); Staker (2006); Markle (2008); and Santucci and Kirkland (2010).

The geologic record exposed at ZION covers the Lower Permian to the Upper Cretaceous, and the Quaternary (see the appendix for a geologic time scale). Its geologic record fits between that of Grand Canyon National Park (GRCA) to the south and Bryce Canyon National Park (BRCA) to the northeast (Gregory 1947), the three forming an impressive display of geologic history. The lower Permian rocks of ZION and its surroundings derive from shallow marine and coastal settings as a continental sea transgressed (advanced over the land) and regressed (retreated) from the west and northwest. The two Permian formations present within or just outside of ZION (the Toroweap and Kaibab formations) each record one major transgression-regression cycle. At the time, southern Utah was near the equator (Biek et al. 2010).

A gap of some 10 million years follows at ZION. A mountain-building event known as the Sonoma Orogeny created a basin in southwestern Utah and adjacent areas, where the Moenkopi Formation was deposited (Stewart et al. 1972a; Biek 2010). Triassic rocks at ZION are principally represented by the Moenkopi Formation of the Early Triassic and the Chinle Formation of the Late Triassic. Southern Utah during the Early and Middle Triassic was a broad coastal plain that was occasionally flooded by a sea to the west (Stewart et al. 1972a; Blakey et al. 1993; Hintze and Kowallis 2009). The top of the Moenkopi Formation is separated from the Chinle Formation by millions of years, with intervening erosion (Pipiringos and O'Sullivan 1978). The Chinle climate is thought to have been seasonal, with wet and dry periods (Lucas 1993). Fluvial and lacustrine deposition alternated over time (Blakey and Gubitosa 1983). Distant volcanic eruptions occurred to the west and south during deposition; these eruptions contributed substantial amounts of volcanic sediment to the Chinle Formation (Riggs et al. 2003). Highlands were present west of ZION in eastern Nevada (Blakey 1989).

As the deposition of the Chinle Formation ended, aridity increased. During its deposition, the Four Corners area of the Southwest moved from near the equator to arid-zone latitudes, and aridification began (Tanner 2003). The beginning of the Jurassic featured the shallow lacustrine system of the Moenave Formation in southwestern Utah, with an erg (giant dune field) to the east (Tanner and Lucas 2007, 2009a). A brief period of increased humidity resulted in the deposition of the Kayenta Formation, on a broad, gentle floodplain separated from the sea to the west (Blakey 1994; Peterson 1994). About 190 Ma (million years ago), this setting was gradually replaced by the vast erg that became the Navajo Sandstone (Harshbarger et al. 1957; Dickinson and Gehrels 2003, 2009; Rahl et al. 2003; Lucas et al. 2005), the largest erg known from North America (Blakey 2008; Blakey and Ranney 2008). This Sahara-like desert was found throughout the Four Corners area into Nevada and Wyoming (Hintze and Kowallis 2009). At this time, the latitude of the area was only about 10° N, near the western edge of the supercontinent Pangaea (Loope et al. 2004a). Strong cross-equatorial winds blew from the northwest during the winter, driving dune movement, and monsoonal rains occurred during the summer (Loope et al. 2004a, 2004b).

Eventually another shallow sea began to encroach on the ZION area, inundating the Navajo erg and setting the stage for another series of transgression-regression cycles and their resulting shallow marine and coastal terrestrial deposition (Biek et al. 2010). The shallow sea, known as the Carmel–Twin Creek Seaway (Tang et al. 2000), moved into the area from the north (Imlay 1980), reaching as far south as northern Arizona. The Carmel–Twin Creek Seaway was active at about the same time as an outbreak of volcanic activity on the southern margin of the Colorado Plateau dated to approximately 170 Ma (Blakey and Parnell 1995); minor crustal extension also occurred to the northeast in central Utah (Blakey 1989; Marzolf 1993). Events following the Middle Jurassic are poorly represented by rock units at ZION, with only a single area of Cretaceous rocks in the Kolob Canyons District (Biek et al. 2009, 2010), but are well known to the north (Biek et al. 2011) and east (Sable and Hereford 2004).

The Sevier Orogeny began during the Jurassic in response to the collision of the North American and Farallon plates along the western margin of North America (Biek et al. 2010), and persisted into the Paleogene (Hintze and Kowallis 2009). During this orogeny, thrust faulting began moving east from the margin of the continent into Utah, with the Kanarra anticline (concave-down fold) an example of the orogeny's effects in the vicinity of ZION (Biek et al. 2010). Terrestrial processes continued in the area until about 100 Ma, when a shallow continental sea called the Western Interior Seaway or Cretaceous Interior Seaway began to encroach upon the central United States. The Dakota Formation signals the onset of several cycles when the Western Interior Seaway transgressed and regressed over the land (Gustason 1989; Aubrey 1991; Ulicny 1999). The Seaway extended from the Arctic to the Gulf of Mexico, bisecting North America (Elder and Kirkland 1993, 1994). Its waxings and wanings influenced the geology of numerous NPS units both within and outside of the Northern Colorado Plateau Network (NCPN). Rocks deposited by the seaway are not present in ZION, but can be found just outside of it to the east (Graham 2006; Biek et al. 2009, 2010). Local formations dating from the Late Cretaceous through to the Neogene are widely exposed at BRCA and Cedar Breaks National Monument (CEBR) (Biek et al. 2011, in prep.).

Throughout much of western North America, the final withdrawal of the Seaway happened concurrently with the beginning of the Laramide Orogeny (Cather 2003). This mountain-building event took place from the Late Cretaceous to the Eocene and was responsible for much of the present form of the Rocky Mountains (Aubrey 1991), although is not much in evidence at ZION, where the Sevier Orogeny and recent seismic activity were more influential (Grant 1987; Biek et al. 2009, 2010). The Colorado Plateau was less deformed in comparison, instead forming a series of broad uplifts and monoclines (step-like folds) (Tindall 2000; Davis and Bump 2009). Seismic activity has persisted at ZION, which is near the eastern margin of Basin and Range faulting (Biek et al. 2010). Several faults have developed in and around ZION, most notably the Hurricane fault zone, which has been active for at least a few million years (Lund et al. 2007). Movement on the Hurricane fault zone increased the gradient of the Virgin River system in the future ZION area and promoted downcutting, which formed the park's canyons (Willis and Biek 2001; Biek et al. 2010).

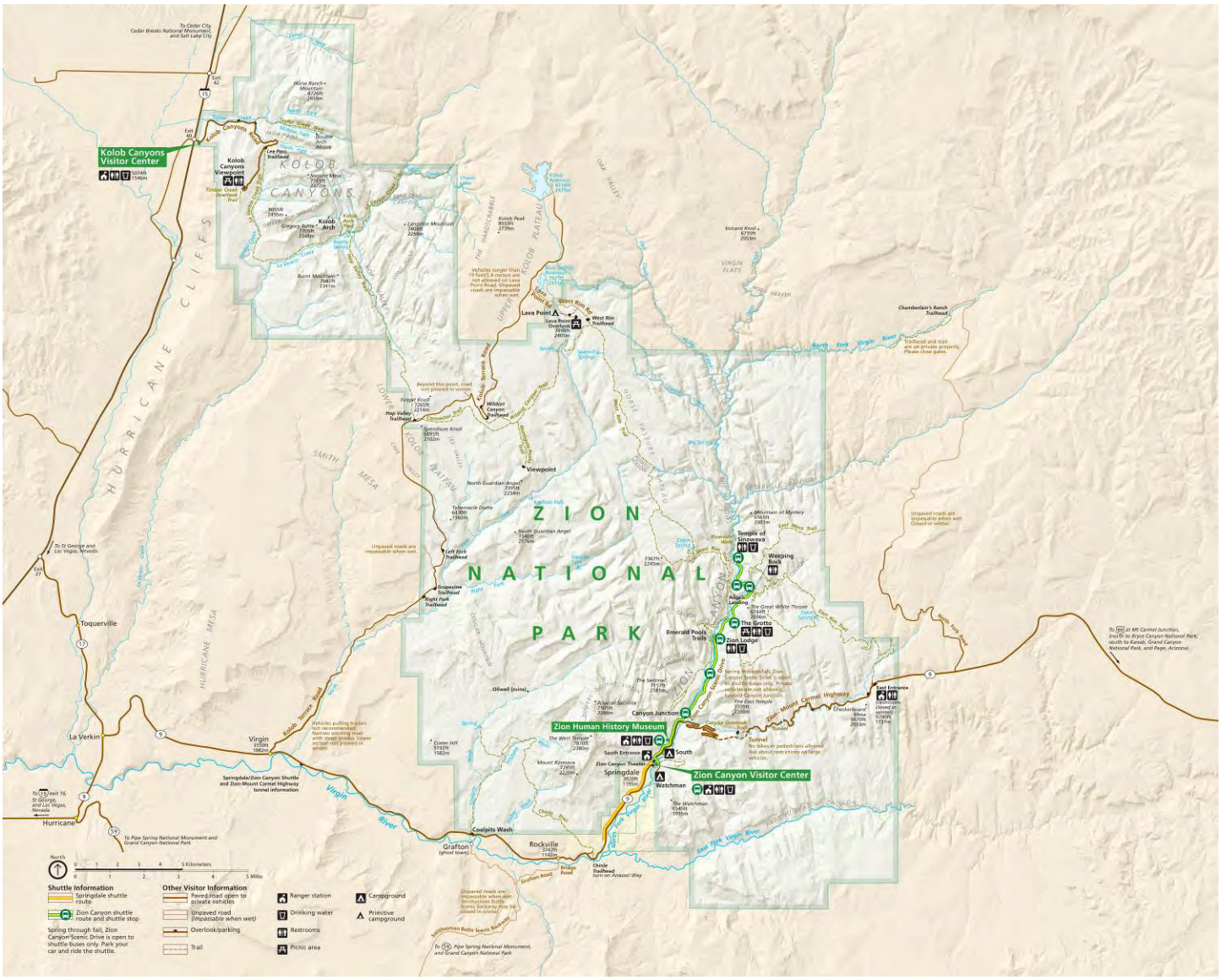
In addition to periods of uplift, the Colorado Plateau, especially along its western margin, underwent local igneous activity during the late Eocene, Oligocene, and Miocene. A local example of igneous activity from this time is the Pine Valley Mountains west of ZION, formed by an intruding laccolith at

about 21 Ma. About 17 Ma, igneous activity transitioned from explosive andesitic eruptions to smaller bimodal basaltic and rhyolitic eruptions due to major tectonic changes at the western margin of North America (Rowley et al. 1979). The youngest volcanic rocks in and near ZION are basaltic lava flows that coincide with the northern end of the Western Grand Canyon basaltic field. Rocks from six volcanic flows dated between 1.44 Ma and 125,000 years ago can be found in ZION, with three cinder cones within park boundaries (Biek et al. 2010).

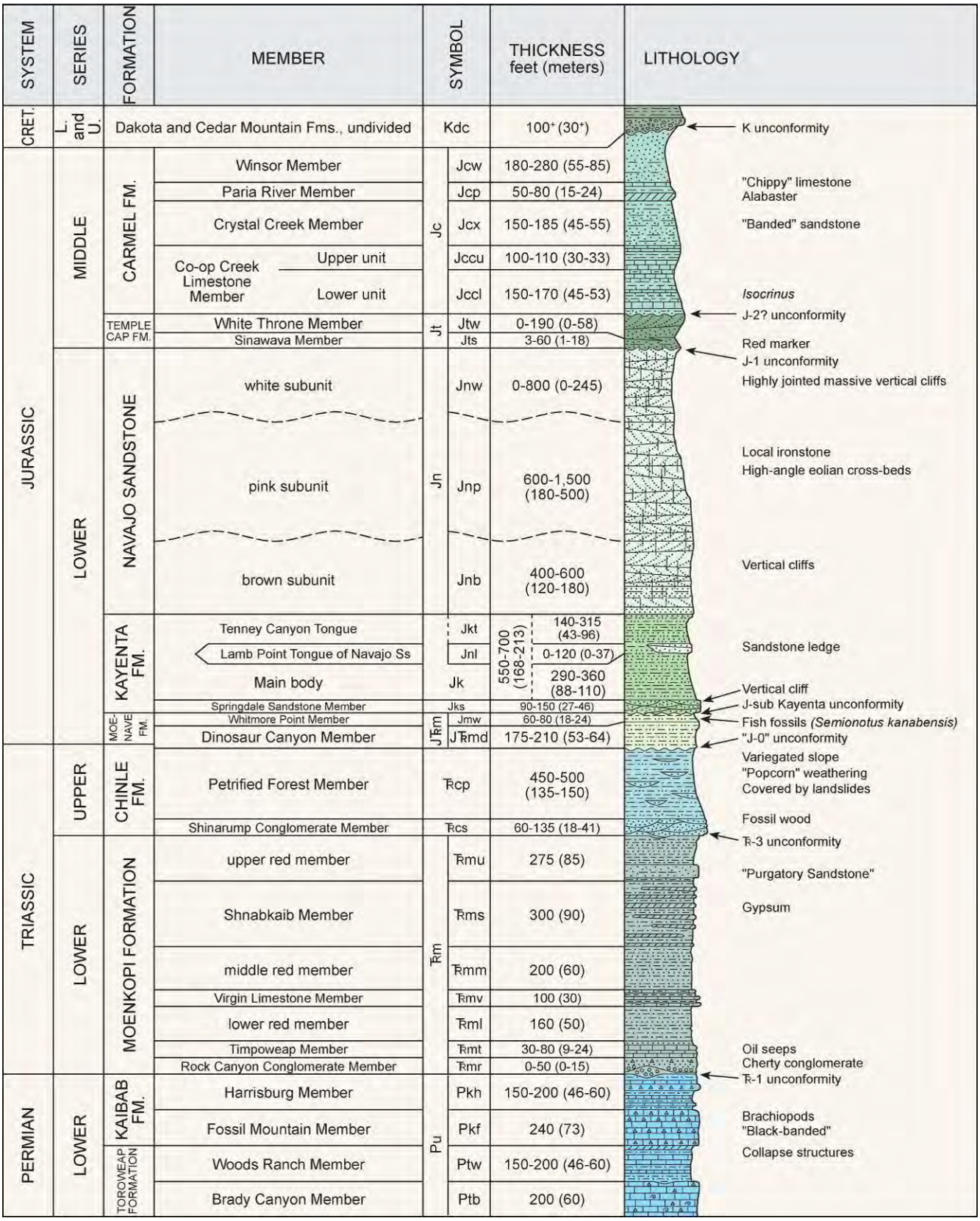
Strata of Cretaceous and Tertiary age were stripped from ZION by erosion, except for the aforementioned outlier of Upper Cretaceous beds at Horse Ranch Mountain in the Kolob section. Because ZION is above all else a monument to erosion, even the Quaternary Period is sparsely represented at ZION. The Pleistocene is famous for its glaciation, but glaciation did not directly affect ZION. Instead, the park was a site of local basaltic volcanism, as mentioned above, and small lakes. At least fourteen lakes are known to have existed in ZION at various times during the Quaternary, formed in response to the damming of drainage by lava flows or landslides (Biek et al. 2010). The Colorado Plateau was cooler and more moist during the last glacial period, becoming more arid afterward. It has been most arid during the last 1,500 years (Elias et al. 1992).

Bedrock units exposed at ZION include, from oldest to youngest: the Kaibab Formation (Lower Permian); the Moenkopi Formation (Lower Triassic); the Chinle Formation (Upper Triassic); the Moenave Formation, Kayenta Formation, and Navajo Sandstone of the Glen Canyon Group (Upper Triassic–Lower Jurassic); the Temple Cap Formation and Carmel Formation of the San Rafael Group (Middle Jurassic); a poorly exposed Cretaceous interval that includes the Cedar Mountain Formation and the lowest part of the Dakota Formation (Lower–Upper Cretaceous); and Quaternary volcanic flows. Surficial units include a variety of unconsolidated Quaternary deposits (Pleistocene–Holocene). In general, the park's exposed bedrock becomes older going east to west approaching the Hurricane fault zone, with broad areas of exposed Carmel Formation rocks along the north and east boundaries and Navajo Sandstone rocks exposed in most of the central part of the park. Permian and Triassic rocks are exposed in the southwestern and northwestern parts of the park. Cretaceous rocks are only found on Horse Ranch Mountain in the Kolob Canyons District. Several cinder cones and associated basaltic lava flows are located along a general north-south line in west-central ZION (see ZION Figure 3 for a stratigraphic column and ZION Figure 4 for a geological map) (Gregory 1950a; Hamilton 1978; Graham 2006; Biek et al. 2010). The Springdale Sandstone Member was assigned to the Moenave Formation in the park's geologic inventory (Graham 2006), but it has since been transferred to the Kayenta Formation (Lucas and Tanner 2006; Biek et al. 2010). It will be discussed within its own subsection between the Moenave Formation and Kayenta Formation, to permit easier referencing. DeBlieux et al. (2005), Graham (2006), and Biek et al. (2010) also discussed the Toroweap Formation, a Lower Permian unit underlying the Kaibab Formation. The Toroweap Formation is found just outside of ZION but not within it, and so will not be discussed here except in passing.

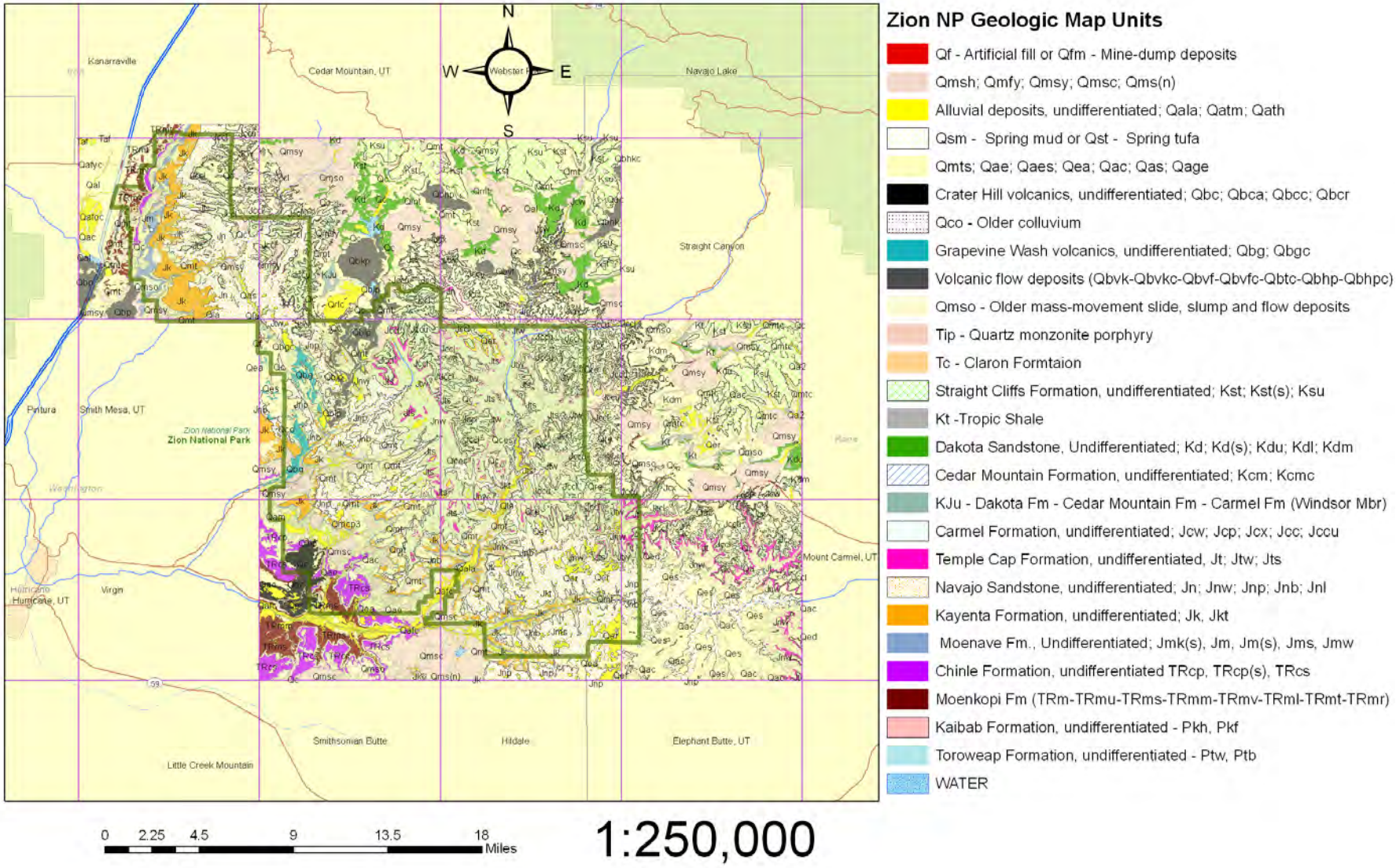
All of these rocks units, with the exception of the Quaternary igneous rocks, are potentially fossiliferous, and most of them have produced fossils within ZION. The only exception at this time is the Temple Cap Formation. In addition, fossiliferous rocks eroded from formations not exposed in ZION have been found in the park. Paleontological resources at ZION are known from the Kaibab Formation (marine invertebrates), the Moenkopi Formation (marine invertebrates and terrestrial trace fossils), the Chinle Formation (petrified wood and other plant fossils, bone fragments, trace fossils), the Moenave Formation (petrified wood and other plant fossils, bone fragments, trace fossils), the Kayenta Formation (tetrapod tracks), the Navajo Sandstone (dinosaur tracks), the Carmel Formation (marine invertebrates), the Cretaceous rocks (plant impressions and bivalves), and Quaternary sediments (pollen, wood and stumps, mollusks, invertebrate trace fossils, and rare vertebrate body and trace fossils) (ZION Table 1). ZION has some museum specimens, and also has what may be petroglyph representations of dinosaur footprints.



ZION Figure 2. Geography of ZION (NPS).



ZION Figure 3. Stratigraphic column for ZION (from Biek et al. [2010], used with permission of the Utah Geological Association).



ZION Figure 4. Schematic geological map of ZION. Map graphic adapted from digital geologic map data available at the following URL: https://irma.nps.gov/App/Reference/Profile?Code=1041895.

Formation	Age	Fossils Within ZION	Depositional Environment
Quaternary sediments	Pleistocene–Holocene	Diatoms, pollen, spores, peat, foliage impressions, stumps, carbonized roots and stems, bivalves, gastropods, invertebrate traces, a fish vertebra, a bison vertebra, other isolated mammal bones, possible bird and gopher bones, and bird and artiodactyl tracks	Alluvial, eolian, fluvial, lacustrine, landslide, and talus deposits
Quaternary volcanic flows	Pleistocene	None to date; fossils are unlikely but not impossible	Volcanic flows
Fossiliferous eroded rocks	Not applicable; originally deposited from the Late Cretaceous to the Middle Eocene	A Late Cretaceous bivalve from the Straight Cliffs Formation and rock fragments with stromatolites from the Paleocene–Middle Eocene Claron Formation	Not applicable; originally diverse terrestrial and marine coastal settings for the Straight Cliffs Formation, and fluvial, alluvial, and lacustrine settings for the Claron Formation
Cedar Mountain Formation–Dakota Formation	Early–Late Cretaceous	Plant impressions and freshwater bivalves	Fluvial, floodplain, and lacustrine settings, then becoming shallow marine over time
Carmel Formation	Middle Jurassic	Microbial structures, bivalves, gastropods, and crinoids (especially star-shaped columnals of the crinoid *Isocrinus*)	Shallow marine to coastal mudflat settings, changing over time because of several transgression-regression cycles
Temple Cap Formation	Middle Jurassic	None to date; palynomorphs are most likely	Heterogeneous, changing from shallow marine to tidal or salt flat to eolian deposition
Navajo Sandstone	Early Jurassic	Wood and dinosaur tracks	Desert with large eolian sand dunes
Kayenta Formation	Early Jurassic	Petrified wood and other plant fragments, root traces, invertebrate burrows, and dinosaur tracks (Springdale Sandstone Member); fish scales and tracks of dinosaurs and mammal relatives (overlying)	Fluvial, floodplain, and eolian settings
Moenave Formation	Late Triassic–Early Jurassic	Microbial structures, plant fragments, rare petrified wood, mollusks, invertebrate burrows, fish remains (scales, bones, and partial skeletons), and dinosaur tracks	Fluvial, floodplain, lacustrine, and eolian settings
Chinle Formation	Late Triassic	Petrified wood and other plant fragments, bivalves, invertebrate burrows, fragmentary vertebrate remains representing fish, metoposaurs, aetosaurs, phytosaurs, and others, coprolites, and reworked fossils	Fluvial, floodplain, lacustrine, and eolian settings, becoming drier over time
Moenkopi Formation	Early Triassic	Brachiopods, bivalves, gastropods, asteroids, crinoids, invertebrate traces, tetrapod tracks, and reworked fossils	Coastal plain with occasional marine incursions

Formation	Age	Fossils Within ZION	Depositional Environment
Kaibab Formation	Early Permian	Bryozoans, bivalves, and crinoids	Shallow marine to coastal plain settings

ZION Table 1. Concise tabulation of ZION stratigraphy, fossils, and depositional setting; details and references can be found in the text. Igneous rocks are included here for completeness, but are not discussed in detail in the text because of the improbability of fossils.

Sources of Paleontological Resources

The fossils of ZION present opportunities for education, interpretation, and continued or future scientific research in the park.

The first described fossil from ZION was a partial fish skeleton. This specimen, now at the University of California's Museum of Paleontology, was collected by R. A. Brown approximately "100 yards northwest of Pine Creek Bridge" (Hesse 1935; the outcrops are actually east [D. Sharrow, ZION hydrologist, pers. comm., May 2011]). It was discovered while a road was being built from Zion Canyon to Zion Tunnel, and was damaged by blasting. It was described by Hesse (1935) as an example of *Semionotus* cf. *gigas*. Hesse's fish specimen was originally thought to have come from the Chinle Formation, but is now known to have come from the Whitmore Point Member of the Moenave Formation. The site, which would now be next to Highway 9 near a bridge over Pine Creek, has been sought but not relocated (DeBlieux et al. 2006). The year 1935 was also the year of the first report of trace fossils from the park. The park's internal "Zion–Bryce Nature Notes" mentioned the presence of dinosaur tracks near the highway at the base of the Mountain of the Sun. The highway was moved later that year and the tracks have not been relocated (Smith and Santucci 2001).

After 1935, the fossil resources of ZION were generally mentioned only in passing for over 60 years, with the exception of a dinosaur track site described by Stokes and Bruhn (1960) and the Quaternary lake fossils documented by Hamilton (1979) and Hevly (1979). Interest in fossils at ZION was renewed when in 1997 Southern Utah University students Aimee Painter and Rex Taylor relocated several fossil sites in the park. This was the impetus for a series of field inventories, in 1999, 2000, 2002, and 2003. These inventories have been collaborative enterprises, with personnel from the Utah Geological Survey and volunteers from the Utah Friends of Paleontology assisting NPS interns. As a result of this work, over 120 new fossil sites have been found in ZION (DeBlieux et al. 2005, 2006). Finds of note from these efforts include tracks that may show gregarious theropod dinosaurs (Smith et al. 2002), an unusual reptile tooth (Heckert et al. 2006), and what may be the oldest Mesozoic vertebrate tracks in North America (Mickelson et al. 2006).

Vulnerability of fossils to erosion

The same spectacular rate of erosion in ZION that exposes rock outcrops and fossils is also a threat to the existence of some of those fossils. The regional rate of erosion has been estimated at 400 m (1,300 ft) per million years based on the extent of exposure of an inverted valley with a dated basalt cap (Willis and Biek 2001; Biek et al. 2010). There have been two instances in the past decade where trace fossils have been inventoried on the undersides of exposed cliff ledges, only for it to be discovered on a subsequent visit that the ledge had collapsed. The first was documented in the Kayenta Formation by DeBlieux et al. (2006) in 2002 and found to have fallen in 2008 (Markle 2008). The second was discovered in the Springdale Sandstone by Markle (2008) and found by park staff to have collapsed in 2010 (D. Sharrow, pers. comm., May 2011).

Kaibab Formation (Lower Permian)

The Kaibab Formation is a heterogeneous formation composed mostly of limestone and dolomite (Blakey and Knepp 1989). Early workers divided it into the Gamma, Beta, and Alpha members (McKee 1938), which have since been subsumed into the Fossil Mountain and overlying Harrisburg members. The Fossil Mountain Member is the equivalent of the Gamma and Beta members, and the Harrisburg Member is the equivalent of the Alpha Member (Blakey and Knepp 1989). Both the Fossil Mountain Member and the Harrisburg Member are present in the ZION area, along the Hurricane Cliffs (Hamilton 1978; Biek 2007a). The Fossil Mountain Member is a cliff-forming gray limestone and cherty limestone, up to 73 m (240 ft) thick (Biek et al. 2010). The black chert produces black bands on the cliff faces. The Harrisburg Member is composed of gray limestone, cherty limestone, and siltstone, and forms a series of ledgy slopes; it is gypsiferous in the greater St. George area (Biek et al. 2009). It is 46 to 60 m (150 to 200 ft) thick at ZION (Biek 2007a). The Kaibab Formation dates to the late Early Permian (Spamer 1984). The upper contact with the Moenkopi Formation is a prominent unconformity (Nielson 1991; Biek et al. 2009).

The Kaibab Formation formed in a variety of settings, from shallow marine to salt flat to coastal plain. The sea transgressed from west to east, transgressing and regressing several times during deposition of the Fossil Mountain Member, and then regressing during deposition of the Harrisburg Member (Blakey and Knepp 1989). The Fossil Mountain Member of ZION is interpreted as a shallow marine setting, and the Harrisburg Member is interpreted as a shallow marine to coastal sabkha (salt flat) setting (Biek et al. 2010).

Fossils are abundant in the Kaibab Formation, but the exposures in ZION are small (DeBlieux et al. 2005). Bryozoans (moss animals), bivalves, and crinoids (sea lilies) have been found in the upper Fossil Mountain Member (Kaibab Beta) either within or just outside of northwestern ZION at Section 10, T38S, R12W (Stewart et al. 1972a). Fossils in the Kaibab Formation of the vicinity, particularly the Hurricane Cliffs, include sponges, corals, bryozoans, brachiopods (lamp shells), mollusks, crinoids, echinoids (sea urchins), and conodonts (extinct eel-like chordates) (Gregory and Williams 1947; Hamilton 1984; DeBlieux et al. 2005; Biek 2007a); a taxonomic list for the ZION area is included in Gregory (1950a). Several recent reports (for example DeBlieux et al. [2005]) appear to be mistaken in citing McKee (1952) as referring to fossils in this formation or the underlying Toroweap Formation in the ZION area.

Considering the Kaibab Formation over its entire area of deposition, the taxonomic list is extensive. In the formation as a whole, there are abundant fossils of marine organisms, including trace fossils, algae (Mather 1970), foraminifera (amoeba-like protists that form "shells"), sponges, corals, conulatans (extinct distant relatives of corals that formed conical structures from rods), bryozoans, brachiopods, bivalves, cephalopods (octopi, squid, and mostly extinct shelled relatives), gastropods, scaphopods (tusk shells), ostracodes (seed shrimp), trilobites, crinoids, echinoids and conodonts (Kirkland 1963; Hopkins 1990; Spamer 1992), and a few reports of chondrichthyan (cartilaginous fish) teeth, spines, and dermal denticles, and ray-finned fish teeth and tooth plates (Hunt et al. 2005). There are distinct mollusk and open marine faunas (Hopkins 1990). Kaibab Formation fossils may also be found as rock fragments incorporated into other formations (McKee 1937). The Fossil Mountain Member has yielded sponges, rugose corals, conulatans, bryozoans, brachiopods, bivalves, cephalopods, gastropods, scaphopods, trilobites, crinoids, echinoids, trace fossils (Sorauf and Billingsley 1991; Spamer 1992), shark teeth (Ossian 1976), and microfossils of mollusks, ostracodes, conodonts, and fish dermal denticles (Thompson 1995). The Harrisburg Member has yielded foraminifera, bryozoans, brachiopods, bivalves, cephalopods, gastropods, scaphopods, trilobites (Nicol 1944; Spamer 1992), crinoids (Jenson 1986), worm tubes (Nicol 1944), conodonts (Clark 1979), and shark teeth (Ossian 1976). The most diverse assemblages are low in this member (Nicol 1944).

Moenkopi Formation (Lower Triassic)

The Moenkopi Formation is a heterogeneous coastal plain and shallow marine unit present in several western states. It has been divided into several members depending on location. At ZION it is divided into the following members, from oldest to youngest: the Rock Canyon Conglomerate Member, Timpoweap Member, lower red member, Virgin Limestone Member, middle red member, Shnabkaib Member, and upper red member (Stewart et al. 1972a; Biek et al. 2009, 2010). There are a few points of historical nomenclatural disagreement. The Rock Canyon Conglomerate Member was long disused as a combination of Permian and Triassic rocks, the latter assigned to the Timpoweap Member (Stewart et al. 1972a). It was briefly reinstated as a formation in its own right (Nielson 1986, 1991), but is now considered to be part of the Moenkopi Formation again. Lucas et al. (2007a) proposed a reorganization of the Moenkopi Formation that would replace the Timpoweap Member's name with that of its eastern equivalent (the Sinbad Member), move the members dominated by carbonate and evaporate rocks to the Thayne Formation (the Sinbad, Virgin Limestone, and Shnabkaib members), and raise both the Thayne Formation and Moenkopi Formation to groups, making all of the members formations; this reorganization has not been widely adopted. The Moenkopi Formation dates to the Early and early Middle Triassic (Heckert et al. 2005a); the members at ZION date to the Early Triassic (Blakey et al. 1993), between about 250 and 245 Ma.

The Moenkopi Formation of ZION consists of a basal conglomeratic member under three pairs of alternating limestone/evaporate units and red-bed units, representing parts of three transgression-regression cycles. At the base is the Rock Canyon Conglomerate, a brownish gray pebble and cobble conglomerate and breccia, up to 15 m (50 ft) thick (Biek 2007; Biek et al. 2009, 2010). It formed from weathering and erosion of the Harrisburg Member of the Kaibab Formation. Its deposits are a combination of valley and channel fill of paleotopography (some channels are up to 150 m [500 ft] deep in the White Hills southwest of St. George [Hayden 2011]), and *in situ* weathering of the Harrisburg Member (Nielson 1991; Biek et al. 2009, 2010). It is overlain by the Timpoweap Member, which is composed of brownish-gray limestone with increasing quantities of yellowish-brown siltstone and sandstone in the upper section, and oil seeps throughout. It is a cliff-forming unit 9 to 24 m (30 to 80 ft) thick (Biek 2007a; Biek et al. 2009, 2010). It was deposited during a marine highstand (Stewart et al. 1972a; Blakey et al. 1993), with an open sea west of the present-day Hurricane Cliffs and an alluvial plain to the east (Blakey 1979). The overlying lower red member is a slope-forming unit composed of reddish-brown mudstone and siltstone, about 50 m (160 ft) thick at ZION (Biek 2007a; Biek et al. 2010). It is a coastal plain and tidal flat unit (Stewart et al. 1972a). The upper contact with the Virgin Limestone Member may be disconformable (McKee 1954; Stewart et al. 1972a), and there may be local disconformities throughout the member (McKee 1954). It represents the regressive part of the transgression-regression cycle began in the Timpoweap Member (Stewart et al. 1972a; Blakey et al. 1993), establishing the carbonate/evaporate to red bed pattern seen in the Moenkopi Formation of southwestern Utah.

The next unit is the Virgin Limestone Member. At ZION it is a locally fossiliferous unit composed of yellowish-gray to gray limestone and interbedded gray mudstone. It forms three limestone ledges in the park, and is about 30 m (100 ft) thick (Biek 2007a; Biek et al. 2009, 2010). Another shallow marine

unit, this member was deposited during a transgression (Stewart et al. 1972a). Large stromatolites (columnar layered sedimentary structures formed by sediment trapped by cyanobacteria) and other microbial structures formed in the absence of grazing animals (Pruss and Bottjer 2004a, 2004b). The upper contact with the middle red member is gradational (McKee 1954). The middle red member is similar to the lower red member, except for the presence of gypsum beds near its base. It is about 60 m (200 ft) thick at ZION (Biek et al. 2010), and is a tidal flat and coastal plain unit deposited during a regression (Stewart et al. 1972a). The upper contact with the Shnabkaib Member is intertonguing and can be difficult to establish (Stewart et al. 1972a). The Shnabkaib Member is the most heterogeneous of the members at ZION, with mudstone, siltstone, gypsum, and thin dolomite beds. It forms slopes and ledges with bands of red mudstone and siltstone and white gypsum, and is about 90 m (300 ft) thick (Biek 2007a; Biek et al. 2010). It includes tidal flat, sabkha, and restricted marine depositional settings, and has been compared to the Trucial Coast of the Persian Gulf and lagoons in Ojo de Liebre, Baja California (Lambert 1984). This member represents another transgression, although not as extensive as previous events. The upper contact with the upper red member is gradational and arbitrary (Stewart et al. 1972a). The upper red member at ZION is similar to the lower and middle red members, except it lacks gypsum and has some sandstone, and forms some ledges as well as slopes. It is about 85 m (275 ft) thick, and like the other red members is a tidal flat and coastal plain unit (Biek 2007a; Biek et al. 2010).

Several of the members are fossiliferous within ZION. In general, the marine units produce fossils such as bivalves, ammonites (typically coiled nautilus-like cephalopods), and gastropods (DeBlieux et al. 2006), and the red beds may produce bits of wood and bone (Gregory and Williams 1947). The Rock Canyon Conglomerate Member includes fossiliferous pebbles probably derived from the underlying Kaibab Formation; fossils include bryozoans, brachiopods, and crinoids (Gregory and Williams 1947 [Timpoweap Member of their usage]). The Timpoweap Member preserves bivalves within the park (Gregory and Williams 1947). The lower red member has fossil bivalves (Mickelson et al. 2006), invertebrate traces (Gregory and Williams 1947; Mickelson et al. 2006) and tetrapod tracks in the park. The tetrapod tracks, from the upper part of the member in the Kolob Canyons District, are mostly examples of *Rhynchosauroides*, with some *Chirotherium* prints and swimming traces as well (Mickelson et al. 2006). The tracks are notable for their age; with the exception of a tracksite reported in the Timpoweap Member (Zeilstra and Lohrengel 2001), they are the oldest Mesozoic vertebrate tracks in North America (Mickelson et al. 2006). Milner and Spears (2007) assigned these tracks to the Virgin Limestone Member. The Virgin Limestone Member has long been recognized as fossiliferous at ZION (Gregory 1947). Gregory and Williams (1947) reported brachiopods and bivalves in this member on Taylor Creek. Santucci and Kirkland (2010) mentioned internal molds of mollusks and fossils of asteroids (sea stars). Biek et al. (2010) reported that the Virgin Limestone contains brachiopod, gastropod, and crinoid fossils in the park. Loose slabs of Virgin Limestone in ZION collections include apparent arthropod traces as well as resting traces of brittle starfish (*Astrosoma*) (Mickelson et al. 2006). The Timpoweap and Virgin Limestone members contain bivalves in Section 10, T38S, R12W, potentially within ZION (Stewart et al. 1972a). A list of taxa found in the Timpoweap, Virgin Limestone, and Shnabkaib members of the ZION area can be found in Gregory (1950a).

Most fossils in the Moenkopi Formation of southwestern Utah come from the Timpoweap or Virgin Limestone members. The fossil assemblage of the Timpoweap Member includes stromatolites (Lucas et al. 2007a), brachiopods, bivalves, ammonites (particularly *Meekoceras*) (Stewart et al. 1972a) gastropods, ostracodes, echinoderms, conodonts, fish fragments (ichthyoliths) (Lucas et al. 2007a), and reptile tracks (from a short distance west of Virgin City; Zeilstra and Lohrengel 2001). Poorly preserved bivalves are locally abundant in this member in the quadrangles including northwestern ZION (Biek 2007a). The Virgin Limestone Member has the most diverse fossils. Its assemblage includes stromatolites, oncolites (similar to stromatolites but spherical) (Sanderson 1967; Pruss and Bottjer 2004a, 2004b), microbial wrinkle structures (Pruss et al. 2004), probable microbial spheroids (Pruss and Payne 2009), algal filaments (Sanderson 1967), bryozoans, brachiopods, bivalves, orthoceratid and ammonoid cephalopods, gastropods, ostracodes, the crustacean *Halicyne*, crinoids (Poborski 1954; Hintze 1986; Jenson 1986; Stewart et al. 1972a), echinoids (Moffat and Bottjer 1999), sea stars (Santucci and Kirkland 2010), and several types of invertebrate trace fossils (Pruss and Bottjer 2004a). In the vicinity of the Kolob Canyons District fossils are locally abundant, including brachiopods, gastropods, and crinoids (Biek 2007a).

The other five members have much smaller fossil assemblages. The Rock Canyon Conglomerate Member so far has only reworked fossils from other units (Gregory and Williams 1947). The lower red member has produced bivalves (Mickelson et al. 2006), possible gastropods (Stewart et al. 1972a), invertebrate traces (Gregory and Williams 1947; Jenson 1986; Mickelson et al. 2006), and tetrapod tracks (Mickelson et al. 2006). The middle red member has produced unspecified fossils (Lundberg and Lohrengel 2005). The Shnabkaib Member has produced stromatolites, algal filaments, brachiopods, bivalves, ammonites, gastropods, crinoids, and invertebrate trace fossils. Microbial structures are common, but fossils are otherwise rare (Lambert 1984). Bivalves, gastropods, and crinoids may be found in carbonate beds (Hintze 1986; Jenson 1986). Brachiopods, bivalves, and gastropods have been found in this member near Virgin City (McKee 1954). The upper red member has produced tetrapod tracks (Milner and Spears 2007; Klein and Lucas 2010), such as those of *Rotodactylus* (Stewart et al. 1972a). In general, red members may yield stromatolites, microbial-coated grains, burrows and bioturbation (general disruption of sediments by organisms) (Reif and Slatt 1979),

and bits of wood and bone (Gregory and Williams 1947). Limestone beds interbedded with the red beds may contain skeletal fragments of brachiopods, bivalves, gastropods, and ostracodes (Reif and Slatt 1979).

Chinle Formation (Upper Triassic)

The Late Triassic-age Chinle Formation is an important fossiliferous unit of the Colorado Plateau. It is a heterogeneous terrestrial unit largely deposited in various fluvial and lacustrine settings, and is divisible into several members depending on location (Stewart et al. 1972b). At ZION, it includes two widely-used members, the Shinarump Conglomerate Member at the base of the formation and the Petrified Forest Member above (Biek et al. 2009, 2010). It may be possible to divide ZION's Petrified Forest Member into additional units (Heckert et al. 2006; Biek et al. 2009). The lower contact with the Moenkopi Formation and the upper contact with the Moenave Formation are disconformities at ZION (Biek et al. 2009, 2010). The Chinle Formation spans much of the Late Triassic. Its base dates to about 225 Ma (Ramezani et al. 2011), and its top predates the Triassic–Jurassic boundary (201.6 Ma) because this division is known to be in the overlying Moenave Formation (Lucas et al. 2011). There are probably at least 10 million years between the Chinle Formation and underlying Moenkopi Formation, a gap known as the Tr-3 unconformity (Pipiringos and O'Sullivan 1978).

Historically, the ZION area has been an important region for the study of the Chinle Formation (Stewart et al. 1972b). Early conceptions of the formation were much greater in stratigraphic extent. Gregory (1950a) proposed a four-member Chinle Formation at the park, including in ascending order the lower sandstone, Petrified Forest, Springdale Sandstone, and upper members. The Shinarump Conglomerate was a separate formation under this proposal. Most of these rocks are now in the Moenave and Kayenta formations of the Glen Canyon Group (Averitt et al. 1955; Stewart et al. 1972b; Biek et al. 2009), with the Chinle Formation restricted to Gregory's Shinarump Conglomerate, lower sandstone member, and lower Petrified Forest Member (Stewart et al. 1972b). This can lead to some confusion when going back through the literature. For example, Hesse's (1935) ZION *Semionotus* specimen was described as coming from the Chinle Formation, but it derived from part of the formation that was later assigned to the Moenave Formation (Stewart et al. 1972b).

The Shinarump Conglomerate Member is one of the most widely exposed units of the Chinle Formation, and is present in Utah, Nevada, Arizona, and New Mexico (Colbert 1972; Ash 1974). It was deposited during the early Late Triassic (Lucas et al. 1997). At ZION, it is a brown cliff-forming unit, composed of medium- to coarse-grained sandstone and pebbly sandstone with common petrified wood fragments. Staining by iron-manganese oxides produces the noted "picture stone" effect. It is 18 to 41 m (60 to 235 ft) thick in the park and is interpreted as the product of north- to northwest-flowing braided streams (Stewart et al. 1972b; Biek et al. 2010). The braided streams were part of a drainage system flowing north into the center of the depositional basin (Parrish 1989). The climate is interpreted as monsoonal (Demko et al. 2005).

The Shinarump Conglomerate Member of ZION and the vicinity has long been known for producing petrified wood (Gregory 1947, 1950a; Gregory and Williams 1947; Stewart et al. 1972b). This characteristic is a stock descriptor of the member in geologic maps of the park (Doelling et al. 1989; Willis and Hylland 2002; Willis et al. 2002; Lockley and Milner 2006; Biek 2007a; Milner and Spears 2007; Biek et al. 2009, 2010). Petrified wood from this member and the Petrified Forest Member is traditionally assigned to *Araucarioxylon* (Gregory 1950a), but recent research has shown that multiple forms are present in the material referred to this name (Savidge 2007; Philippe 2011). Fossils known from the Shinarump Conglomerate Member in the park include petrified wood and other plant fragments, logs assigned to *Araucarioxylon* and *Woodworthia*, a reptile vertebra, and other bones and bone fragments (DeBlieux et al. 2005, 2006; Santucci and Kirkland 2010). Fossils are generally scattered and show signs of being rolled by water (DeBlieux et al. 2006). Fossiliferous pebbles reworked from the Kaibab and Moenkopi formations are also present, with marine fossils such as fusulinid foraminifera, brachiopods, bivalves, and crinoids (Gregory 1950a). Rare Shinarump Conglomerate Member tetrapod tracks are known from outside the park in the St. George area (Lockley and Milner 2006; Milner and Spears 2007). They include tracks of a quadruped, not a tridactyl bipedal dinosaur as first reported, and probably pertain to a phytosaur (a crocodile-like reptile) or aetosaur (an armored herbivorous reptile) (Milner and Spears 2007).

Over its entire depositional area, the most common fossils of the Shinarump Conglomerate Member are plants (Colbert and Gregory 1957). Types of plants found in the Shinarump Conglomerate across its depositional area include horsetails, ferns and fern-like plants, bennettitales, cycads, and conifers and other gymnosperms (Ash 2005a). Some specimens show traces of arthropod attacks (Ash 1996). Other fossils are also known from this unit. Bone fragments are commonly preserved as cobbles, usually too eroded to identify. Some can be identified as belonging to metoposaur amphibians or phytosaurs (Colbert and Gregory 1957; Parrish 1989). Tetrapod footprints, swim traces (Lockley and Milner 2006), and coprolites (fossil feces) are also known (Heckert et al. 2002). For invertebrates, tentaculites (enigmatic possible mollusks, misidentified as shells of the polychaete worm *Spirorbis*), two types of insect eggs, and possible beetle feeding traces have been found (Ash 2005a). Reworked marine fossils from older formations may also be found, such as foraminifera (McKee 1937; Miller and Breed 1962).

The Petrified Forest Member of ZION is a brightly-colored variegated slope-forming unit, 135 to 150 m (450 to 500 ft)

thick (Biek et al. 2009, 2010). It is notably heterogeneous, composed mostly of mudstone and claystone with lesser amounts of siltstone, sandstone, pebbly sandstone, and limestone. Petrified wood is common. It was deposited in a landscape of floodplains, lakes, and high-sinuosity rivers. The upper contact of the member is marked by a thin pebble conglomerate at the base of the overlying Moenave Formation (Biek et al. 2009, 2010). This member is divisible into smaller units at locations such as Petrified Forest National Park (PEFO). Chinle Formation fossil vertebrates are most diverse in the Petrified Forest Member's various units (Parrish 1989).

The Petrified Forest Member of ZION preserves a diverse fossil assembles, including petrified wood and other plant material, invertebrate burrows, coprolites, and fragmentary vertebrate remains from fish, metoposaurs, aetosaurs, phytosaurs, and others (ZION Figure 5) (DeBlieux et al. 2005, 2006; Santucci and Kirkland 2010). Like the Shinarump Conglomerate Member, the Petrified Forest Member of the park has been noted as a producer of petrified wood for decades (Gregory 1947, 1950a; Gregory and Williams 1947). The greatest concentration of fossil logs in the Chinle Formation of ZION is in the lower Petrified Forest Member, just above the Shinarump Conglomerate. Petrified logs are locally abundant enough to be called "petrified forests" (DeBlieux et al. 2005, 2006). Previously, a trail in ZION had been named the Petrified Forest Trail, but it has been renamed to protect the fossils. Petrified wood is frequently collected from areas just outside the park (DeBlieux et al. 2005). Unauthorized collection using large machinery such as backhoes has been an issue in these adjacent areas in recent years (Milner et al. 2009). The Petrified Forest Member has produced the majority of vertebrate body fossils known from ZION. Bones and teeth are typically found in nodules (DeBlieux et al. 2005, 2006). One recent find of interest is a tooth (ZION 15675) of a herbivorous archosauriform similar to the armored *Revueltosaurus*, a genus well-known from PEFO. This tooth was found near Cougar Mountain in southwestern ZION, along with fragments of petrified wood, bone fragments, phytosaur teeth, and small coprolites (Heckert et al. 2006). This is the "ornithischian" (beaked herbivorous dinosaur) tooth of earlier reports (DeBlieux and Kirkland 2003; DeBlieux et al. 2003, 2005); revueltosaur and similar teeth were previously thought to belong to ornithischian dinosaurs. A phytosaur specimen was excavated from land adjacent to the park in the middle 1990s by the College of Eastern Utah (DeBlieux et al. 2006). As with the Shinarump Conglomerate Member, petrified wood is abundant in the Petrified Forest Member of the area, and is frequently noted in passing (Doelling et al. 2002; Willis and Hylland 2002; Willis et al. 2002; Biek 2007a).

There are also a few reports of note for ZION where the member was not specified. Park collections include indeterminate metoposaur fossils from the Chinle Formation identified by the eminent paleontologist E. H. Colbert (Heckert et al. 2006). Other fossils from unspecified beds at ZION include unionid clams (Stewart et al. 1972b) and metoposaur shoulder bones from near Cougar Mountain (Hamilton 1984). Given the similarity of the two reports of metoposaur bones, it would be useful to establish if they refer to the same material.



ZION Figure 5. Phytosaur teeth from the Chinle Formation of ZION (photo courtesy Jim Kirkland, Utah Geological Survey).

The Petrified Forest Member and other members of the Chinle Formation above the Shinarump Conglomerate Member have similar types of fossils, but differ in the genera and species present and which are most commonly found. Plants are represented by charophyte algae (Lucas and Keitzke 1993), diverse palynomorphs (Gottesfeld 1972; Fisher and Dunay 1984; Parker 2004), and leaves, wood, and other remains of lycopods, horsetails, ferns and fern-like plants, cycads, bennettitales, ginkgoes, conifers, and taxa of uncertain classifications (Ash 1972, 1989, 2003, 2005b; Stewart et al. 1972b; Parker and Ash 2004). Fungal remains are also present (Ash 1972; Stewart et al. 1972b). Invertebrates are represented by bivalves, gastropods, ostracodes, conchostracans (clam shrimp) (Breed 1972; Stewart et al. 1972b), crabs (Rinehart et al. 2003), crayfish (Miller and Ash 1988), insects (Breed 1972; Stewart et al. 1972b), and various trace fossils (Caster 1944; Breed 1972; Gillette et al. 2003; Demko et al. 2005), including borings in wood (Walker 1938; Tapanila and Roberts 2012) and a variety of feeding traces on plants (Ash 1996).

The Chinle Formation's vertebrate assemblage has been summarized in a number of documents, including Stewart et al. (1972b), Colbert (1972), Parrish (1989), Long and Murry (1995), Heckert et al. (2005a, 2005b), Irmis (2005a), Parker (2005, 2009), Lucas et al. (2007b), Loughney et al. (2011), and Parker and Martz (2011). Vertebrates include: freshwater sharks; ray-finned fish; lungfish; coelacanths; several types of enigmatic tetrapods known from isolated bones and teeth; metoposaur amphibians; procolophonids (stocky lizard-like herbivorous and insectivorous reptiles); sphenodontians (relatives of the modern tuatara); trilophosaurids (reptiles that resembled large herbivorous lizards); tanystropheids (long-necked piscivorous reptiles); the unusual aquatic reptile *Vancleavea*; drepanosaurids (beaked, possibly aquatic chameleon-like reptiles); several forms of crocodile-like

phytosaurs, known under a variety of names; several forms of herbivorous armored aetosaurs, also known under a variety of names; the small unusual armored *Revueltosaurus*; "rauisuchian" crocodile relatives including poposaurids and rauisuchids, which resembled large carnivorous dinosaurs, and shuvosaurids (toothless forms that resembled ostrich-mimic dinosaurs); "sphenosuchians" (small terrestrial crocodile relatives); lagerpetonids (small bipedal dinosaur relatives); silesaurids (quadrupedal dinosaur relatives); several early dinosaur taxa; and dicynodonts (a type of non-mammalian therapsid, or "mammal-like reptile"). Lungfish, metoposaurs, aetosaurs, and phytosaurs are the most common forms. Vertebrate trace fossils include coprolites (Hunt et al. 1998), swim traces, and footprints possibly attributable to sphenodontians, tanystropheids, phytosaurs, aetosaurs, carnivorous and herbivorous dinosaurs, and mammal relatives (Hunt and Lucas 2007). Bones accumulated in floodplains, bogs, small ponds, and in or near channels (Parrish 1989).

Glen Canyon Group: Moenave Formation (Upper Triassic–Lower Jurassic)

The Moenave Formation of ZION includes, in ascending order, the Dinosaur Canyon and Whitmore Point members (Biek et al. 2009, 2010). The Springdale Sandstone Member, included in the Moenave Formation in Graham (2006), was recently removed from the formation and assigned to the overlying Kayenta Formation (Lucas and Tanner 2006). It is treated separately below. The Moenave Formation was defined as a separate formation after most of the other formations at the park were recognized, so older references, particularly those of Gregory, lack it.

The Dinosaur Canyon Member was initially a separate formation (Colbert and Mook 1951). At ZION, it is composed of reddish-brown sandstone and silty sandstone, with lesser amounts of siltstone and mudstone. It forms ledges and slopes, and is 53 to 64 m (175 to 210 ft) thick (Biek et al. 2009, 2010). The depositional environment is interpreted as fluvial to floodplain (Clemmensen et al. 1989; Blakey 1994; Peterson 1994). It has a complex relationship with the Whitmore Point Member: the Dinosaur Canyon Member both underlies the Whitmore Point Member and laterally intertongues with part of it (Tanner and Lucas 2009a). This has consequences for the age of the Moenave Formation units. The Moenave Formation straddles the Late Triassic–Early Jurassic boundary, which in southwest Utah is thought to be in the upper Dinosaur Canyon Member (Kirkland and Milner 2006) or lower Whitmore Point Member (Lucas et al. 2011), whereas in east-central Arizona, where the Whitmore Point Member does not exist, fossiliferous strata of the upper Dinosaur Canyon Member are thought to be above the boundary (Lucas et al. 2011). The Whitmore Point Member of ZION is predominantly varicolored mudstone, with some sandstone, siltstone, and dolomitic limestone. It is a slope-forming unit 18 to 24 m (60 to 80 ft) thick, interpreted as representing lacustrine and floodplain deposits. The upper contact with the Springdale Sandstone Member is a disconformity (Biek et al. 2009, 2010).

The Moenave Formation is a mosaic of fluvial, lacustrine, and eolian environments deposited under a semiarid to arid climate. The Whitmore Point Member's ephemeral and perennial lake system, centered in southwestern Utah, received drainage from the Dinosaur Canyon Member's fluvial system, flowing to the north and northwest from northern Arizona. To the east there was a dune field that later became the Wingate Sandstone (Tanner and Lucas 2007, 2009a), a formation widely exposed in other NCPN units.

Both members are fossiliferous in ZION, although the Whitmore Point Member has more abundant fossils. The Dinosaur Canyon Member of the park preserves plant fragments, mollusks, invertebrate burrows, and tridactyl dinosaur tracks. There is a new plant site of interest in the Kolob Canyons District. Invertebrate traces and dinosaur tracks are the most abundant fossils in this member in the park (DeBlieux et al. 2005, 2006). The Whitmore Point Member of the park preserves microbial structures, plant fragments and rare petrified wood, invertebrate trace fossils, fish scales and bones, partial fish skeletons (ZION Figure 6), and dinosaur tracks (DeBlieux et al. 2005, 2006), including small bird-like tracks (Smith and Santucci 2001). Invertebrate burrows and theropod (bipedal carnivorous dinosaurs) tracks assigned to *Eubrontes* and *Grallator* are abundant. Greenish-gray dolomitic beds are particularly productive with regard to dinosaur tracks (DeBlieux et al. 2005, 2006).



ZION Figure 6. A partial semionotid from the Moenave Formation within ZION; inset is a reconstruction of Semionotus kanabensis from Milner and Kirkland (2006) (figure courtesy Jim Kirkland, Utah Geological Survey).

Over its entire deposition area, the Dinosaur Canyon Member has a smaller assemblage than the Whitmore Point Member. It has yielded palynomorphs (organic microfossils such as spores and pollen) (Cornet and Waanders 2006), plant fragments, root

traces (Kirkland and Milner 2006), mollusks, invertebrate trace fossils (DeBlieux et al. 2006), the fish *Semionotus*, an undescribed lepidosaur (lizards, snakes, and tuatara), the early crocodylomorph (crocodiles and their close relatives) *Protosuchus*, two pelvises of a small theropod, tracks of crocodylomorphs (Lucas et al. 2005), tracks of theropod and ornithischian dinosaurs (Tanner and Lucas 2009a), and possible tritylodont synapsid (a type of "mammal-like reptile") burrows (Tanner and Lucas 2009b). *Protosuchus* and some of the track forms only appear high in the member (Tanner and Lucas 2009a). The relevant ichnofossil taxonomy can be complicated. A recent summary for the Chinle Formation and Glen Canyon Group can be found in Lucas et al. (2006a).

The Whitmore Point Member microbial and plant assemblage includes stromatolites, microbial laminations (Kirkland and Milner 2006), palynomorphs (Cornet and Waanders 2006), ferns, horsetails, conifers (Tidwell and Ash 2006), and root traces (Milner and Spears 2007). Invertebrates are represented by conchostracans (clam shrimp) (Lucas and Milner 2006), ostracodes (Schudack 2006), and several types of trace fossils (Lucas et al. 2006b). Vertebrates are represented by hybodont sharks, palaeoniscid and semionotid fish, lungfish, coelacanths (Milner and Kirkland 2006), a turtle shell (Olsen and Padian 1986), rare theropod bones, coprolites, fish swim markings, and several types of tetrapod footprints, including tracks of synapsids, sphenodontians (tuatara, represented by *Exocampe*), crocodylomorphs (*Batrachopus*), large and small theropod dinosaurs, and ornithischian dinosaurs (*Anomoepus*) (Hunt and Lucas 2006a; Milner and Spears 2007). Local fish scales and bone fragments can be found in the immediate area of ZION (Doelling et al. 2002; Willis and Hylland 2002; Willis et al. 2002; Biek 2007a).

Of particular note for the Moenave Formation of the ZION region are the numerous new finds in the St. George area. By far the best-known of these is the St. George Dinosaur Discovery Site at Johnson Farm, a prolific locality discovered by accident in 2000. Multiple fossiliferous beds have been found here in the upper Dinosaur Canyon and Whitmore Point members. These fossils are associated with the shorelines of an ancient lake or series of lakes known as Lake Dixie (Kirkland and Milner 2006; Milner and Kirkland 2006; Milner and Spears 2007). The fossil assemblage found here includes stromatolites, ferns, horsetails, conifers (Tidwell and Ash 2006), root traces (Milner and Spears 2007), conchostracans (Lucas and Milner 2006), ostracodes (Schudack 2006), invertebrate traces (Lucas et al. 2006b), the hybodont shark *Lissodus johnsonorum*, a new palaeoniscid fish, new semionotid fish, the lungfish *Ceratodus stewarti*, a new taxon of coelacanth (Milner and Kirkland 2006), a theropod vertebra and two types of theropod teeth, coprolites, fish swim markings, and several types of tetrapod footprints, including tracks of synapsids, *Exocampe*, *Batrachopus*, large and small theropod dinosaurs (walking, swimming, and crouching), and *Anomoepus* (Hunt and Lucas 2006a; Milner and Spears 2007). The main track layer is at the base of the Whitmore Point Member. Most fossils are in the Whitmore Point Member, but the upper Dinosaur Canyon Member has produced some of the plant fossils, invertebrate traces, and reptile tracks (Milner and Spears 2007). Fish scales and bones may be found in long cylindrical concretions formed by wave action (Chin et al. 2003). There are many wading theropod tracks; the predators may have been hunting the various large fish (Milner and Kirkland 2007).

Glen Canyon Group: Kayenta Formation (Lower Jurassic): Springdale Sandstone Member

As noted previously, the Springdale Sandstone Member was originally assigned to the Chinle Formation (Gregory 1950a). It was later moved into the Moenave Formation (Harshbarger et al. 1957), and is now assigned to the Kayenta Formation (Lucas and Tanner 2006). It is partially equivalent to the lower sandy Kayenta Formation to the east (Riggs and Blakey 1983; Lucas and Tanner 2006). The Springdale Sandstone Member of ZION is a cliff-forming yellowish gray to pink unit composed mostly of sandstone, with lesser amounts of mudstone and conglomerate. It is 27 to 46 m (90 to 150 ft) thick (Biek et al. 2009, 2010) and is interpreted as a braided stream and floodplain unit (Lucas and Tanner 2006). It grades upward into the Kayenta Formation (Lucas and Tanner 2006). During its deposition, this area of North America was in a monsoonal belt, moving into arid climates (Riggs and Blakey 1983).

The Springdale Sandstone Member is fossiliferous at ZION. Numerous dinosaur track localities have been found at the top of the member; *Eubrontes* is the most common type of track. Tracks are rare lower in the member (DeBlieux et al. 2006). Among the tracksites in the upper Springdale Sandstone Member at the park is a locality interpreted as showing three theropods traveling together (Smith et al. 2002). Another tracksite has ornithopod (beaked herbivorous bipedal dinosaurs)-like footprints (Smith and Santucci 2001). Aside from the dinosaur tracks, other Springdale Sandstone Member fossils in ZION include root traces (Biek et al. 2004a), petrified wood, poorly preserved plant fragments, and bioturbation from invertebrate burrowing (DeBlieux et al. 2005, 2006). Small pieces of petrified wood have been found in this member in Zion Canyon (Santucci and Kirkland 2010).

The ZION tracksites at the top of the member are part of a megatracksite (Lucas and Tanner 2006; Milner and Spears 2007). This megatracksite is found through southern Utah and northern Arizona at the top of the unit, from St. George, Utah, to Tuba City, Arizona, over 185 km (115 miles) southeast of ZION (Lucas et al. 2005; Hamblin et al. 2006; Lucas and Tanner 2006; Milner and Spears 2007). It is sometimes attributed to the base of the main body of the Kayenta Formation instead (DeBlieux et al. 2005). Dinosaur tracks, particularly *Eubrontes* and *Grallator*, are the most common fossils found in this unit (Lucas and Tanner 2006). Some unusual finds have been made in this unit in southwestern Utah. One trackway shows a theropod moving at about 41 km/hr (25 mph), just shy of the estimated record of

43 km/hr (27 mph) (Hamblin et al. 2006; Milner and Spears 2007). Another site has more than 200 *Eubrontes* prints (Hamblin et al. 2006). Aside from the dinosaur tracks, the fossil assemblage of the Springdale Sandstone Member includes petrified wood, invertebrate traces, fish scales, and bones (Doelling et al. 1989; Lucas and Tanner 2006). Hesse's fish is occasionally listed incorrectly as coming from the Springdale Sandstone Member (Harshbarger et al. 1957; Clark and Fastovsky 1986).

Glen Canyon Group: Kayenta Formation (Lower Jurassic)

The Kayenta Formation of ZION and the immediate vicinity is composed mostly of reddish-brown siltstone beds, with some fine-grained sandstone, mudstone, and rare thin beds of dolomite. A tongue of the Navajo Sandstone (the Lamb Point Tongue) divides it into a main body and overlying Tenney (sometimes misspelled Tenny) Canyon Tongue. The Tenney Canyon Tongue is similar to the main body of the formation, except it becomes sandier going up-section. The main body of the formation is 88 to 110 m (290 to 360 ft) thick, the interposed Lamb Point Tongue is up to 37 m (120 ft) thick (but usually much thinner in ZION [Willis et al. 2002; Biek et al. 2009, 2010]), and the Tenney Canyon Tongue, where distinguished by the presence of the Lamb Point Tongue, is 43 to 96 m (140 to 315 ft) thick. The lower contact of the formation with the Springdale Sandstone Member is conformable, and the upper contact with the Navajo Sandstone is gradational (Biek et al. 2009, 2010). The upper Kayenta Formation intertongues with the lower Navajo Sandstone because the dunes of the latter overrode the streams of the former, advancing from the northeast (Riggs and Blakey 1983). The age of the Kayenta Formation is now thought to be middle Early Jurassic, between approximately 197 and 190 Ma (Lucas et al. 2005). The Kayenta Formation was formed by shifting freshwater braided and meandering streams (Colbert 1974); floodplain deposits are also known (Lucas et al. 2005). At ZION, the major depositional environments are rivers and playas, with minor lake deposition. The Lamb Point Tongue of the Navajo Sandstone represents a short-lived eolian (wind-blown) dune field incursion (Blakey et al. 1988; Blakey 1994; Peterson 1994).

The Kayenta Formation includes two major facies: a sandstone-rich facies near its type locality in northeastern Arizona, and a siltstone-rich facies to the west. These are known respectively as the typical (or sandy) facies and silty facies (Harshbarger et al. 1957). ZION is entirely within the silty facies (Blakey 1989), expressed as interbedded siltstone, mudstone, and sandstone. It is thought that the silty facies represent the sluggish lower part of Kayenta Formation rivers and streams (Harshbarger et al. 1957), where low-energy drainage fed lush floodplains (Jenkins and Shubin 1998). The silty facies is much more fossiliferous than the sandy facies, which is best known for its vertebrate tracks (Hamblin and Foster 2000; Lucas and Tanner 2007). Examples of sandy-facies fossils can be found at Canyonlands National Park (CANY) (Santucci et al. 1998; Santucci and Kirkland 2010), Capitol Reef National Park (CARE) (Santucci et al. 1998), and Glen Canyon National Recreation Area (GLCA) (Lockley et al. 1998; Santucci and Kirkland 2010), extending through the formation proper into the Kayenta Formation–Navajo Sandstone transition zone.

The Kayenta Formation is fossiliferous within ZION. One of the earliest reports of fossils from ZION is of a Kayenta Formation tracksite, discovered on the Left Fork of North Creek by Nelson Maloney about 135 m (450 ft) below the upper contact of the formation. The tridactyl tracks were attributed to a herbivorous dinosaur (Stokes and Bruhn 1960). Tracks continue to be the most prominent Kayenta Formation fossils in ZION, with dozens of tracksites now known. The formation has the highest concentration of fossils of any of the park's formations (DeBlieux et al. 2006). Most of the tracks are of the theropod types *Eubrontes* and *Grallator* (Santucci and Kirkland 2010). One locality has a track similar to that of *Brasilichnium*, attributed to "mammal-like reptiles" (Smith and Santucci 2001), and DeBlieux et al. (2005) included a photo of a four-toed track. Tracks are particularly common at the Springdale Sandstone Member–Kayenta Formation contact (DeBlieux et al. 2005). ZION tracksites are also described in Santucci et al. (1998), Smith and Santucci (1999), and Santucci et al. (2006). Some Kayenta Formation carbonate beds include ganoid fish scales; similar beds just east of ZION have produced fossils of semionotid fish and coelacanths (Santucci and Kirkland 2010), and in the area are known to produce trace fossils from gastropods or worms (Hamilton 1984). DeBlieux et al. (2005, 2006) reported finding scales of *Semionotus kanabensis* in ZION. Kayenta Formation dinosaur tracks are also known in the St. George area (Hamblin et al. 2006; Lockley et al. 2006a; Milner and Spears 2007). Fish fossils, coprolites, and crocodylomorph-like tracks are also known from some of these sites (Milner and Spears 2007). The top of the Kayenta Formation is a megatracksite in much of southern Utah; the Kayenta-Navajo transition zone of the Red Cliffs Recreation Area provides a local example (Milner and Spears 2007).

Most vertebrate body fossils from the Kayenta Formation are from the silty facies of the formation, and were discovered in the Gold Spring Wash area of northeastern Arizona, over 200 km (125 miles) southeast of ZION in the vicinity of Cameron and Tuba City (Lucas et al. 2005). Per Lucas et al. (2005), taxa represented in the formation by body fossils include sharks, bony fish, the frog *Prosalirus*, the caecilian (limbless amphibian) *Eocaecilia*, the turtle *Kayentachelys*, a sphenodontian reptile, numerous crocodylomorphs including *Calsoyasuchus*, *Eopneumatosuchus, Kayentasuchus*, *Protosuchus*, and unnamed taxa, the pterosaur (flying reptile) *Rhamphinion*, theropod dinosaurs *Dilophosaurus*, "Megapnosaurus" *kayentakatae*, and the undescribed "Shake N Bake" theropod, a prosauropod dinosaur (since named *Sarahsaurus* [Rowe et al. 2011]), armored dinosaurs *Scelidosaurus* and *Scutellosaurus*, an unnamed heterodontosaurid dinosaur, the tritylodontid synapsids

Dinnebiton, *Kayentatherium*, and *Oligokyphus*, possible early mammal *Dinnetherium*, and a haramyid mammal (there are discrepancies between the table and text in Lucas et al. [2005]; the text seems more accurate). Also reported are lungfish (Jenkins and Shubin 1998), salamanders (Curtis and Padian 1999), lizards (Jenkins et al. 1983), the theropod *Kayentavenator* (Gay 2010), and morganucodontid mammals (Jenkins et al. 1983). Sandy facies fossils include some of the crocodylomorphs. Silty facies fossils are more diverse, and include petrified wood, amphibians, turtles, lizards, abundant sphenodontians, *Eopneumatosuchus*, pterosaurs, *Dilophosaurus* and other theropods, theropod tracks, prosauropods, *Scutellosaurus*, and haramyid and morganucodontid mammals. Fossils in both facies include coprolites, fish scales, crocodylomorphs, and tritylodontids such as *Kayentatherium* (Clark and Fastovsky 1986). Frogs are the rarest component of the assemblage and are only found in the silty facies (Jenkins and Shubin 1998).

Kayenta Formation vertebrate trace fossils include coprolites (Luttrell and Morales 1993), therapsid tracks, lizard-like tracks (Hamblin and Foster 2000), and dinosaur tracks such as *Grallator* and *Eubrontes* (Lucas and Tanner 2007). Kayenta Formation fossils that are not vertebrates or vertebrate traces include microbial limestone (Luttrell and Morales 1993), petrified wood, which is locally abundant in the silty facies (Jenkins and Shubin 1998), plant impressions (Rigby et al. 1971), freshwater bivalves and gastropods (Harshbarger et al. 1957), ostracodes (Lucas and Tanner 2007), and invertebrate trails and burrows (Luttrell and Morales 1993).

Glen Canyon Group: Navajo Sandstone (Lower Jurassic)

The Navajo Sandstone is a prominent component of ZION's geology and landscape. It is a cliff-forming fine- to medium-grained sandstone, usually with large cross-beds from dunes but sometimes featuring planar bedding instead, from deposition in a sabkha at the Kayenta–Navajo transition and scattered interdunal ponds. It grades out of the underlying Kayenta Formation but has a disconformable upper contact with the Temple Cap Formation. As noted above, a relatively thin tongue of Navajo Sandstone, the Lamb Point Tongue, interrupts the Kayenta Formation (Doelling 2002; Willis et al. 2002; Biek et al. 2009, 2010). This tongue had been misinterpreted as the Wingate Sandstone (Gregory 1950a), but was later reidentified (Averitt et al. 1955). The main body of the Navajo Sandstone is informally divided at ZION into three subunits, based on color. In ascending order they are the brown, pink, and white subunits. The brown subunit is 120 to 180 m (400 to 600 ft) thick, the pink subunit is 180 to 500 m (600 to 1,500 ft) thick, and the white subunit is 0 to 245 (0 to 800 ft) thick. The brown and white subunits are somewhat better consolidated than the pink subunit, which is instead known for its localized iron mineralization (Biek et al. 2010). While dating of Glen Canyon Group units has proven difficult to establish, the bulk of the Navajo Sandstone probably dates to the middle and late Early Jurassic (Lucas et al. 2005), between approximately 190 and 175 Ma. Formerly, the formation was thought to intertongue with Middle Jurassic units, but this was based on misinterpretations of similar eolian sandstones (such as the White Throne Member of the Temple Cap Formation) above it (Peterson and Pipiringos 1979).

The Navajo Sandstone represents the largest erg known from North America (Blakey 2008), and Kocurek (2003) identifies it as the largest known on Earth. As discussed in the "Kayenta Formation" section, the upper Kayenta Formation intertongues with the lower Navajo Sandstone as the dunes of the latter overrode the streams of the former. Recent research on geochronology of detrital zircon grains in the Navajo Sandstone suggests that most of the sand was eroded from the ancestral Appalachian Mountains, transported westward by a continental-scale river system to the western shore of Jurassic North America, and was then blown southward and incorporated into the Navajo–Nugget–Aztec dune field (Dickinson and Gehrels 2003, 2009; Rahl et al. 2003). Preservation of this tremendous thickness of wind-blown sand was made possible because of basin subsidence associated with Early Jurassic compressional deformation near the west margin of North America. This deformation caused the continental interior to flex downward, creating accommodation space for sediment accumulation (Allen et al. 2000). The Navajo Sandstone formed in a tropical arid setting with eolian, playa, and sabkha deposits (Mahaney et al. 2004). Wetter restricted interdunes are also known, and are characterized by mudstone with planar bedding, dolomitic limestone, and sandstone (Irmis 2005b). Interdune deposits are best known from the vicinity of Moab, ranging from fluvial sandstones to lake limestones. During periods of increased precipitation, vegetation spread (Wilkens 2008). Fossils are mostly associated with interdunes (Loope et al. 2004b). As time went on during the deposition of the Navajo Sandstone, the climate became drier (Irmis 2005b), although there were several periods of increased precipitation with durations on the order of tens of thousands of years (Milan et al. 2008). The cyclicity of the Navajo Sandstone at ZION has been described by Chan and Archer (2000). They noted the presence of 30-year and 60 year bundles of seasonal or annual cycles, possibly due to climate oscillations driven by orbital factors (similar to modern Milankovitch cycles) and/or seasonal precipitation (Chan and Archer 2000).

The Navajo Sandstone is not extensively fossiliferous, but does preserve fossils within ZION. Most of the fossils are trace fossils. Dinosaur footprints have been observed along the trail to Observation Point in Zion Canyon, and tracks of several types of animals are visible on a fallen Navajo Sandstone boulder near Parunuweap Canyon (DeBlieux et al. 2005, 2006; Santucci et al. 2006; Santucci and Kirkland 2010). A photograph of the boulder shows two types of tracks from tridactyl bipeds and tracks from a quadruped (DeBlieux et al. 2005). Tracks may actually be more common than is observed, but circumstances are unfavorable for finding them. Tracks are found on bedding

planes, which can only been seen at the tops of cliffs and on fallen blocks. Even where a plane is exposed, there must be some erosion to make the tracks stand out (DeBlieux et al. 2005). There is also a report of poorly preserved fossil wood (DeBlieux et al. 2006; Santucci and Kirkland 2010). It was observed in 1962 in a shale lens (Santucci and Kirkland 2010). Finally, general bioturbation has been observed in Navajo Sandstone rocks in ZION and its immediate surroundings (Freeman and Visher 1975; Doelling 2002; Doelling et al. 2002; Willis and Hylland 2002; Willis et al. 2002).

To date, fossil organisms and traces represented in the Navajo Sandstone over its entire depositional area include stromatolites, root traces (Irmis 2005b), pollen, spores, horsetails, ferns (Loope et al. 2004b), cycadeoids, conifers (Wilkens 2008), petrified wood (Irmis 2005b), possible fungal filaments (Mahaney et al. 2004), worm burrows (Stokes 1978), insect burrows, scorpion tracks (Loope et al. 2004b), unionid bivalves (Wilkens 2008), ostracodes, conchostracans (Harshbarger et al. 1957), a crocodylomorph, the small theropod dinosaur *Segisaurus* (Irmis 2005b), at least two types of prosauropods including *Seitaad* (Sertich and Loewen 2010), a tritylodontid synapsid, and vertebrate trace fossils (Irmis 2005b). The assemblage may be a persistent remnant of the Kayenta Formation assemblage (Winkler et al. 1991). Sandstone dune deposits have mostly yielded vertebrates, while limestone interdune beds have mostly yielded plants and invertebrates (Stokes 1991), although arthropod tracks (*Octopodichnus*) are known from the dunes (Hunt and Lucas 2006b). Interdune fossils include stromatolites, root casts, horsetails, ostracodes, and insect traces, with locally common petrified wood (Irmis 2005b).

Vertebrate trace fossils are much more common than body fossils in the Navajo Sandstone, and have been extensively described. More than 60 dinosaur track sites are known in the Navajo Sandstone across its depositional area (Loope 2006; Milan et al. 2008). The various vertebrate trace fossils have gone under different names and interpretations, depending on the author. Recent works (Irmis 2005b; Hunt and Lucas 2006b; Loope 2006) seem to have settled on seven or eight primary ichnogenera: *Brasilichnium* (synapsid, but see Milan et al. 2008 for an interpretation as a sauropodomorph dinosaur), *Batrachopus* (crocodylomorph), *Grallator* (small theropod), *Anchisauripus* (medium theropod), *Eubrontes* (large theropod), *Navahopus* (small ?sauropodomorph), *Otozoum* (larger sauropodomorph), and *Anomoepus* (ornithischian dinosaur) (Irmis 2005b). A few oddities are also known, including lizard-like tracks (*Lacertipus*), a track once thought to be pterosaurian in origin ("*Pteraichnus*"), bird-like *Trisauropodiscus* tracks (Irmis 2005b), undescribed large ornithopod tracks (Santucci and Kirkland 2010), and large vertebrate burrows thought to have been made by mammals and/or mammal relatives (Lucas et al. 2006c; Riese et al. 2011). So far, large traces (*Anchisauripus*, *Eubrontes*, and *Otozoum*) are only known from interdunes, with smaller traces (*Batrachopus*, *Brasilichnium*, and *Grallator*) the only tracks found in dune sands (Loope 2006), although this differentiation may be due to preservational bias (Irmis 2005b). In general, the vertebrate trace fossils are consistent with the known body fossils.

San Rafael Group: Carmel Formation (Middle Jurassic)

The Carmel Formation is a heterogeneous unit, reflecting its depositional origin in and around a shallow sea (Blakey 1989). It and its members have complicated nomenclatural histories at ZION. The modern definition of the formation at the park includes from oldest to youngest the Co-op Creek Limestone, Crystal Creek, Paria River, and Winsor members (Biek et al. 2009, 2010). Earlier, they had been assigned to four different formations, from oldest to youngest the Carmel Formation, Entrada Sandstone, Curtis Formation, and Winsor Sandstone (Gregory 1950a). Thompson and Stokes (1970) redescribed the units and introduced the modern names, with the exception of the Co-op Creek Limestone Member: they named it the Kolob Limestone Member, but another rock unit from the area was already named "Kolob," so it was changed to the Co-op Creek Limestone (Doelling et al. 1989). For a period of time, the four members were also known as the limestone, banded, gypsiferous, and Winsor members (Peterson and Pipiringos 1979). The name of the Co-op Creek Limestone's equivalent to the east, the Judd Hollow Member, has also sometimes been applied in the ZION area (Blakey et al. 1983; Blakey 1989). To further complicate matters, Nielson (1990) assigned informal members for the formation in nearby southwestern Utah. Members A through D pertain to the Co-op Creek Limestone Member, the E Member is thought to be the Crystal Creek Member, and the F Member is thought to be the lower Paria River Member (Nielson 1990).

The Carmel Formation of ZION is exposed through much of the park, but most of the exposures are of the Co-op Creek Limestone. The Co-op Creek Limestone Member is a gray limestone and shale unit divided into a slope-forming lower unit and cliff-forming upper unit. The lower unit is 45 to 53 m (150 to 170 ft) thick, and the upper unit is 30 to 33 m (100 to 110 ft) thick (Biek et al. 2009, 2010). This member is shallow marine in origin (Imlay 1980). Some of the basal part of the member has been transferred to the underlying Temple Cap Formation as the Esplin Point Member (Sprinkel et al. 2011); this was anticipated by Hylland (2010), who reported that a red shaly interval at the base of the Co-op Creek Member might be part of the upper Temple Cap Formation. The Crystal Creek Member is only found in northern ZION within the park. It is composed of reddish brown siltstones and sandstones deposited in coastal sabkha and tidal flat settings. It is 45 to 55 m (150 to 185 ft) thick. The Paria River Member is also only found in the northern part of the park. It is a ledge- and cliff-forming unit. The lower three-quarters is primarily gypsum with sparse interbedded mudstone and sandstone. The upper part is a fossiliferous limestone. This member was deposited in restricted shallow marine settings, becoming deeper over time, and is 15

to 24 m (50 to 80 ft) thick. Finally, the Winsor Member is present only on Horse Ranch Mountain at ZION. It is composed mostly of yellowish-gray to reddish-brown sandstone, with some siltstone and pebbly sandstone, and represents deposition on a sandy mudflat. It is 55 to 85 m (180 to 280 ft) thick (Biek 2007a; Biek et al. 2009, 2010).

A pair of large-scale transgression-regression cycles are responsible for the variable geology of the four members. Marine transgressions from the Carmel–Twin Creek (or Sundance) Seaway to the north in central Utah produced the Co-op Creek Limestone and Paria River members, and regressions produced the Crystal Creek and Winsor members (Blakey et al. 1983). Another erg (the Page Sandstone erg) was present to the east (Riggs and Blakey 1983). The setting has been compared to the Trucial Coast of the Persian Gulf (Wilson and Palmer 1994). Not surprisingly, the two shallow marine units are the most fossiliferous. Volcanic activity was in progress during deposition, from a volcanic arc to the west and southwest (Kowallis et al. 2001). This corresponds to volcanic activity on the southern margin of the Colorado Plateau that occurred around 170 Ma (Blakey and Parnell 1995). The volcanism and fossil content help to date the formation. The Co-op Creek Limestone Member is regarded as dating from about 169 to 167 Ma, the Crystal Creek and Paria River members date from about 166 to 165 Ma, and the Winsor Member dates from about 164 to 162 Ma (Sprinkel et al. 2011). The upper contact of the formation is a disconformity at ZION (Biek et al. 2010).

The Carmel Formation of ZION has long been known for its fossils. Gregory (1947) noted the presence of abundant fossils in the formation (when it was equivalent to the Co-op Creek Limestone only). Marine invertebrates, typically bivalves, gastropods, and star-shaped columnals of the crinoid *Isocrinus*, are found throughout the park in the Co-op Creek Limestone Member (Biek et al. 2010). Biek et al. (2004a) describes a few locations within ZION where the member is fossiliferous, and Gregory (1950a) includes a taxonomic list for ZION and the area. The Paria River Member is also known to be fossiliferous in the ZION area, preserving bivalves, gastropods, and crinoids (Gregory 1950a; described as the Curtis Formation). Microbial structures, bivalves, and gastropods are present in an unspecified member along Wildcat Trail (Hamilton 1984). Fossiliferous Carmel Formation material is also found as float in the park (Biek et al. 2004a). For the area including ZION and its surroundings, the presence of marine fossils is frequently noted in passing for the Co-op Creek Limestone Member (Sable and Doelling 1993; Doelling 2002; Doelling et al. 2002; Willis and Hylland 2002; Willis et al. 2002; Biek 2007a, 2007b; Hylland 2010) and the Paria River Member (Sable and Doelling 1993; Biek 2007a, 2007b).

Over its entire depositional area including outside of ZION, the Co-op Creek Limestone Member has a diverse marine assemblage, including stromatolites, foraminifera, radiolarians (similar to foraminifera, but secrete silica skeletons), corals, bryozoans, bivalves, ammonites, gastropods, ostracodes, sea stars, crinoids, echinoids, and worm tubes and other invertebrate trace fossils (Imlay 1964; Nielson 1990; Wilson 1998). The Co-op Creek Limestone Member's fossils are sometimes found in association with what had been cavities (Wilson 1998; Baker and Wilson 1999) or as ostreoliths (oyster balls) (Wilson et al. 1998). Crinoid fossils can be abundant enough to form the majority of limestones, which are called encrinites. They are among the most recent examples in the world. The crinoid that produces the common star-shaped columnal elements is *Isocrinus*, previously known as *Pentacrinus* (Tang et al. 2000). The Crystal Creek Member is sparsely fossiliferous, with only bioturbation reported (Blakey et al. 1983). The Paria River Member has produced bivalves, gastropods, crinoids (Gregory 1950a), and trace fossils including tracks and fecal pellets (Blakey et al. 1983). No specific reports of fossils have been attributed to the Winsor Member. Sprinkel et al. (2011) reported the presence of microfossils (dinoflagellate plankton, spores, and pollen) and megafossils in all four members, allowing them to assign the members to various stages of the Middle Jurassic. Terrestrial fossils from the Carmel Formation include spores, pollen (Sprinkel et al. 2011), and tracks of crocodylomorphs and dinosaurs (Lockley and Hunt 1995).

Cedar Mountain Formation/Dakota Formation (Lower–Upper Cretaceous)

The youngest Mesozoic interval at ZION is an area of rocks dating roughly to the middle of the Cretaceous, at the top of Horse Ranch Mountain in the Kolob Canyons District. Generally, this interval has been solely attributed to the Dakota Formation, but recent mapping indicates that it mostly composed of the Cedar Mountain Formation, with some Dakota Formation at the top (DeBlieux et al. 2005; Biek 2007a; Biek et al. 2009, 2010). The Horse Ranch Mountain section of Cretaceous rocks is poorly exposed (Biek 2007a) and difficult to observe because of forest cover (DeBlieux et al. 2005). It is estimated to be 6 to 8 m (20 to 25 ft) thick, and is largely composed of yellowish-brown conglomerate and pebbly sandstone. There are also angular blocks of Dakota Formation-like sandstone in Horse Ranch Mountain colluvium, which may be evidence of a thin concealed *in situ* layer of Dakota Formation above the conglomerate and below the younger cover (Biek 2007a). The age, lithology, and stratigraphic position of these beds suggest correlation to the Cedar Mountain Formation of central Utah, but this correlation is not universally accepted (Titus et al. 2005). Both the Cedar Mountain Formation and Dakota Formation of ZION are interpreted as having a fluvial origin (Biek 2007a; Biek et al. 2010).

The Cedar Mountain Formation is often divisible into multiple members. At ZION and in nearby quadrangles, the unit is designated as an informal conglomerate unit and overlying mudstone unit. A date of 97.9 ± 0.5 Ma has been established for the mudstone unit, which is corroborated by late Early Cretaceous pollen (Biek 2007a, 2007b; Biek and Hylland 2007; Hylland 2010). This mudstone bed may correlate to the

Mussentuchit Member of central and eastern Utah, with the underlying conglomerate representing a major river system (Biek et al. 2009). The upper contact with the Dakota Formation is a disconformity in the area (Biek 2007a, 2007b).

The Cedar Mountain Formation is extensively fossiliferous in central and eastern Utah, and its various members have distinct fossil assemblages. Common non-vertebrate fossils include palynomorphs (Tschudy et al. 1984; Garrison et al. 2007), petrified wood (Doelling 2010), tree ferns (Tschudy et al. 1984), cycads, cycadeoids, conifers (Kirkland et al. 1999), angiosperms (Tidwell and Thayn 1985), root traces (McDonald et al. 2010), bivalves, gastropods, and invertebrate trace fossils (Garrison et al. 2007). Common vertebrates include sharks, ray-finned fish, lungfish, turtles, sphenodontians, crocodylomorphs, theropods, sauropods, armored dinosaurs, ornithopods, mammals (Kirkland et al. 1997, 1998, 1999; Cifelli et al. 1999; Kirkland and Madsen 2007), and vertebrate tracks, particularly of dinosaurs (Lockley et al. 1999, 2004).

The Dakota Formation (also sometimes called the Dakota Sandstone) is a heterogeneous rock unit with exposures as far from Utah as Minnesota. Because it was named in Nebraska from rocks on the opposite shore of the Western Interior Seaway, its use in Utah may be inappropriate (Witzke and Ludvigson 1994; Titus et al. 2005). The Dakota Formation of the Colorado Plateau is often divided into three informal members, with a conglomeratic base, a heterogeneous coaly middle, and an upper marine sandstone. This division reflects environments changing from channels at the base, to floodplains and coastal plains in the middle, to barrier islands and coastal deposits at the top as the Western Interior Seaway approached from the east and flooded the surface (Kirkland et al. 1997). The age of the formation differs by location, because different areas were affected by the encroaching sea at different times, but overall it was deposited within a few million years of the Early–Late Cretaceous boundary of approximately 100 Ma. In Grand Staircase–Escalante National Monument to the east, the middle of the formation was deposited around 96 Ma (Titus et al. 2005).

The best-known terrestrial fossils from the Dakota Formation are plant fossils (particularly angiosperm leaves) and vertebrate tracks. As early as the middle 1800s, fossil wood and dicotyledonous angiosperm leaves were reported from the Dakota Formation (Meek and Hayden 1862). Ferns, clubmosses, the horsetail *Equisetum* (Skog and Dilcher 1994), cycads, conifers (Baxter 1954), and angiosperms (Lesquereux 1892) have been recognized. The floral assemblage in Utah includes coal (Molenaar and Cobban 1991), palynomorphs (including moss spores, fern spores, and gymnosperm and angiosperm pollen; Nichols 1997), the tree fern *Tempskya* (Ash and Read 1976), *Equisetum*, possible gymnosperms, and angiosperms (Tidwell et al. 1976). More recently, abundant dinosaur tracks in the Dakota Group of eastern Colorado, northeastern New Mexico, and the panhandle of Oklahoma have received a great deal of attention. More than 60 sites have been found, with more than 1,000 trackways; collectively, the sites are known as the "Dinosaur Freeway" (Kukihara and Lockley 2012). The tracks come from Dakota Group rocks that are several million years older than the rocks at ZION, and include footprints of crocodilians, pterosaurs, theropod, ornithopod, and armored dinosaurs, and birds (Lockley et al. 2000, 2006b; Kukihara and Lockley 2012).

Vertebrate body fossils are historically rare in the Dakota Formation, although microvertebrate remains are a new area of interest. Recent publications on screenwashed microfossils from nonmarine Dakota Formation rocks in Utah reported eggshells, gastropods, ostracodes, fish, lizards, crocodilians, dinosaurs, and mammals; locations included areas near BRCA (see the BRCA summary in this report for more details) (Eaton et al. 1999; Eaton 2004). Terrestrial vertebrates are typically found in the middle part of the formation. Other nonmarine fossils include algal cysts (Kovach and Dilcher 1988), thecamoebians (a type of "shell"-secreting protist; van Hengstum et al. 2007), unspecified shelled freshwater invertebrates (Stanton 1922), trace fossils from bees and wasps (Elliott and Nations 1998), freshwater sharks and rays, lungfish, frogs, and turtles (Eaton 1993). Dakota Formation marine fossils include algae, foraminifera (Tibert et al. 2003), dinoflagellate cysts (Nichols 1997), bivalves, gastropods (Tibert et al. 2003), cephalopods, arthropods, echinoderms (Cobban and Hook 1989), and trace fossils, such as marks left by the shells of dead ammonites striking the seafloor (Landman and Cobban 2007). The lower conglomeratic parts of the Dakota Formation may also include reworked fossils from older rock units, such as Paleozoic fusulinid foraminifera, corals, bryozoans, brachiopods, and echinoderms (Peterson 1969a), and reworked petrified wood (Titus et al. 2005).

The Cedar Mountain–Dakota interval of ZION is fossiliferous. Plant impressions and freshwater bivalves are present in the sandstone above the conglomerate on Horse Ranch Mountain (Hamilton 1978; DeBlieux et al. 2005, 2006; Santucci and Kirkland 2010). Similar fossils are present nearby. In the Kolob Reservoir quadrangle outside of ZION, petrified wood is present in conglomeratic beds of the Cedar Mountain Formation, and bivalves and gastropods are present in the Dakota Formation (Biek 2007b). The Dakota Formation of the Clear Creek Mountain quadrangle, which includes a small area of northeastern ZION, has yielded coal, palynomorphs, bivalves, gastropods, and invertebrate trace fossils (Hylland 2010). In the Temple of Sinawava quadrangle, petrified wood up to log size is present in the conglomerate of the Cedar Mountain Formation, and coal is present in the Dakota Formation (Doelling 2002). Dakota Formation coal from the area is also mentioned by Given et al. (1985) and Kirschbaum and Gustason (1989).

Fossiliferous eroded rock fragments (Upper Cretaceous–Paleogene for original deposition)

Although there are no *in situ* remnants of rocks deposited between the Dakota Formation and the past few million years at

ZION, eroded fragments of such formations can be found in the various drainages of the park. Some of them are fossiliferous. DeBlieux et al. (2005) reported finding a Straight Cliffs Formation (Upper Cretaceous) cobble bearing fossils of the bivalve *Crassostrea* in the Virgin River, and clasts with stromatolites from the Eocene-age Claron Formation. The Late Cretaceous-age Tropic Shale, another formation in this series of nearby units, yielded a partial plesiosaur (marine reptile) skeleton on Muddy Creek about 18 km (11 miles) east of ZION (Gillette et al. 1999). All three of these formations are described in more detail in the BRCA summary of this document; the BRCA and CEBR summaries together also include many of the region's Upper Cretaceous and Cenozoic formations.

The Tropic Shale is a shallow marine formation, reflected in its fossils. Coal, wood (Gregory 1951), foraminifera (Tibert et al. 2003; McCormick 2006), corals, gastropods, bivalves, ammonites (Gregory 1951), ostracodes (Hazel 1969), invertebrate trace fossils (Peterson 1969a), sharks (Santucci and Kirkland 2010), bony fish, turtles, plesiosaurs (Albright et al. 2007), mosasaurs (marine reptiles descended from monitor lizards) (Peterson 1969a), and a therizinosaurid dinosaur (Zanno et al. 2009) are known to date.

The Straight Cliffs Formation was deposited near sea level, and thus has a variety of both terrestrial and shallow marine deposits (Peterson 1969b). It has a diverse fossil assemblage. Typical fossils include coal (Peterson 1969b), petrified wood (Peterson 1969b; Sable and Hereford 2004), foraminifera, bivalves, ammonites, gastropods, ostracodes, invertebrate trace fossils (Peterson 1969a; Eaton et al. 2001), and vertebrate microfossils belonging to sharks, rays, bony fish, amphibians, turtles, lizards, crocodiles and crocodylomorphs, various dinosaurs, and early mammals (Eaton et al. 1999).

The Claron Formation records fluvial deposition transitioning to lacustrine deposition (Goldstrand 1990; Davis and Pollock 2010). It is sparsely fossiliferous, and its fossil assemblage is mostly limited to microbial structures such as stromatolites (DeBlieux et al. 2005), sparse bivalves and gastropods (Gregory 1950b), and arthropod traces (Bown et al. 1997).

Quaternary rocks and sediments (Pleistocene–Holocene)

The Quaternary rocks and sediments of ZION are a combination of basaltic lava flows and unconsolidated sediments (alluvium, colluvium, eolian deposits, lake deposits, landslides, and talus) with grain sizes from clay to gravel and larger. Deposits are found throughout the park, and cover large areas around drainages, in the vicinity of the extinct cinder cones, in the southwestern part of the park, and near the western border of the Kolob Canyons District (Hamilton 1978; Graham 2006; Biek et al. 2010). Typical Quaternary fossils of the Colorado Plateau include isolated bones of large mammals (such as sloths, mastodons, mammoths, horses, camels, and bison) and fossil material useful for paleoecological and paleoclimatological studies (such as pollen and packrat middens). Summaries of Quaternary fossil sites in southern Utah and the vicinity include Agenbroad and Mead (1989), Mead and Agenbroad (1992), Jefferson et al. (1994), Gillette and Miller (1999), Anderson et al. (2000), and Coats et al. (2008). Reworked older fossils may also be found; for example, deposits of the ancestral La Verkin Creek have fossiliferous limestone cobbles from the Carmel Formation (Biek et al. 2004a).

The Quaternary deposits of ZION are fossiliferous and mostly fall into the paleoecological category. To date, all of the Quaternary fossils of the park appear to be associated with lake deposits, with the possible exception of leaf impressions found in tufa (spring deposits) from the Zion Narrows; these specimens are probably mostly recent (DeBlieux et al. 2005). As noted, there is evidence for at least fourteen Quaternary lakes in ZION (Hamilton 1978; Biek et al. 2010). Fossils have been described or reported from several of them (Hamilton 1979, 1984; Hevly 1979; DeBlieux et al. 2005; Biek 2007a; Biek et al. 2010). Abundant microfossils including pollen, spores, and a few diatoms (algae that secrete a silica "shell") have been recovered (Hevly 1979), and many of the lake deposits have yielded mollusks (Hamilton 1979). In addition, carbonized roots and stems are present in some lacustrine beds, particularly in those of the Holocene lakes (Hevly 1979).

Coalpits (or Coal Pits) Lake existed in Coalpits Wash during the Pleistocene, after the Crater Hill flow blocked drainage around 125,000 years ago. It was perhaps 33 m (110 ft) deep at most, and probably was periodically dry (Biek et al. 2010). Its deposits have yielded the most diverse fossil assemblage of the Quaternary lakes, including impressions of the mosquito fern *Azolla* and a pondweed, shells of bivalves and gastropods, mollusk burrows, bivalve impressions, traces left by ants, beetles, horsehair worms, and aquatic insect larvae, and tracks of a large bird and an artiodactyl (even-toed ungulate) (ZION Figure 7) (Hamilton 1979; Hevly 1979). The bird track may be from a stork, and the mammal track may have come from a camel or elk (DeBlieux et al. 2005). The lake existed when the area's climate was much cooler and wetter (Hamilton 1979). Trail Canyon Lake was another Pleistocene lake. It appears to have formed twice, once in response to the Grapevine Wash basalt flow sometime between 310,000 and 220,000 years ago, and later after a large landslide (Willis et al. 2002; Biek et al. 2010). Its fossils include pollen, gastropods, a fish vertebra, and a bison vertebra (Hamilton 1979, 1984; DeBlieux et al. 2005). Hevly (1979) also mentioned bird and pocket gopher (*Thomomys*) bones.



ZION Figure 7. Artiodactyl track associated with Coalpits Lake deposits (VLS).

Sentinel Lake in east-central ZION was one of the largest Holocene lakes, attaining depths of at least 60 m (200 ft). Organic material indicates it formed between 7,150 ± 810 and 6,800 ± 580 radiocarbon years before present (9,910–6,280 and 8,795–6,405 calendar years before present per Calib 6.0 [http://intcal.qub.ac.uk/calib/calib.html]; note that radiocarbon dates are not the same as calendar dates, and that radiocarbon years require calibration to yield calendar years) and persisted until about 3,600 ± 400 radiocarbon years before present (4,970–2,920 calendar years before present) (Doelling 2002). Its fossils include bivalves and gastropods (Hamilton 1979).

Several lakes were present in the Kolob Canyons District over the Holocene. Paria Lake has yielded charcoal, pollen, basal *in situ* stumps, and gastropods (Hamilton 1979, 1984; Biek 2007a). Dates on the organics indicated it lasted from 3,900 ± 60 radiocarbon years before present (4,445–4,150 calendar years before present) to 2,800 ± 200 radiocarbon years before present (3,390–2,435 calendar years before present) (Biek 2007a). Hamilton (1979) also reported a bighorn sheep's radius-ulna near a stump, but noted it could be younger. Hop Valley "Lake" was the largest of the district's lakes, though detailed stratigraphic and sedimentological observations did not yield any evidence for a body of standing water; rather, the exposed valley fill is interpreted to be fluvial and eolian in nature (Foos 2002). Peat deposits from 2,640 ± 60 radiocarbon years before present (2,875–2,695 calendar years before present) may provide a basal date, and wood from 670 ± 200 radiocarbon years before present (1,000–290 calendar years before present) approximates its end (Biek et al. 2004a). Its sediments are also bioturbated (Foos 2002). Small Currant (or Current) Creek Lake and nearby Cane Creek Lake also have peat lenses and existed at about the same time as Hop Valley Lake (Biek 2007a). A stump (Biek 2007a) and another radius-ulna are also reported from Currant Creek Lake (DeBlieux et al. 2005).

Comparable pond and bog fossils are known from CEBR and its surroundings to the north-northeast. Within CEBR is Alpine Pond, about 33 km (21 miles) north-northeast of ZION. Its fossil record goes back approximately 3,000 years and includes pollen, spores, conifer needles, and spruce wood (Anderson et al. 1999). CEBR also has a nearly continuous record of Holocene mollusks (Sharpe 1993), and Pleistocene and Holocene bog deposits with peat (Hatfield et al. 2010). Similar fossils are known from the more thoroughly-described Lowder Creek Bog, about 36 km (22 miles) north-northeast of ZION (Mulvey et al. 1984; Anderson et al. 1999; Madsen et al. 2002a, 2002b). Lowder Creek Bog is associated with a small former alpine glacier (Mulvey et al. 1984; Anderson et al. 1999). Its fossil record goes back approximately 17,000 years, as the site went through the glacial period and then transitioned from a lake to a bog (Madsen et al. 2002a). Fossils from Lowder Creek Bog include peat, plant debris, wood fragments (Madsen et al. 2002a), pollen, spores, conifer needles and other plant macrofossils, zooplankton, and hard parts of aquatic arthropods (chironomid midges, beetles, and mites) (Anderson et al. 1999). Another well-studied paleoecological site is the drained Red Valley bog, about 35 km (22 miles) northeast of ZION. It has a discontinuous fossil record centered on the latest Pleistocene, approximately 11,795 ± 90 radiocarbon years before present (13,840-13,420 calendar years before present), and the middle Holocene. Fossil material from this site ranges from peat and charcoal, to pollen and spores, to plant macrofossils such as seeds, cone scales, needles, and branches, to remains of 33 insect and arachnid taxa. Most of the arthropods are beetles, but ants, shield bugs, midges, caddisflies, moths, and mites are also present (Madsen et al. 2002b).

Madsen et al. (2002b) surveyed ZION and adjacent land to the north for packrat middens. Although there is extensive cliff space, they observed few middens. They postulated that the cliff rocks in the park are unsuitable for preserving middens, perhaps because the cliffs are seasonally too wet, or erode too quickly. Middens that are potentially ancient are preserved along the eastern end of ZION, and are known to be present in Parunuweap Canyon just south of the park (Madsen et al. 2002b). Otherwise, the nearest well-documented midden site is near Shinumo Creek in Grand Canyon National Park (GRCA), over 110 km (68 miles) to the southeast. A large midden at this location has yielded plant fossils (Van Devender and Mead 1976; Betancourt et al. 1984) and lizard and rodent bones (Van Devender et al. 1977) from the latest Pleistocene (Mead et al. 1978; Betancourt et al. 1984).

The fossil record of Quaternary vertebrates at ZION is currently limited to the few bones and trace fossils associated with the lake sediments. Fossils of large Quaternary mammals on the Colorado Plateau tend to be found in the vicinity of rivers, as shown by the pattern of localities in Agenbroad (1997). The most commonly-found large mammals are sloths, mammoths, horses, tapirs, camels, bison, oxen, and mountain goats (Agenbroad 1997). The nearest find of an extinct Pleistocene mammal consists of mammoth bones found near Newcastle, about 36 km (22 miles) northwest of ZION (Larson 1999).

Potential Sources of Paleontological Resources

Fossils have not yet been documented from the Temple Cap Formation within ZION. However, it is known to preserve fossils elsewhere, and future field investigations within the park may recover fossils from it.

Temple Cap Formation (Middle Jurassic)

The Temple Cap Formation (formerly the Temple Cap Sandstone) was originally named as a member of the Navajo Sandstone based on exposures in ZION (Gregory 1950a). Peterson and Pipiringos (1979) raised it to a formation and divided it into two members, again based on ZION exposures: the Sinawava and overlying White Throne members. At first, the formation was interpreted as a pair of wedges, with the Sinawava Member pinching out to the east and the White Throne Member pinching out to the west (Peterson and Pipiringos 1979; Blakey 1989). More recent work has found a more complicated relationship in the ZION area. The Sinawava Member is at the base of the formation, overlain by the White Throne Member, which intertongues with marine and marginal marine sandstones from the west related to a unit called the Manganese Wash Member. Above the White Throne Member is the Esplin Point Member, composed of beds formerly included in the basal Co-op Creek Member of the overlying Carmel Formation (Sprinkel et al. 2011). The Sinawava Member of ZION is a reddish-brown unit composed of fine-grained sandstone, siltstone, and mudstone. It is 1 to 18 m (3 to 60 ft) thick in the park (Biek et al. 2010; Hylland 2010), and is interpreted as representing coastal sabkha and tidal flat deposition (Blakey 1994; Peterson 1994). The White Throne Member of ZION is mostly composed of yellowish-brown fine- to medium-grained sandstone similar to that of the Navajo Sandstone, but with a thin red-bed section at its top (Biek et al. 2009, 2010). It is 0 to 58 m (0 to 190 ft) thick (Biek et al. 2010), and is interpreted as representing coastal dune deposits (Blakey 1994; Peterson 1994). The White Throne Member pinches out near the western border of ZION (Blakey 1989; Biek et al. 2009), and is absent from the Kolob Canyons District. The Sinawava Member is unusually thin there as well, suggesting that there was a paleotopographic high there during deposition (Biek 2007a; Biek et al. 2010). The Esplin Point Member is similar to the Sinawava Member, with some thin limestone and gypsum beds. It is about 9 m (30 ft) thick and is interpreted as a shallow marine to marginal marine unit (Sprinkel et al. 2011). Overall, the Temple Cap Formation of the ZION area can be seen as the result of fluctuating eolian (White Throne Member) and marginal marine deposition (the other members), with eolian deposition proceeding from the east and marginal marine deposition proceeding from the west (Riggs and Blakey 1983; Blakey 1989; Sprinkel et al. 2011).

The same pulse of volcanic activity that helps to date the Carmel Formation is also found in the Temple Cap Formation (Kowallis et al. 2001), and a restricted shallow sea was present to the north (Riggs and Blakey 1983). The Temple Cap Formation probably dates to between 173 and 170 Ma; dates as old as nearly 179 Ma in the Pine Valley Mountains may not be accurate (Sprinkel et al. 2011). Biek et al. (2010) reported that the upper contact with the Carmel Formation is disconformable, but this was before the introduction of the Esplin Point Member. The contact at the base of this member was thought to be the J-2 regional unconformity, but it appears that the J-2 hiatus is generally very short to nonexistent in Utah south of the Uinta Mountains (Sprinkel et al. 2011). The Temple Cap Formation is not known for its fossils, and is not known to be fossiliferous at ZION (DeBlieux et al. 2005). Dinoflagellate cysts are known from the Sinawava and Esplin Point members (Sprinkel et al. 2011).

Park Collections

ZION collections include fossils, with noteworthy specimens such as Virgin Limestone slabs with possible arthropod traces and resting traces of brittle starfish (*Astrosoma*) (Mickelson et al. 2006), indeterminate metoposaur fossils from the Chinle Formation identified by the eminent paleontologist E. H. Colbert, the revueltosaur-like tooth from the Petrified Forest Member (Heckert et al. 2006), bones and scutes of metoposaurs and phytosaurs collected by Helmut Ehrenspeck from Petrified Forest Member mudstones near Cougar Mountain (DeBlieux et al. 2006), and a partial specimen (ZION 667) of the fish *Semionotus kanabensis* (DeBlieux et al. 2006). In addition, fossil collections from BRCA have been moved to ZION (S. Eagan, BRCA chief of resources management, pers. comm., October 2010).

The specimen of *Semionotus* cf. *gigas* described by Hesse is in the collections of the University of California Museum of Paleontology (Berkeley, CA) as UCMP 29530 (Hesse 1935).

Cultural Resource Contexts

ZION has an unusual case of a possible fossil resource-cultural resource association. At the South Gate petroglyph site just inside of ZION's south entrance, there are six Ancestral Puebloan (Virgin Anasazi) petroglyphs that appear to replicate tridactyl dinosaur tracks. Near the site is a fallen slab of Kayenta Formation rock with five similar-looking dinosaur footprints, suggesting that the petroglyph maker was inspired by the tracks (Staker 2006). A similar site can be found at Flag Point, Grand Staircase–Escalante National Monument, 52 km (32 miles) east-southeast of ZION. In this case the dinosaur tracks possibly represented in both petroglyphs and pictographs were also preserved in nearby Kayenta Formation rocks along the same route through the cliffs (Stokes and Bruhn 1960; Lockley et al. 2006c). Kenworthy and Santucci (2006) presented an overview and cited selected examples of National Park Service fossils found in cultural resource contexts.

One of the authors of this document (VLS) has observed a tridactyl dinosaur track contained in the building stone of one of the park housing units.

Paleontological Resource Management, Preliminary Recommendations

- The paleontological resources inventory (DeBlieux et al. 2005) and monitoring report (Markle 2008) presented a number of recommendations, particularly in regard to future research and site monitoring. Several years have passed, so it would be useful to assess the state of action on these recommendations.

- ZION reports annually to GPRA Goal Ia9 (Paleontological Localities Condition) concerning 153 localities. A Geologic Resource Monitoring Manual by the Geological Society of America and NPS Geologic Resources Division includes a section on paleontological resource monitoring (Santucci et al. 2009). Fossil theft and vandalism are potential threats; theft of petrified wood and vertebrate fossils has occurred at fossil sites in the area, and tracks have been damaged by off-road vehicles and vandalized by amateur casting attempts using plaster and concrete (Milner et al. 2009).

- Park staff should be encouraged to observe exposed sedimentary rocks and associated eroded deposits for fossil material while conducting their usual duties. Staff should photodocument and monitor any occurrences of paleontological resources that may be observed *in situ*. Fossils and their associated geologic context (surrounding rock) should be documented but left in place unless they are subject to imminent degradation by artificially accelerated natural processes or direct human impacts. When opportunities arise to observe paleontological resources in the field and take part in paleontological field studies with trained paleontologists, park staff should take advantage of them.

- Within the NCPN, the Chinle Formation is exposed in ARCH, CANY, CARE, COLM, DINO, NABR, and ZION. A multi-park, cooperative research effort to study the ancient ecosystems preserved in the Chinle Formation throughout the NCPN and parks in the SCPN would provide invaluable information and increase understanding of this important fossil-bearing unit. A similar USGS–NPS project was recently completed for the Jurassic Morrison Formation ecosystems in 11 NPS units of the Intermountain Region. A project to redefine the stratigraphic nomenclature of the Chinle Formation exposures within Petrified Forest National Park is underway and could form a framework for continued Chinle investigation.

- Fossils found in a cultural context should be documented as other fossils, but will also require the input of an archeologist. Any fossil with cultural context may be culturally sensitive as well (e.g. subject to NAGPRA) and should be regarded as such until otherwise established. The Geologic Resources Division can coordinate additional documentation/research of such material.

- Future archeological excavations or infrastructure developments should consider scheduling site monitoring by a trained paleontologist in order to document and protect fossil resources.

- Fossil packrat middens are typically found in dry caves and rockshelters, and resemble piles or mounds of plant material with a dark glossy coating of crystallized packrat urine. Fossil middens can provide important paleoecological information. If one is located, there are several midden researchers in the Southwest who may be contacted. The GRD maintains a list of active researchers and can facilitate communication between the park and these researchers.

- Contact the NPS Geologic Resources Division for technical assistance with paleontological resource management issues.

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Data Sets

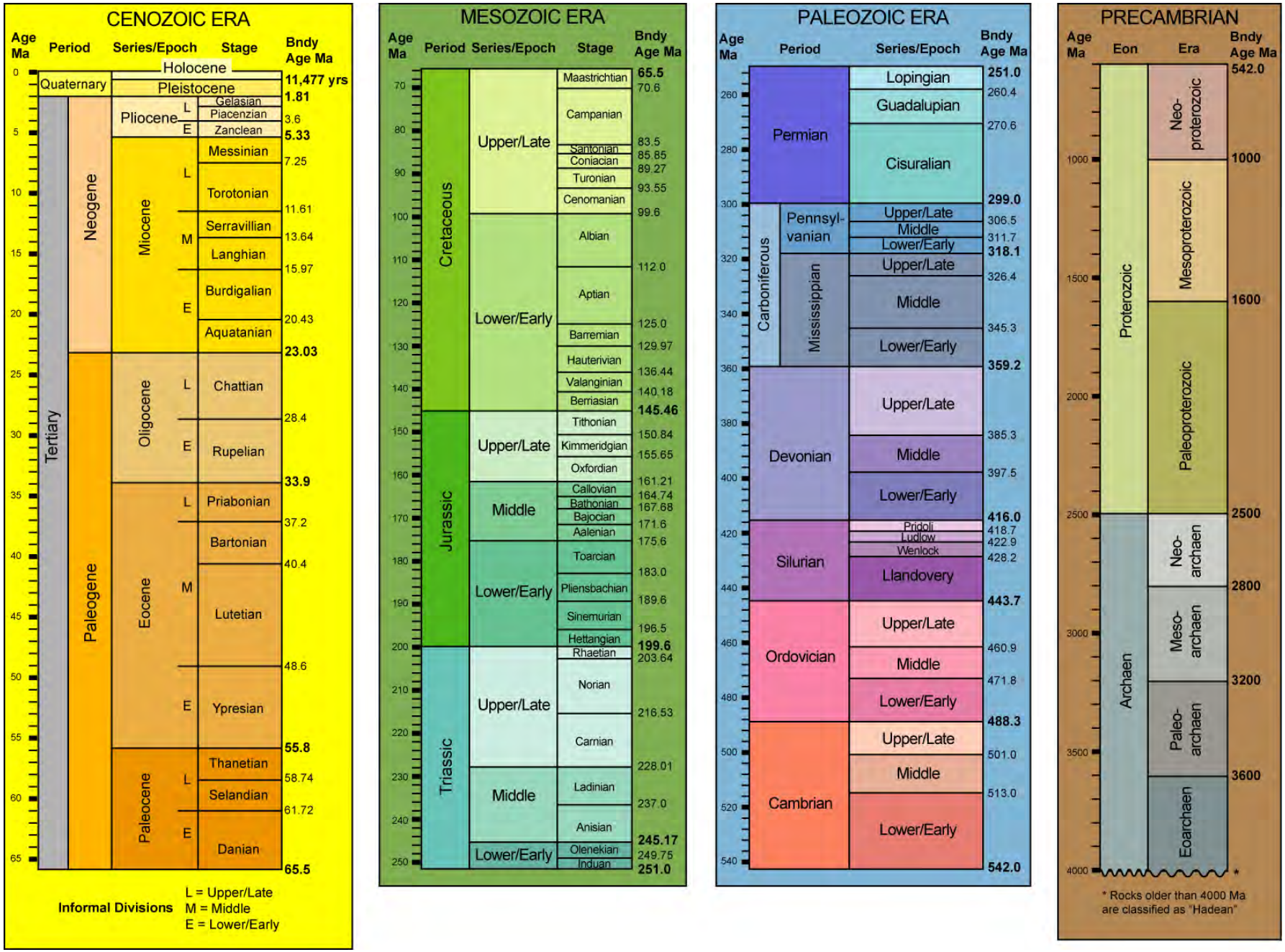
Zion National Park Paleontological Archives
5/1985–present. (hard copy data; reports; electronic data; photographs; maps; publications). Originated by Santucci, Vincent; status: Active.

Zion National Park Collections
1909–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by ZION staff; status: Active.

Zion National Park Digital Geologic Map
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University of California Museum of Paleontology
1935–present. (museum specimens; associated specimen notes; collection records; maps; etc.). Originated by UCMP staff; status: Active.



Appendix: Geologic Time Scale. *Ma = Millions of year old. Bndy Age = Boundary Age*. Colors are standard USGS colors for geologic maps. Modified from 1999 Geological Society of America Timescale (www.geosociety.org/science/timescale/timescl.pdf). Dates and additional information from 2004 International Commission on Stratigraphy (www.stratigraphy.org/gssp.htm) and USGS Fact Sheet 2007-3015 (http://pubs.usgs.gov/fs/2007/3015/).

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