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GRAND CANYON GEOLOGY

GRAND CANYON GEOLOGY

Second Edition

Edited by
Stanley S. Beus
Michael Morales

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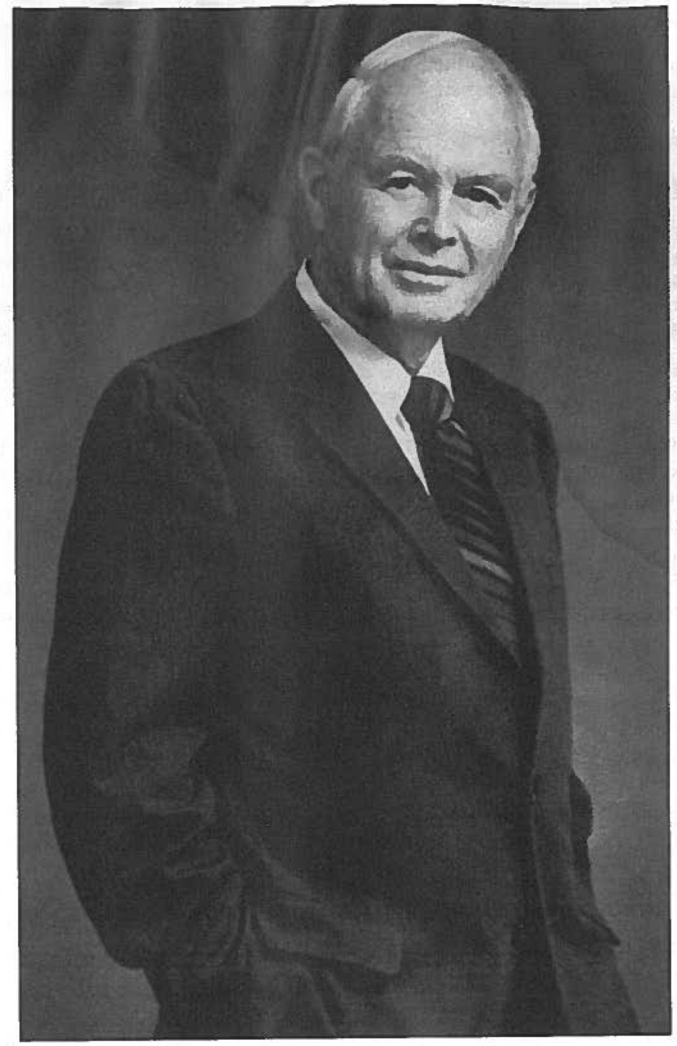
DEDICATION

This book is dedicated to the memory of Dr. Edwin Dinwoodie McKee (1906–1985), who stands as one of the premier scientists of the Grand Canyon in this century. Eddie began his career in 1929 as a park naturalist at Grand Canyon where for eleven years he made the natural history of this marvelous place a labor of love. In that time, he made extensive collections and observations in all areas of natural history, increased the interpretive program of the park, and became an expert on the biology, archaeology, and geology of the Grand Canyon. He also initiated studies of Paleozoic rock units that would occupy him, at least part time, for the rest of his career and would result in the publication of seven books on Grand Canyon geological history and four other monographs, together with more than 200 shorter articles, on geology or other topics of natural history.

His hiking exploits in the Grand Canyon became legendary—especially his frequent weekend journeys on foot from the south rim to the north rim and back (42 miles round trip) to woo and win a young biology student, Barbara Hastings. Barbara and Eddie were married in 1929 to begin a union that would last more than 50 years.

Following his years with the Grand Canyon National Park, Eddie served for a time as Chairman of the Geology Department at the University of Arizona, Tucson, as well as assistant director of the Museum of North Arizona, Flagstaff. In 1953 he began a 30-year career with the U.S. Geological Survey in Denver. From there, he conducted worldwide and world-class scientific investigations in stratigraphy, sedimentation, and depositional processes. These investigations included frequent returns to the Grand Canyon, where he continued his work on the Paleozoic strata.

An entire generation of geologists who have studied the Grand Canyon, including the authors of this book, have benefited from the stimulation, encouragement, and inspiration of Edwin D. McKee. A number of us have spent long-remembered days with him in the field. His tireless efforts to develop new insights, stimulate discussion, ask the right questions, and share the excitement and curiosity of scientific investigation serve as a model for all who would pursue the hidden secrets of the Grand Canyon.



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FOREWORD

It is strangely ironic that the Grand Canyon of the Colorado River, one of the most frequently visited natural wonders of the world, was for three centuries ignored by the Europeans who explored and exploited western North America. Of course, the Native Americans of the Southwest had known the canyon for thousands of years before some of them led a little band from the Coronado expedition to the south rim of the canyon in 1540. The Spaniards were impressed, and said so, but that was about the extent of their interest in this magnificent natural feature. During the long interval from that fateful day, four hundred and fifty years ago, when Don García López de Cárdenas and his companions gazed across the canyon until the third decade of the nineteenth century, very few people of European heritage visited or bothered to describe the canyon.

Then, in 1831 there appeared a description of the canyon by mountain man James Ohio Pattie, the first account to be written by an American. Yet his published portrayal of the canyon did not bring a rush of people to see it. It was hard to reach, and its breathtaking proportions made it a barrier to travel along the southwestern lines of latitude. Indeed, it was as a barrier that those Americans who knew anything about the canyon were prone to view it.

In the middle years of the nineteenth century, there were several United States Army surveying expeditions through the Southwest, in part to search for a feasible railroad route to the West Coast. One of the expeditions, led by Lieutenant Joseph Christmas Ives, reached the floor of the canyon near the mouth of Diamond Creek in 1858. The canyon seemed anything but inviting to Ives, who wrote that "the increasing magnitude of the colossal piles that blocked the end of the vista, and the corresponding depth and gloom of the gaping chasms into which we were plunging, imparted an earthly character to a way that might have resembled the portals of the infernal regions."

Fortunately, Ives had with him Dr. John Strong Newberry, one of the great pioneer American geologists, and it was Newberry who first looked at the canyon with a geologist's eye. To him the canyon did not look in the least like the gates of hell, for he wrote that it was "the most splendid exposure of stratified rocks that there is in the world."

Then, in 1869 and 1871 John Wesley Powell led his exploring parties down the Colorado River in their frail, wooden boats, penetrating the canyon from Green River to the Grand Wash Cliffs. Suddenly, people throughout the world were made aware of this great natural wonder. From then until the present day, the canyon has been scientifically explored and studied, so that there now exists a vast body of literature on all aspects of the canyon. All of this has taken place within two lifetimes.

This I know in a personal way, for one evening in the thirties I sat on a couch in the Explorers' Club, New York, with none other than Frederick Dellenbaugh, one of the members of Powell's second expedition. For an hour or more, I enjoyed a firsthand account of Powell and his party and of their adventures.

Grand Canyon Geology is dedicated to Edwin D. McKee—"Eddie" to all of us who knew him. No more appropriate dedication could be made. Eddie was the foremost modern authority on the geology of the Grand Canyon. Geologists today and in years to come will benefit from the profound studies of canyon geology that have come from Eddie's pen.

This book is not intended for the casual reader; he who turns its pages will need a certain amount of geological sophistication to appreciate the twenty chapters that describe and interpret Grand Canyon geology. Because each chapter is written by one or more authorities in their respective fields, there is here a storehouse of geological knowledge, much of it new. For geologists and those interested in geology, this is a book to be read, consulted, and treasured for years to come.

Edwin H. Colbert

GRAND CANYON GEOLOGY

• 1 •

Introducing the Grand Canyon

Stanley S. Beus and Michael Morales

The Grand Canyon contains a marvelous record of geologic and paleontologic events spanning most of the last two billion years, nearly one half of the life span of this planet. Although it is neither the deepest nor the longest canyon in the world, it is one of the few places where so many chapters of earth history are clearly displayed. The igneous and metamorphic rocks of the canyon's inner gorge are part of the basement of the North American continent. Two great packages of sedimentary and volcanic rocks—one of Middle Proterozoic age and one of Paleozoic age—make up most of the canyon's walls. Beyond the rim to the north and east are the giant stairsteps of Mesozoic and lower Cenozoic strata whose cliffs have been retreating from the canyon's edge for millions of years. In the central part of the canyon, upper Cenozoic volcanic rocks have poured intermittently into the gorge, temporarily blocking the Colorado River and altering its cutting action. During the past two billion years, many dramatic, earth-changing events have occurred in western North America. We are incredibly fortunate that the rocks exposed in the Grand Canyon record this time period like no other place on the continent. This great chasm is truly a unique "window into the past."

The Grand Canyon (Fig. 1.1) lies entirely in the northwestern part of Arizona. It extends nearly 278 miles (448 kilometers) between Lake Powell on its eastern end and Lake Mead to the west. In 1919, the region became a national park, which today encompasses approximately 1900 square miles (4921 square kilometers) of land. This most famous of all canyons was formed by swiftly flowing waters of the Colorado River cutting into rock layers of the southwestern Colorado Plateau, a vast uplifted tableland that includes a large portion of the Four Corners states: Arizona, Colorado, New Mexico, and Utah. The land surrounding the Grand Canyon includes six local plateaus and one low-lying platform, all of which are bounded by faults or monoclines (Fig. 1.2). In a west-to-east cross section to the north of the canyon (Fig. 1.3), you can see how each of these blocks of land has been uplifted, downdropped, or tilted relative to its neighbor.

At its narrowest, the Grand Canyon is a little less than a mile (1.6 kilometers) across. Along the canyon's north and south rims, the relief is relatively gentle, except for incisions made by runoff waters flowing into the gorge. Within the canyon itself, however, the topography is quite varied and spectacular. The maximum depth of the canyon at any single place is about 6000 feet (1829 meters) from the rim to the floor. The maximum drop in elevation of the canyon as a whole, however, is approximately 6600 feet (2012 meters) between Point

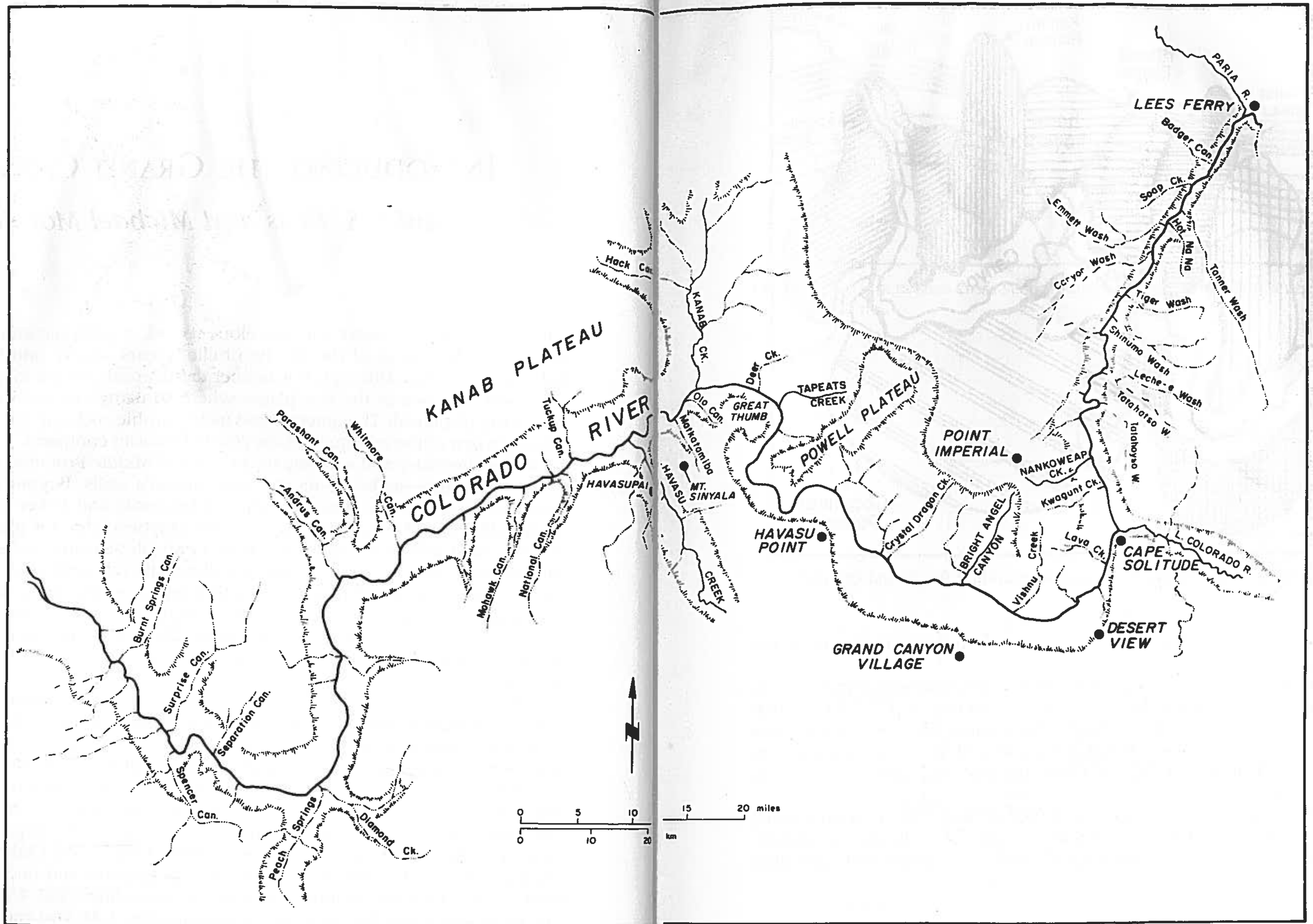

Figure 1.1. Map of the Grand Canyon.

Imperial on the north rim (8803 feet [2683 meters]) and the floor of the canyon near Lake Mead (1200 feet [366 meters]).

Many people are surprised to learn that there is a difference in elevation between the north and south rims. On the canyon's southern edge, the altitude ranges from 6000 to 7500 feet (1829 to 2286 meters) above sea level. The northern edge, however, is 1000 to 1200 feet (305 to 366 meters) higher even though both rims are capped by the same rock unit, the Kaibab Formation. The difference in height occurs because the various rock layers into which the canyon has been cut do not lie completely flat in this region. Instead, they arch or dome upward. The Colorado River carved the canyon through the southern flank of the dome, where the rock layers are tilted gently down to the south. Layers on

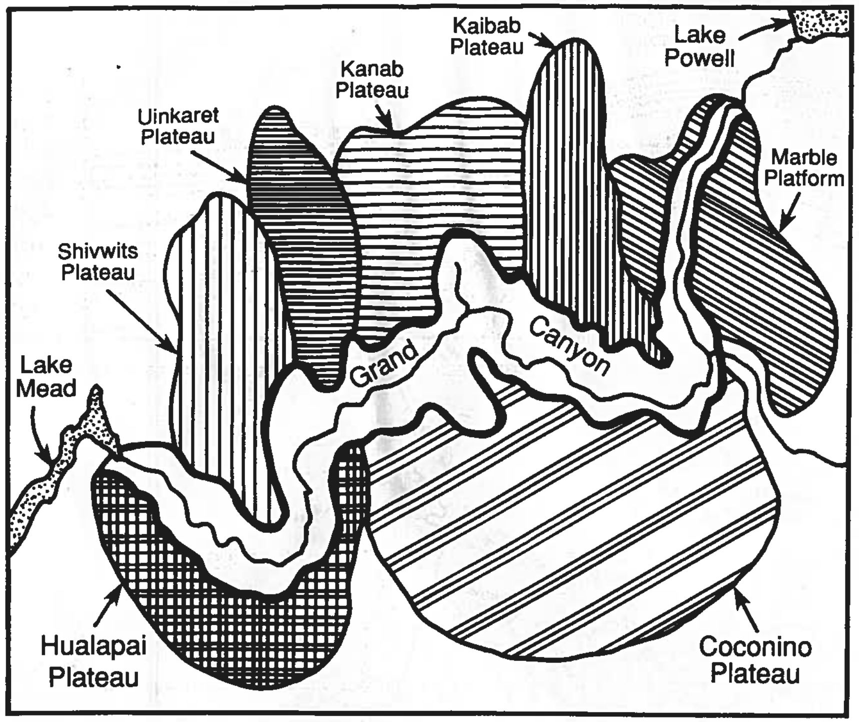

FIGURE 1.2. Generalized map of the area surrounding the Grand Canyon.

the north rim, therefore, have been uplifted higher than the same layers on the south rim (Fig. 1.4).

Water that flows into and through the Grand Canyon comes primarily from four merging rivers—the Green, San Juan, Little Colorado, and Colorado—that drain hundreds of square miles of the Four Corners states (Fig. 1.5). As it courses through the canyon, the Colorado River drops about 2000 feet (610 meters) in elevation (Fig. 1.6). This steep gradient allows the river to continue its erosion of the chasm's floor.

Humans have been in the Grand Canyon for at least four thousand years, but the early records are sparse. Rare artifacts, remains of prehistoric dwellings, and the numerous mescal pits all attest to the early exploration and habitation

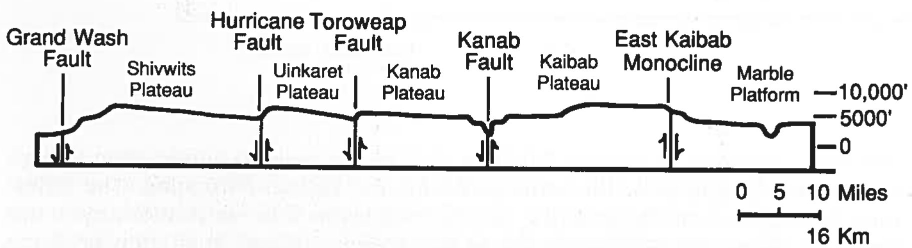

FIGURE 1.3. Generalized west-to-east cross section of the area just north of the Grand Canyon.

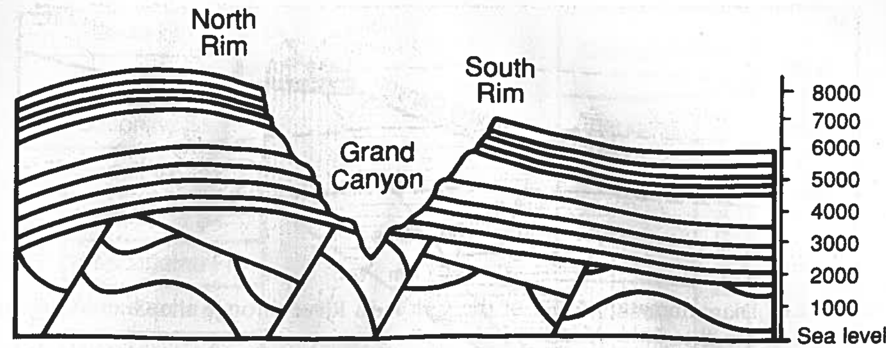

FIGURE 1.4. Generalized cross section through the Grand Canyon from the north rim to the south rim.

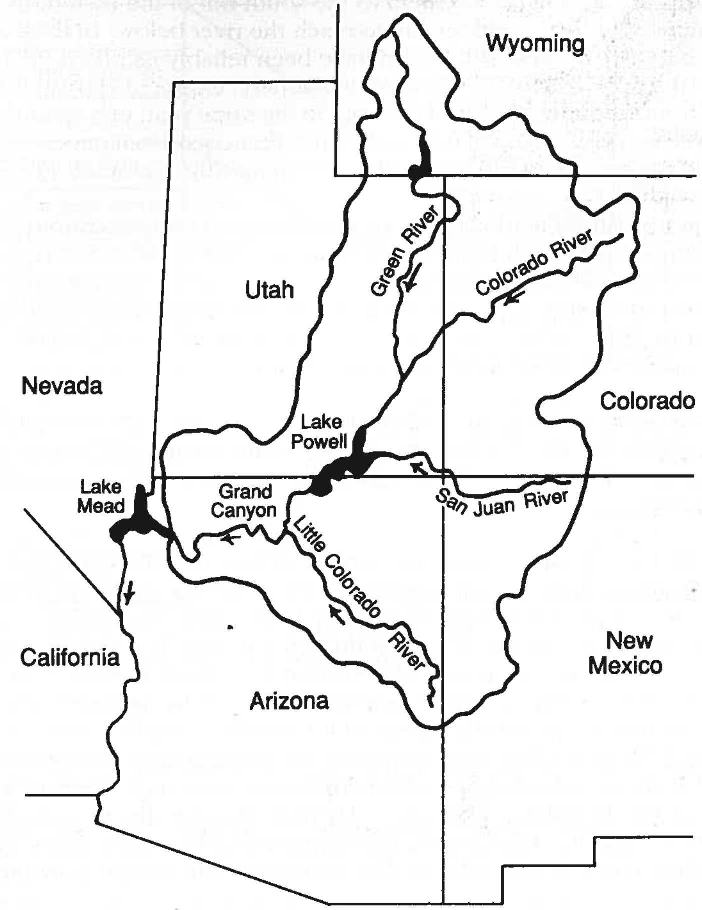

FIGURE 1.5. Drainage area for runoff water that flows into and through the Grand Canyon.

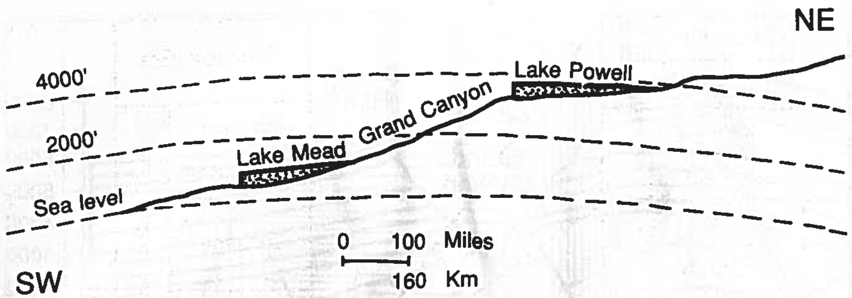

Figure 1.6. Diagrammatic profile of the Colorado River through the Grand Canyon, from Lake Powell to Lake Mead.

by American Indians such as the Anasazi and the Cohonina. The first Europeans to see the canyon were a Spanish party of thirteen in search of the fabled lost cities of gold, under the command of Captain Don García López de Cárdenas. In 1540, Hopi Indian guides led them to the south rim in the eastern part of the Grand Canyon, but they were unable to reach the river below. In the next three centuries, only two visits to the region have been reliably recorded. In 1776, Father Francisco Tomás Garcés, a Spanish missionary, explored Havasu Canyon in the south-central part of the Grand Canyon. In the same year, two Spanish priests, Father Silvestre Vélez de Escalante and Father Francisco Domíngues, led an expedition to the region and discovered a ford across the Colorado River (Crossing of the Fathers).

Among the earliest geological reports of the Grand Canyon country are those of Jules Marcou (1856) and John Strong Newberry (1861), who described the region's Paleozoic stratigraphy from their explorations of the canyon and the land to the south. Newberry was a geologist in the War Department-sponsored Ives expedition of 1857–1858. Ives' rather discouraging and, as it turned out, unprophetic statement about the Grand Canyon was:

> Ours has been the first, and will doubtless be the last, party of whites to visit this profitless locality. It seems intended by nature that the Colorado River, along the greater portion of its lonely and majestic way, shall be forever unvisited and undisturbed.

In 1869, John Wesley Powell led a party of ten men (reduced later to nine and finally only six) on an epic journey by boat down a thousand miles of the Colorado River from Green River, Wyoming, across Utah, and finally through the Grand Canyon to the mouth of the Virgin River at what is today the north end of Lake Mead. Powell's work was followed up by a small group of outstanding scientists through the turn of the nineteenth century. These included G.K. Gilbert, who was the first to apply formal rock unit names to Grand Canyon rocks; C.E. Dutton, who wrote the first monograph on the geology and geologic history of the Grand Canyon; A.R. Marvine, who participated in the U.S. Geographical Survey West of the 100th Meridian; and C.D. Walcott, who described both Paleozoic and Precambrian rocks in the canyon's central and eastern parts. These pioneer studies laid the groundwork for all subsequent research in the Grand Canyon.

Edwin D. McKee, to whom this book is dedicated, stands out as the premier research scientist of Grand Canyon geology in the twentieth century. Be-

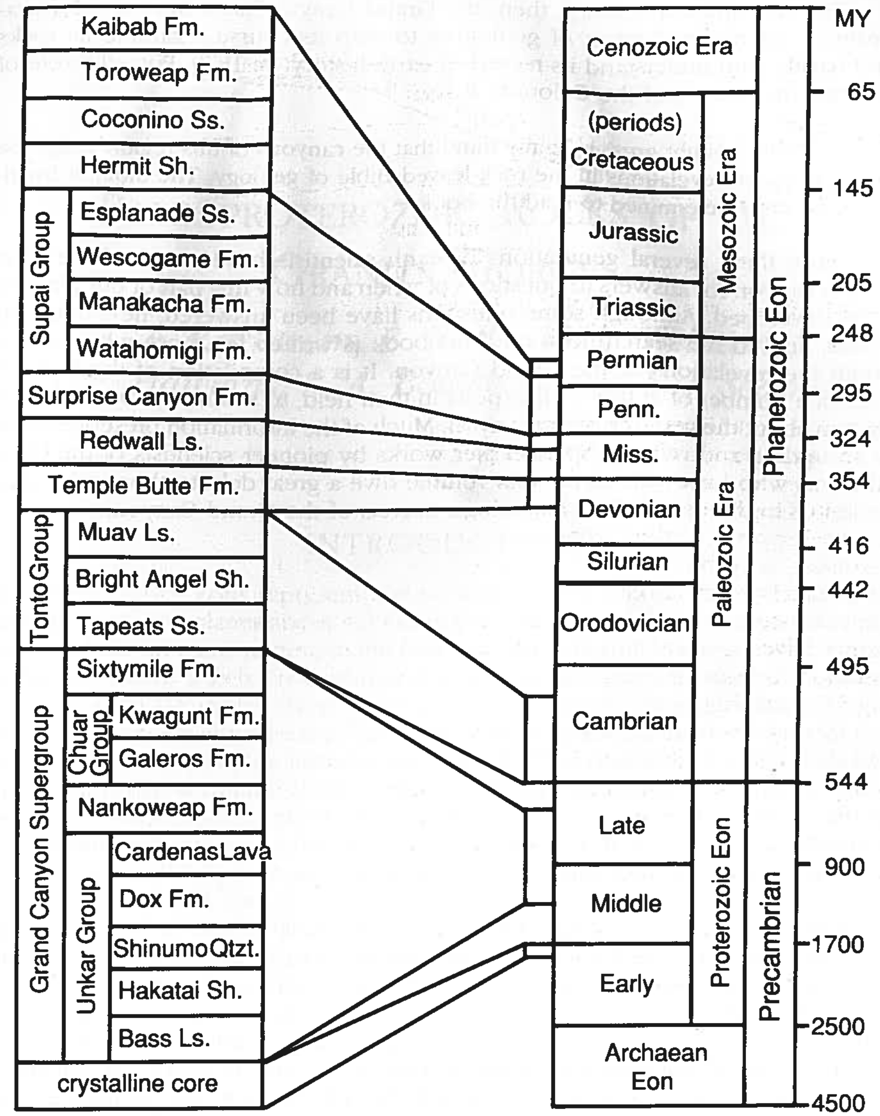

Figure 1.7. Comparison of the geologic column of the Grand Canyon with the Geologic Time Scale. (After Haq and Van Eysinga 1987.)

tween 1933 and 1982, McKee either authored or coauthored five monographic publications on various Paleozoic rock units of the canyon—including the Redwall Limestone, Supai Group, Coconino Sandstone, the Toroweap and Kaibab Formations, and Cambrian units. McKee also organized a special symposium in 1964 at the Museum of Northern Arizona to summarize the data and interpretations on the origin and evolution of the canyon. Much of our present understanding of the stratigraphy and age of Grand Canyon rocks (Fig. 1.7) is based on McKee's seminal work.

For more than a century, then, the Grand Canyon has attracted the extraordinary interest and effort of geologists to map its course, examine its rocks and fossils, and understand its record of earth history. In 1870, Powell wrote of the canyon country of the Colorado River:

> . . . the thought grew in to my mind that the canyons of this region would be a Book of Revelations in the rock-leaved Bible of geology. The thought fructified and I determined to read the book.

Since then, several generations of earth scientists have searched the river and its canyon for answers to questions of when and how this part of our planet's crust developed. Although some questions have been answered, new ones are raised, and so the search goes on. This book is written for those who inquire about the revelations of the Grand Canyon. It is a compilation of the best efforts of a number of authors, all experts in their field, to summarize what is now known about the geology of the canyon. Much of the information presented here is an update and refinement of earlier works by pioneer scientists of the past. All of us who have written for this volume owe a great debt to those who preceded us in the search for the geologic secrets of the Grand Canyon.

• 2 •

Paleoproterozoic Rocks of the Granite Gorges

K. E. Karlstrom, B. R. Ilg, M. L. Williams, D. P. Hawkins, S. A. Bowring, and S. J. Seaman

INTRODUCTION

The Precambrian Eons represent eight-ninths of earth history, from the time of formation of our planet about 4.6 Ga (billion years ago), to the rapid diversification of life on earth starting in the Cambrian 545 Ma (million years ago). North America records a rich Precambrian geologic record, from the oldest rocks on earth, 4.0–3.5 Ga in the Slave and Wyoming provinces, through younger belts that were sequentially added to the south (Fig. 2.1). The Archean nucleus of the continent had a complex history, including the development of a rifted shoreline in southern Wyoming about 2 billion years ago (Karlstrom et al. 1983). However, there was no continental lithosphere in the Grand Canyon region before ~1.85 billion years ago, when volcanic arcs began to form on oceanic crust.

The Grand Canyon region provides a spectacular record of the growth of the North American continent in the Paleoproterozoic Era (2.5–1.6 Ga), particularly within a 200-million-year interval from 1.8 to 1.6 Ga. To understand the history of this continental growth, we rely on detailed studies of the age and character of ancient rocks, analyzed in the context of modern plate tectonic analogues. The best modern analogue for growth of continents may be the Indonesian region, where the Australian plate is moving north and colliding with volcanic arcs and back arc basins across a very complex series of subduction zones and transcurrent faults. We envision that the Grand Canyon region resembled the Indonesian region at about 1.75 Ga, with tectonic blocks (terranes) consisting of volcanic island arcs, arc basins, and older continental fragments poised to the "south" and ready to become welded to the continental nucleus during progressive arc-continent collisions (Condie 1984; Karlstrom and Bowring 1988).

This plate tectonic model (Fig. 2.2) suggests that Proterozoic continental crust of the southwestern Unites States was differentiated from the mantle in magmatic arcs above subduction zones. In the Grand Canyon, these arcs range in age from 1.84 to 1.71 Ga. We infer that oceanic island arcs approached each other, and the older Archean part of the North American continent, as a result of subduction of the dense intervening oceanic crust and back-arc basins (Fig. 2.2a). This collision process resulted in the welding together of less dense materials (sediments, volcanics, and plutonic rocks) into new continent (Fig. 2.2b). The main orogeny, or "mountain building" episode, lasted from 1.74 to 1.65 Ga in the

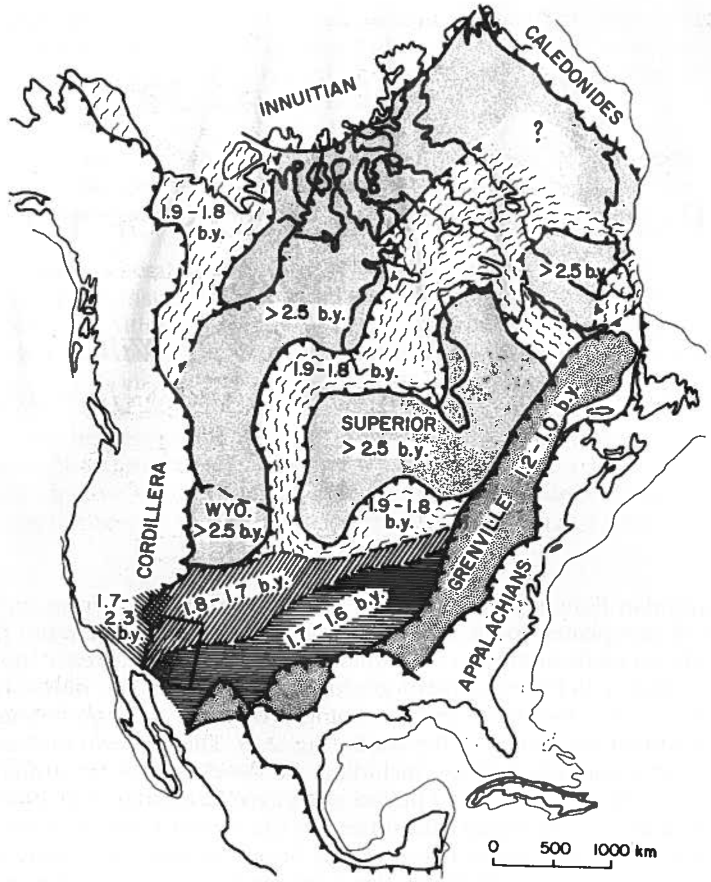

FIGURE 2.1. North American age provinces. The wide belt of Proterozoic crust in the southwestern United States was added to the Archean nucleus of the continent from 1.8 to 1.6 Ga. (Adapted from Hoffman 1988.)

Grand Canyon region, and it culminated with intense contractional deformation at 1.7–1.68 Ga. By 1.65 Ga, rocks were complexly deformed, metamorphosed, and beginning to cool off at depths of 10 km in the middle crust of the thickened orogen (Fig. 2.2b). This type of orogeny did not create high mountains like the Himalayas. Instead, thickening of originally thin crustal fragments appears to have resulted in a relatively low-elevation orogen, like the Indonesian region today (Bowring and Karlstrom 1990).

A continuing goal of our work in the Grand Canyon region is to identify the once-separate arc terranes, the tectonic boundaries between terranes (sutures), and the timing and geometry of collisions. Beyond tectonic models, our research is also designed to simply map, describe, and analyze the rocks in ways that will allow us to understand deep crustal processes, such as the symbiotic interactions between deformation, metamorphism, and plutonism (Williams and Karlstrom 1996; Karlstrom and Williams 1998). In working on such complexly

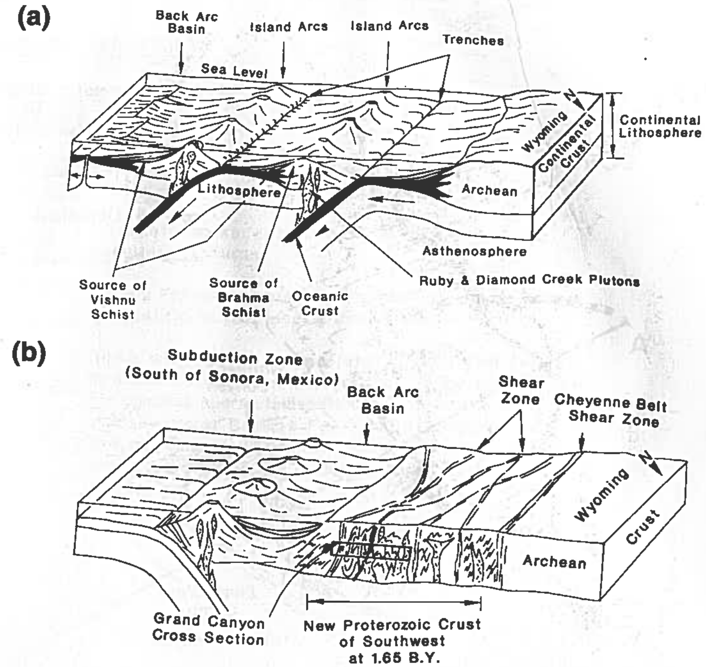

FIGURE 2.2. Plate tectonic model for assembly of Proterozoic crust in the Southwest. (a) Hypothetical geometry of island arcs south of the Archean nucleus at about 1.74 Ga. (b) The continent grew southward by assembly of arcs and accompanying compressional deformation between 1.74 and 1.65 Ga. Rocks of the Granite Gorge were at depths of 10–20 km in the middle crust (box) during assembly and stabilization of the continent.

deformed and metamorphosed rocks, it is important to realize that we cannot directly witness processes that operate at the 10- to 20-km depths from which these rocks have come, nor can we simulate in the laboratory the slow rates and interacting processes that went on during the 100-million-year plate collision period. Thus, many questions about the middle crust can only be answered by detailed collaborative studies such as this one. We approach the metamorphic rocks with a hope that while modern analogues may help us understand ancient mountain belts, the reverse is also true and the old eroded orogens offer a chance to understand tectonic processes that operate deep in the earth today.

PREVIOUS WORK AND NOMENCLATURE

The east–west trending Grand Canyon transect presents spectacular exposures of Paleoproterozoic rocks for 200 km across the ancient orogenic belt and our only window into the nature of the Precambrian rocks under the Colorado Plateau (inset to Fig. 2.3). In the Upper Granite Gorge, these rocks are continuously exposed from river mile 78 to 120 (Figs. 2.3 and 2.4). The Middle Granite Gorge has discontinuous exposures from mile 127 to mile 137. An isolated outcrop oc-

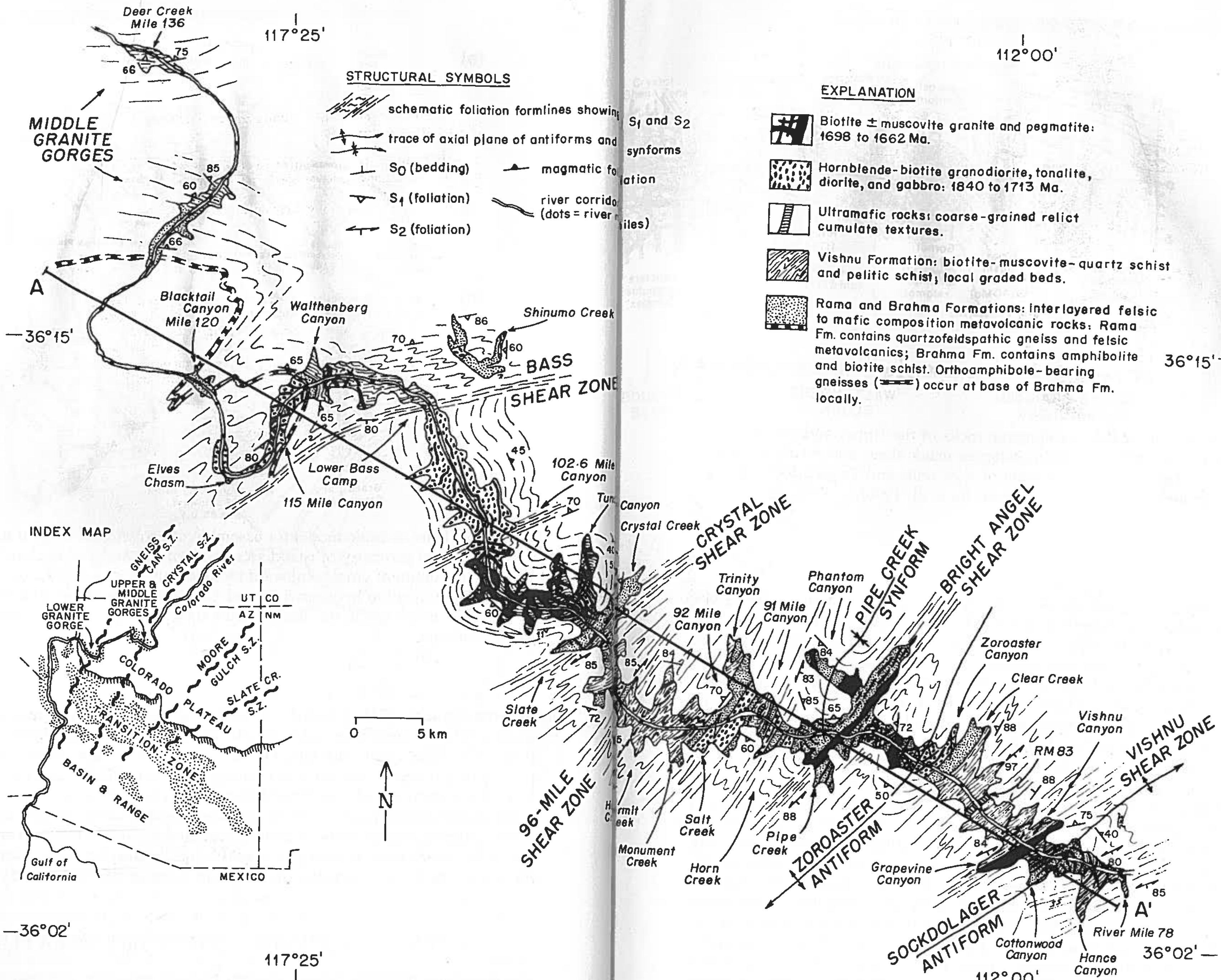

FIGURE 2.3. Geologic map of Paleoproterozoic rocks of the Upper and Middle Granite Gorges showing rock types, foliation patterns, and major shear zones. Index map shows location of Upper + Middle and Lower Granite Gorge transects, major physiographic provinces, Proterozoic rocks (stipple), and regional shear zones (S.Z.). (Adapted from Ilg et al. 1996).

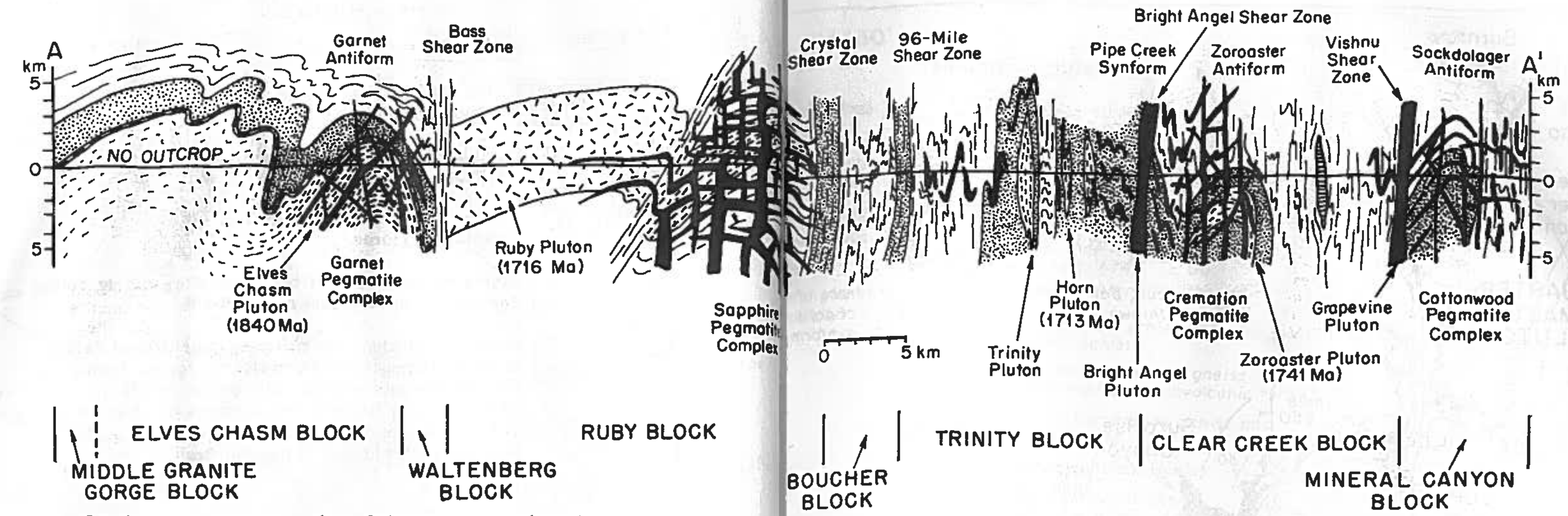

FIGURE 2.4. Cross section of Paleoproterozoic rocks of the Upper and Middle Granite Gorges showing rock types, foliation patterns, major shear zones and folds, and tectonic blocks; see Fig. 2.3 for explanation of rock units and Table 2.2 for description of blocks and shear zones. (Adapted from Ilg et al. 1996).

curs near the Hurricane Fault at mile 190–191. The Lower Granite Gorge contains near-continuous outcrops from mile 207 to mile 261 (Figs. 2.5 and 2.6).

To John Wesley Powell (1876), the Precambrian "granite" and "Grand Canyon schists" were dreaded because these harder rocks were associated with a narrower river and more difficult rapids. Walcott (1889) identified the Vishnu "terrane" as a complex of schist and gneiss. Since then, workers have continued to refine the subdivisions of the Precambrian rocks. Noble and Hunter (1916) provided the first detailed petrologic work by descending into side canyons from the south Tonto Platform (from Garnet Canyon east to Red Canyon). They identified domains of contrasting rock packages and recognized that differences reflected the presence of both metasedimentary and intrusive igneous rocks and that some of the gneisses possibly were a basement on which the metasedimentary units were deposited. Campbell and Maxson (1938), following six years of work and a three-boat expedition, identified different mappable units: Vishnu "Series" and Brahma "Series" (Maxson 1961). However, Campbell and Maxson underestimated structural complexities and probably overestimated stratigraphic thickness when they proposed that the combined stratigraphic sequence of metasedimentary and metavolcanic rocks was 8–16 km thick. This stratigraphic approach was called into question by Ragan and Sheridan (1970), who noted complex folding and interfingering of schist and amphibolite in the Phantom Ranch area. Subsequently, Brown et al. (1979) also emphasized the complex deformational features and lumped all of the metasedimentary and metavolcanic rocks under the name "Vishnu Complex."

Our approach (Ilg et al. 1996) recognizes the need to simultaneously pursue both tectonic and stratigraphic subdivisions of the Proterozoic rocks. The tectonic perspective emphasizes that the complex deformation and metamorphism in these rocks make it impossible to measure thickness and to confidently reconstruct regional stratigraphy. Furthermore, there are important shear zones boundaries of unknown displacement, but some possibly representing suturing of separate tectonic blocks that were once hundreds to thousands of kilometers apart (e.g., Crystal shear zone). Thus, we cannot be certain that all metasedimentary schists are strictly correlative (same age, same basin, same depositional sequence). Nevertheless, our work to date permits a stratigraphic interpretation (Noble and Hunter 1916) that metasedimentary rocks across the transect are broadly of similar rock type and age (1.75–1.73 Ga). Thus, the entire metasedimentary–metavolcanic package could have been deposited within 10–20 m.y. in a single arc basin or in similar, but tectonically separate, basins.

We proposed a new name, the Granite Gorge Metamorphic Suite, for metasedimentary and metavolcanic units in the Grand Canyon (Ilg et al. 1996). According to the U.S. Code of Stratigraphic nomenclature (Henderson et al. 1980), the term "metamorphic suite" can have the stratigraphic significance of "group" or "supergroup" and hence can be subdivided into mappable formations or groups. However, the term also implies complex structures, high-grade metamorphism, and difficulty in unraveling original stratigraphic relationships. We assign names for major mappable rock types in the Upper Granite Gorge: Brahma Schist for mafic metavolcanic rocks (after the Brahma "series" of Campbell and Maxson 1938) and Rama Schist for felsic metavolcanic rocks. We reserve the term Vishnu Schist for metamorphosed sedimentary rocks, as probably intended by Walcott (1894), recommended by Noble and Hunter (1916, their Vishnu schist), and proposed by Campbell and Maxson (1938, their Vishnu "Series"). Rocks in the Lower Granite Gorge also fall into these three basic lithologic groups.

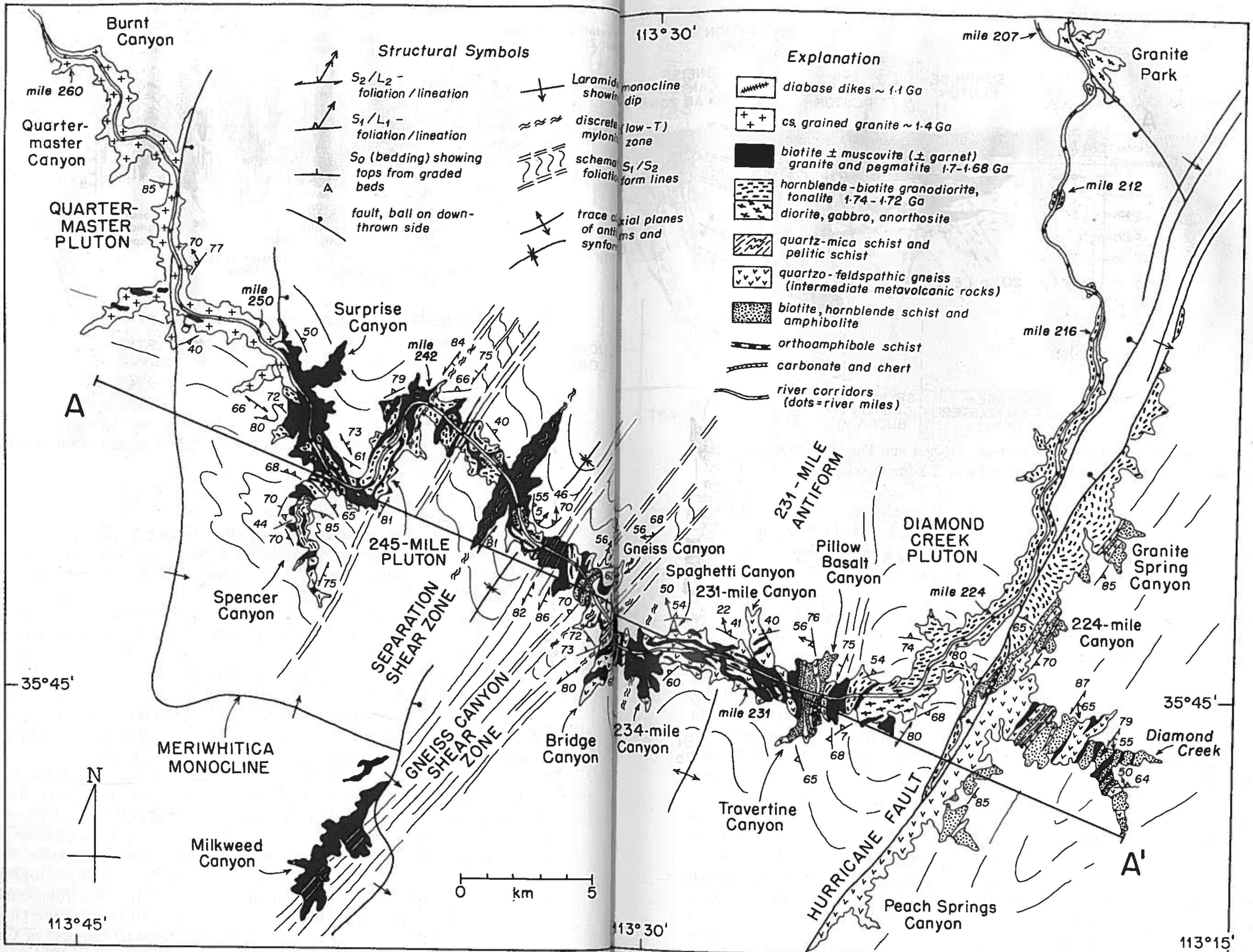

FIGURE 2.5. Geologic Map of the Proterozoic rocks of the Lower Granite Gorge showing rock types, foliation patterns, and major shear zones.

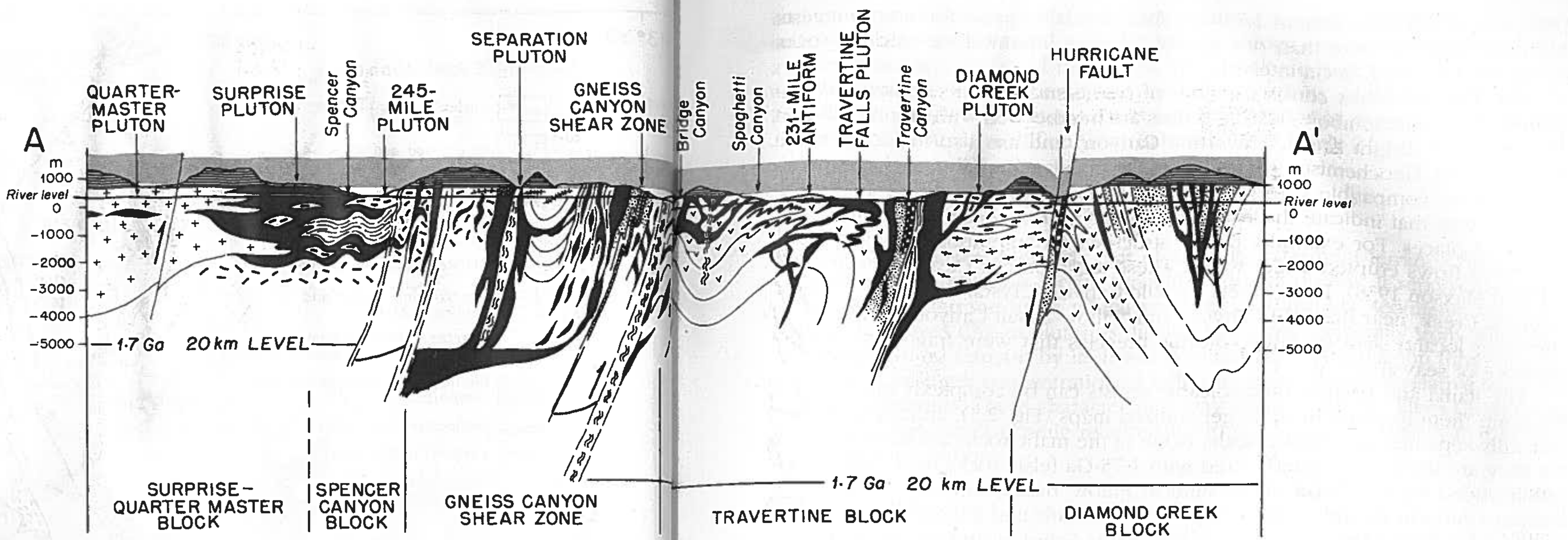

Figure 2.6. Cross section of the Lower Granite Gorge; see Fig. 2.5 for explanation of rock units and structural symbols, and see Table 2.2 for description of blocks and shear zones.

OLDER BASEMENT—1.84-GA ELVES CHASM PLUTON

All sedimentary and volcanic rocks must be deposited on some older substrate or "basement," but "basement" and "sedimentary cover" often get detached from each other and tectonically interlayered during deformation. High-grade metamorphism also obscures the nature of the original sedimentary and volcanic protoliths and the original contact relationships. Noble and Hunter (1916) posed this problem and speculated that some of the gneisses of the Grand Canyon might be basement for the schists. Subsequent workers recognized that gneisses are deformed intrusive rocks (Campbell and Maxson 1938; Brown et al. 1979; Babcock 1990), but there was still debate on whether the Elves Chasm and Trinity gneisses are possible basement for the Vishnu schist.

New mapping and geochronology indicate that the "gneisses" have different ages in different areas. The Trinity "gneiss" (Babcock et al. 1979) and the Zoroaster "gneiss" (Lingley 1973) are deformed and metamorphosed granodiorites that intruded into the Granite Gorge Metamorphic Suite. U-Pb zircon geochronology by Hawkins et al. (1996) shows that the Trinity pluton is 1.73 Ga and that the Zoroaster pluton is 1.74 Ga.

In contrast, the Elves Chasm pluton is 1.84 Ga, the oldest rock known in the southwestern United States and apparently the basement for the turbidites of the Vishnu Schist. The contact zone between the Elves Chasm pluton and the overlying Granite Gorge Metamorphic Suite is exposed in several areas—notably at Walthenberg, 113-mile, and Blacktail canyons—and several places in the Middle Granite Gorge (Fig. 2.3). The contact is gradational over an interval of several meters between the foliated pluton and a distinctive orthoamphibole-bearing gneiss. The composition of this gneiss is unusual (high Ca/K ratio; Clark 1979). Babcock (1990) suggested this alteration took place during weathering, and new mapping suggests that the several occurrences of orthoamphibole gneiss in the Upper and Middle Granite Gorges may represent the same weathering zone repeated by folding (Figs. 2.3 and 2.4). This zone separates the 1.84-Ga pluton from Vishnu Schist that contains 1.75-Ga detrital zircon (D.P. Hawkins, unpublished data from Walthenberg Canyon), indicating that the Vishnu sediments were probably deposited on the older basement.

GRANITE GORGE METAMORPHIC SUITE (1.75–1.73 GA)

Metasedimentary and metavolcanic rocks of the Granite Gorge Metamorphic Suite make up about half of the exposed rocks in the Grand Canyon, the rest being intrusive rocks (discussed below). Overall, the Granite Gorge Metamorphic Suite consists of a variety of schists containing quartz, feldspar, micas, and other metamorphic minerals. In the following discussions, we use descriptive metamorphic rock names for rocks we see in outcrop or thin section. We also infer the original sedimentary or volcanic "protolith" based on rock composition and a limited number of primary structures that survived the deformation and metamorphism.

The Rama Schist includes quartzofelspathic schist and gneiss with locally preserved phenocrysts of quartz and feldspar and possible relict lapilli (e.g., in Shinumo and Diamond Creek) that suggest a felsic to intermediate volcanic origin. Felsite layers within Brahma amphibolite at Clear Creek yield a zircon date of 1.75 Ga, quartzofelspathic gneiss in the Sockdologer antiform give a date of 1.74 Ga, and an intermediate-composition dike in "Pillow Basalt Canyon" (mile

229.5, Fig. 2.5) gives a date of 1.73 Ga. Thus, available geochronology suggests that there may be more than one age of felsic to intermediate volcanic rocks within the 1.75- to 1.73-Ga interval.

The Brahma Schist consists mainly of hornblende–biotite schists and amphibolites. Orthoamphibole-bearing schists are interbedded with amphibolites in several places (Bright Angel, Travertine Canyon) and are also included in the Brahma Schist. Geochemistry suggests that the massive amphibolites have tholeiitic character, compatible with an origin as island arc basalts (Clark 1979). Primary features that indicate the nature of the depositional setting are preserved in several places. For example, pillow structures in amphibolites indicate that mafic lava flows erupted under water. These are found in Clear Creek (Campbell and Maxson 1938), Horn Creek, 92-Mile Canyon, Crystal Creek, Slate Creek, Shinumo Creek, near Blacktail Canyon, and "Pillow Basalt Canyon" (mile 229.5). The latter locality also contains volcanic breccias that were fragmented in the presence of seawater.

The Rama and Brahma metavolcanic schists can be complexly interlayered. We lump them together in some generalized maps (Fig. 2.3), although they are generally separable at 1:24,000 scale. None of the mafic rocks are directly dated, but they are intruded or interlayered with 1.75-Ga felsic rocks near Clear Creek and intruded by a 1.73-Ga dacite dike in Pillow Basalt Canyon. Contact relationships support variable relative ages between mafic and intermediate metavolcanic rocks. In the Upper Granite Gorge, Rama Schist is underneath and older than the Brahma Schist in the Sockdolager antiform and Shinumo areas. In the Diamond Creek and Travertine areas of the Lower Granite Gorges, intermediate volcanic rocks and dikes appear younger than the amphibolites based on younging of pillows and intrusion of the 1.73-Ga dacite dikes into amphibolites.

The Vishnu Schist consists of thick sections of quartz-mica schist and pelitic schist that are interpreted to be metamorphosed sandstones and mudstones that were deposited in submarine conditions on the flanks of the eroding oceanic islands. Thick (several kilometers) sections show rhythmic bedding and graded bedding suggesting deposition as submarine turbidites. The absence of conglomerates and general fine grain size suggests a lack of high-energy proximal facies. Calc silicate pods are numerous and are interpreted to be concretions. Preserved graded bedding indicates that it was deposited stratigraphically above the Brahma Schist in the Walthenberg area. Detrital zircon crystals from the Vishnu Schist of the Upper Granite Gorge range from 3.3 Ga to 1.736 Ga. The older grains were probably transported from the Archean Wyoming province; the younger grains (1.74 Ga) are likely close to the depositional age of the Vishnu Schist, because these rocks overlie the 1.74- to 1.75-Ga Brahma Schist and were cross cut by the 1.74-Ga arc plutons (Trinity and Zoroaster plutons) that contain schist and calc-silicate xenoliths.

INTRUSIVE ROCKS

Intrusive rocks make up the other half of the crystalline rocks of the Grand Canyon. Campbell and Maxson (1933) thought there was a single major period of igneous "invasion," and this led to the convention of lumping all plutonic rocks of the Grand Canyon under a single name such as "Zoroaster Gneiss" (Campbell and Maxson 1933), "Zoroaster Granite" (Maxson 1968), and "Zoroaster Plutonic Complex" (Babcock et al. 1979). However, new mapping (Ilg et al. 1996) and geochronology (Hawkins et al. 1996) show that plutonic rocks record a long

and complex evolution of the crust. For example, magma crystallized at numerous times between 1.84 and 1.37 Ga, and petrology and geochemistry indicate a wide variation in types of intrusions. Thus, a single rock name is misleading, and we prefer to use the names of individual plutons or dike swarms (Table 2.1). We subdivide the intrusive rocks into four groups of plutons, based on age and tectonic groups (modified from Babcock 1990; Table 2.1).

Older Basement

The 1.84 Elves Chasm pluton is part of an older basement terrane. This pluton is dominantly hornblende–biotite tonalite to quartz diorite. It represents a part of an older arc terrane upon which younger arcs and arc sediments were deposited. The Elves Chasm pluton is distinguished geochemically from other plutons in the Grand Canyon by its lower concentration of large ion lithophile elements and its lower concentration of light rare earth elements relative to heavy rare earth elements. These characteristics may suggest that the Elves Chasm pluton bears a less direct genetic relationship to slab subduction than that shown by younger arc plutons.

Arc Plutons

The 1.74- to 1.71-Ga granodiorite (plus gabbro-diorite) complexes are interpreted as arc plutons, compatible with petrologic and geochemical data (Babcock 1990). These plutons formed from melting above the subducting plate, then melts rose to form large magma chambers that, in turn, fed volcanic eruptions within the island arcs. We infer that they were emplaced at shallow levels in arcs because of an absence of contact metamorphic aureoles, but their original shape is not preserved. Some are now large folded sheet-like plutons (Zoroaster, Trinity, and Ruby plutons), whereas others are massive differentiated plutons (Diamond Creek pluton) or smaller stock-like bodies (Pipe Creek, Horn Creek, Boucher, and Crystal plutons). One characteristic of many of these plutons is the presence of enclaves of a range of compositions (gabbro to granodiorite) that record co-mingling of magmas within the arc magma chamber (e.g., Ruby and Diamond Creek plutons). Cumulate-textured ultramafic rocks found in several places (mile 81, 83, 91, Salt Creek, Crystal, Granite Park, Diamond Creek), typically as tectonic slivers, are interpreted to be variably dismembered parts of the base of these arc plutons, rather than as ophiolite fragments. This interpretation is based upon the abundance of phlogopite and light rare earth elements in wehrlite successions, implying a geochemical contribution from subducting slab material. Compositional relations of the mile 83, 91, and Crystal ultramafic bodies are consistent with their origin as a single cumulate succession related by fractional crystallization (Seaman et al. 1997). The arc plutons are relatively rich in the feldspar minerals and hence were relatively strong during deformations. Thus, some of these plutons are not as obviously deformed (foliated) as the schists. However, they were all intruded before the period of intense 1.70- to 1.68-Ga deformation.

Syncollisional Granites

The 1.71- to 1.66-Ga granites and granitic pegmatites have a different composition, intrusive style, and deformational character than the arc plutons. They are

Table 2–1. Plutons and Dike Swarms

Name	*Mile*	*Age-Ma*	*Composition*	*Deformation*	*Interpretation*
Cottonwood Pegmatite Complex	77–82	1685 ± 1; 1680 ± 1	Biotite and biotite–muscovite granite and granitic pegmatite dikes and sills	Dated dikes cross-cut S2 and are weakly deformed	Late syn-D2
Grapevine Camp Pluton	**81.5**	**1737**	**Medium-grained foliated biotite granite**	**Contains S1; alligned along S2 Vishnu shear zone**	**Age and fabric suggests pre-D1 arc pluton, but composition similar to peraluminous plutons**
83-mile ultra-mafic	83	Undated	Layered ultramafic containing cumulate layers of olivine and pyroxene	Layering parallel to S1, truncated by S2-parallel margins	Tectonic slice of cumulate rocks from an arc pluton
Zoroaster Pluton	**84.6–85.5**	**1740 ± 2**	**Foliated medium-grained biotite granite to granodiorite orthogneiss**	**Contains S1, folded by F2**	**Pre- or syn-D1**
Cremation Pegmatite Complex	84–88	1698 ± 1, with older inherited monazite cores	Fine- to medium-grained biotite–muscovite granite and granite pegmatite	Cross-cuts S1, contains weak S2	Late syn-D2
Bright Angel Pluton	88 and Bright Angel Creek	Undated	Coarse-grained friable granite and pegmatitic granite	Intruded along Bright Angel shear zone; contains weak S2	Late syn-D2
Phantom Pluton	Phantom Canyon	1662 ± 1	Fine- to medium-grained biotite–muscovite granite	Cross-cuts S2, contains synmagmatic shear zones	Syn-D3
Pipe Creek Pluton	89	Pb–Pb dates of 1.74–1.69 Ga	Granite to granodiorite	Contains S1, folded by S2	Pre-D1
Horn Creek Pluton	Near 90.6	1713 ± 2	Foliated medium-grained hornblende quartz diorite to tonalite	Contains magmatic S1, solid-state S2	Syn-D1
91-mile ultramafics	**91**	**Undated**	**Cumulate layered ultramafic**	**Contains S1, boudinaged by D2**	**Pre-D2 tectonic sliver of arc pluton cumulate rock**
Trinity Pluton	**91.5**	**1730 ± 93**	**Medium- to coarse-grained biotite granodiorite to granite orthogneiss**	**Contains S1 and S2**	**Pre-D1**
Boucher Pluton	**96.2**	**1714 ± 1**	**Granodiorite to tonalite**	**Weakly foliated**	**Pre- or syn-D1**
Crystal Pluton	**97**	**Undated**	**Granite to granodiorite**	**Weakly foliated**	**Syn-D1**
Tuna Pluton	**99**	**1750–1710 Ga, with inherited zircons >2.0 Ga**	**Medium-grained foliated granodiorite**	**Contains S1 and S2**	**Pre- to syn-D1**
Sapphire Pegmatite Complex	99–104	Undated	Granite and granitic pegmatite dikes and sills	Cross-cut S1, contain S2	Syn-D2
Ruby Pluton	**102–108**	**1716 ± 0.5**	**Hornblende–biotite** granodiorite, diorite, and **gabbro**	**Contains S1 as** magmatic **layering**	**Syn-D1**
Garnet Pegmatite Complex	111–115	1697 ± 1	Biotite and biotite–muscovite granite and granite pegmatite dikes and sills	Cross-cuts S1, weak S2	Syn-D2

(continued)

TABLE 2–1. Plutons and Dike Swarms (*Continued*)

Name	*Mile*	*Age-Ma*	*Composition*	*Deformation*	*Interpretation*
Elves Chasm Pluton	**112–118**	**1840 ± 1**	**Lineated and foliated hornblende–biotite tonalite to quartz diorite and granodiorite, shows minging textures between intermediate and mafic units**	**Contains S1 and S2**	**Pre-D1, older arc basement**
Granite Park Mafic Complex	**207–209**	**Undated**	**Alternating layers of gabbro, anorthosite, and granodiorite with gabbroic pegmatite**	**Contains S1-related, top to SW, thrust-sense shear zones**	**Pre-D1 layered Mafic, may be base of Diamond Creek pluton**
Diamond Creek Pluton	**216–227**	**1736 ± 1**	**Granodiorite, tonalite, diorite and gabbro, shows mingling textures and cumulate ultramafic**	**Contains magmatic S1 and solid-state S2 shear zones**	**Syn-D1**
229-mile granite	228.7	Undated	Fine- to medium-grained biotite granite	Contains strong S2 foliation and S2-parallel psuedotachylite	Syn-D2
Travertine Falls Pluton and dike complex	230–231	1704 ± 1	Medium-grained biotite granite	Injects into and cross-cuts S1, contains variable S2	Syn-D2
232-mile Pluton and dike complex	231.7–232.4	Undated	Medium-grained biotite granite	Injects into and cross-cuts S1, contains variable S2	May be same as Travetine Falls Pluton, syn-D2
234-mile Pluton and dike complex	233.5–235	Undated	Medium-grained biotite granite	Cuts S1, contains variable S2	May be same as Travetine Falls Pluton, syn-D2
237-mile Pluton	236.7–237	Undated	Medium-grained biotite granite	Cuts S1, contains weak and magmatic S2	Syn-D2
Separation Pluton	239.3–239.8	Undated	Coarse-grained to megacrystic biotite, (+muscovite) granite	Contains strong S2	Syn-D2
245-mile pluton	**242–246**	**1720 ± 5**	**Granodiorite, tonalite, to diorite mingled Mafic Pluton intruded by Spencer Pluton**	**Contains S1 and S2**	**Pre-D2**
Spencer Pluton and dike complex	242–245 and in upper Spencer Canyon	Undated	Medium-grained biotite (muscovite) granite	Contains variable S2	Syn-D2
Surprise Pluton	246–252	Undated	Medium grained biotite + muscovite (garnet) granite	Contains weak S2	Syn-D2, same as separation Pluton
Quartermaster Pluton	Injects into Surprise Pluton from 250–261	1375	Coarse-grained granite	Cross-cuts S2	Post-D2

Data from Clark 1976; Babcock et al. 1979; Babcock 1990; Ilg et al. 1996; Hawkins et al. 1996; Hawkins 1996. **Bold = metaluminous arc plutons**; light = peraluminous granites; * = 1.4 Ga granite.

biotite–granite, biotite–muscovite (garnet) granite, and granitic pegmatite that probably formed by partial melting of the lower crust during deformation. Melts rose opportunistically along cracks and shear zones and froze to form dikes or coalesced as small stocks and plutons. It is common for a network of granite dikes to locally make up more than half of the exposed rock volume, blurring the distinction between dike swarms and plutons and indicating that these plutons coalesced from dikes. In the Upper Granite Gorge, intrusive complexes are made up of nearly equal proportions of medium-grained granite and granitic pegmatite. These are named the Cottonwood, Cremation, Sapphire, and Garnet Canyon complexes (Fig. 2.3). In the Lower Granite Gorge, intrusives are more massive and pluton-like (e.g., Travertine Falls, Separation, and Surprise plutons; Fig. 2.5) and commonly are composed of sheets concordant to foliation.

Granites are commonly stretched (boudinaged), folded, and foliated, but also occur as undeformed cross-cutting tabular dikes and orthogonal dike networks. These relationships suggest that they were intruded during regional deformation. Plutonism was also synchronous with peak metamorphism; the heat from the molten rock combined with high ambient temperatures at depth caused chemical reactions and growth of metamorphic minerals in the schists. Highest metamorphic temperatures are recorded in areas of most voluminous granites indicating that granites created zones that were >200°C hotter than surrounding areas (see below). Based on detailed studies, the range of ages of these dikes (1.70–1.68 Ga) is thus interpreted to also be the time of peak metamorphism and the main contractional deformation. Plutonism continued until 1.66 Ga (Phantom Pluton), and local deformation also continued. Cessation of magmatism, deformation, and metamorphism by ~1.65 Ga is interpreted to be the beginning of a long period (200 m.y.) of tectonic stability of this newly formed part of continental North America.

Post-Orogenic Granites

The youngest plutonic rock in the Grand Canyon is the 1.35-Ga Quartermaster pluton (S.A. Bowring, unpublished U/Pb zircon data) and related pegmatites (Fig. 2.5). These represent part of a regionally important period of intracratonic magmatism in the southwestern United States. The Quartermaster pluton is a coarse-grained (rarely megacrystic) friable granite that intrudes and is intermixed with older biotite granite of the Surprise pluton from mile 250–261. A pegmatite dike of similar age was dated in Diamond Creek, and several other subhorizontal dikes of pegmatite in the Lower Granite Gorge may also be this age. These plutons and dikes postdate the main contractional deformation as shown by cross-cutting relations and a lack of fabric. There are no 1.4- to 1.3-Ga plutonic rocks yet known from the Upper Granite Gorge.

DEFORMATIONAL HISTORY

Rocks of the Granite Gorge Metamorphic Suite were deposited at the earth's surface (1.75–1.74 Ga) and then buried to depths of 20–25 km and complexly squeezed by thrusting and folding (Fig. 2.2b). Metamorphic data suggest they were exhumed part way back to the surface (to 10-km depths) during the main deformation at 1.7–1.68 Ga. The tectonic fabrics that give rocks of the Granite Gorge Metamorphic Suite their distinctive vertically layered appearance were formed in the middle crust during a progression of tectonic events from 1.74 to

1.66 Ga. Deformations took place throughout this time interval whenever the stresses generated by arc collisions exceeded the strength of the crust. Thus, deformation was accentuated when rocks were hot and ductile—for example, in areas heated by magmas. The main deformational features are folds and foliations that record NW–SE squeezing and shear zones that record shearing along boundaries of tectonic domains. All of these structures formed when the rocks were solid, but had consistencies something like stiff Silly Putty. The final structures record deformation over a long time interval, variations in temperature and rock strengths from place to place and time to time, and changing stress fields involving combinations of squeezing and shearing. There is also evidence for an interaction of brittle and ductile deformation; for example, dike swarms filled a complex network of cracks that developed even as the rocks were deforming ductilely. The result is a very complex set of structures. We begin to understand the history of this deformation by recognizing a number of generations of structures that can be distinguished based on their style and overprinting relationships (hence relative age; Fig. 2.7).

The most obvious result of cumulative deformation is a profound verticality of tectonic layering below the Cambrian unconformity. This NE-striking vertical fabric is a second-generation composite (S2) fabric that records two effects: (1) the rotation and transposition of original bedding and an earlier tectonic layering (S1) to a subvertical orientation during development of tight folds (such as the Sockdologer and Zoroaster antiforms (Fig. 2.3) and (2) alignment of micas and other inequant grains perpendicular to tectonic shortening. This fabric formed synchronously with metamorphism and granite intrusion mainly during a 15-m.y. period from 1.7- to 1.685-Ga. Evidence for this is that melt pods that give U-Pb dates of 1.7–1.69 Ga are folded and boudinaged, then cross-cut by unfolded dikes of granite that give U-Pb dates of 1.685–1.68 Ga (Hawkins et al. 1996).

The main northeast-trending subvertical foliation and its associated folds are second-generation structures (S2 and F2, respectively) that fold an earlier tectonic layering (as well as bedding). This early S1 fabric can be seen in areas that escaped intense shortening (Sockdologer synform; Walthenburg area) and as aligned mineral trails preserved within the protective housing of metamorphic minerals such as garnet (Fig. 2.8). In these areas, the early (S1) foliation is commonly subparallel to bedding and has a northwest strike and shallow dip. However, in other areas (e.g., Phantom Creek) S1 is also subvertical, presumably due to its rotation during D2. The regional foliation pattern in both the Upper and Lower Granite Gorges (Figs. 2.3 and 2.5) involves domains of northwest-striking S1 that are folded and variably reoriented into the northeast-striking subvertical S2 orientation. We speculate that early (D1) tectonism involved thrusting and subhorizontal foliations, but the detailed configuration of early structures, such as direction of movement of thrusts, has not yet been worked out, mainly due to the intensity of the later (D2) shortening.

D3 structures are defined as ones that fold or deflect the main S2 subvertical layering. These include northwest-striking crenulations, late synplutonic mylonite zones (e.g., in the cores of 1.68-Ga dikes and in the 1.66 Phantom pluton), and low-temperature shear zones that reactivated older zones and may have allowed differential uplift of blocks (e.g., in the Gneiss Canyon shear zone). This group of structures represents several styles and periods of deformation, including both continued 1.68- to 1.65-Ga shortening and possible 1.4-Ga deformation. For the low-T (greenschist grade, <450°C) shearing in the 96-mile shear zone and in several discrete zones within the Gneiss Canyon shear zone (Fig. 2.5), kinematic indicators suggest west-side-up and dextral oblique shear.

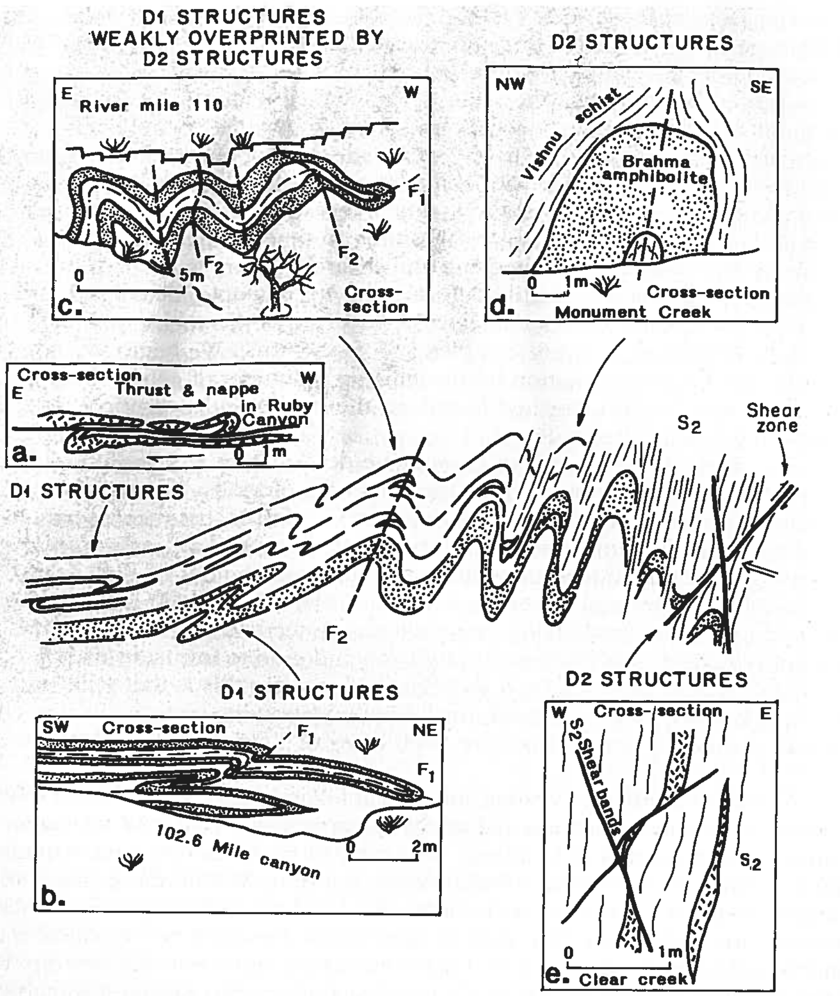

FIGURE 2.7. Overprinting of folds and fabrics and nomenclature for unraveling polyphase deformation. D1 involved thrusting, F1 recumbent folding, and development of S1 foliation; D2 involved NW–SE shortening, development of upright F2 folds, and dominantly west-side-up shear on shear zones; D3 involved minor crenulations and adjustments. Examples from outcrop photos: (a) and (b) recumbent F1 folds in Ruby Canyon and 102.6 Mile Canyon, respectively; (c) F1 recumbent fold refolded by F2 upright folds in area between Shinumo and Walthenberg canyons; (d) F2 fold with axial plane cleavage developed in metasedimentary rocks but not amphibolite, Monument Creek; (e) intense S2 and conjugate shear bands developed in Clear Creek Canyon. (Adapted from Ilg et al. 1996.)

TECTONIC BLOCKS

As is true throughout the southwestern United States, the Proterozoic crust in the Grand Canyon is segmented into blocks of different character by shear zones (Karlstrom and Williams 1998). Figures 2.3 and 2.5 show the main blocks and shear zones of the Grand Canyon transect; Table 2.2 summarizes tectonic evolution of blocks and shear zones. As discussed earlier, there is a general similarity of metasedimentary and metavolcanic rock types, similar types and ages of intrusive rocks, and similar styles and generations of deformational fabric across the transect. However, there are also important differences between blocks that might possibly indicate juxtaposition of once-separate tectonic terranes. The next sections discuss two ways that blocks differ: their isotopic compositions and metamorphic P–T histories.

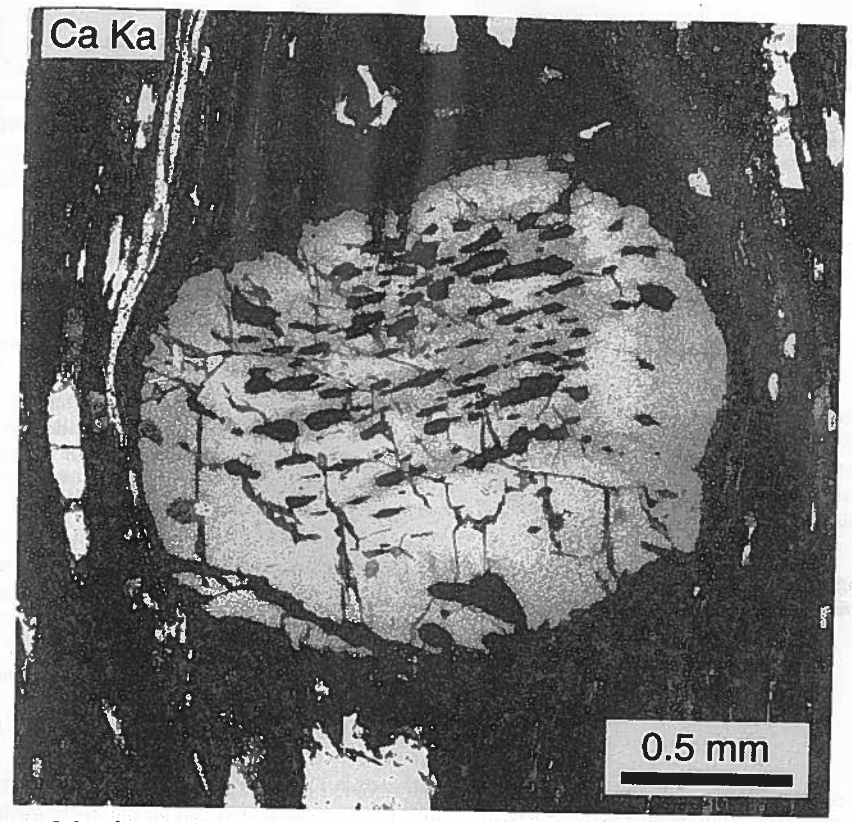

FIGURE 2.8. Electron microprobe, Ca Ka map showing compositional zoning in a single garnet crystal from mile 82. The thin-section view is approximately vertical and perpendicular to NE-striking S2 foliation. Note that the core of the garnet overgrew a slightly folded (S1) fabric. This early fabric was folded and transposed into the vertical S2 orientation now seen in the garnet rim and outside of the crystal. The increasing Ca content of the garnet, along with the composition of other associated minerals, suggests that the garnet core grew as temperature and pressure increased. Peak conditions of approximately 550°C and 6 kbar (20-km depths) were reached during the growth of the outer part the garnet, but the compositions of the extreme garnet rims indicate that pressures decreased during the final increments of growth. The composition of matrix minerals indicate that the final matrix recrystallization occurred at conditions near 550°C and 3 kbar (10-km depths). Thus, this one garnet crystal and its matrix record increasing pressures and the subsequent decompression from 20- to 10-km depths all during D2 deformation. The compositional map was collected on a Cameca SX-50 electron microprobe at the University of Massachusetts.

Table 2–2. Blocks and Shear Zones

Name	*Mile*	*Rocks and Structures*	*Metamorphism and Cooling History*
Mineral Canyon block	Mile 77–81	Migmatitic Rama–Brahma–Vishnu schists, intimately injected by the Cottonwood granite/pegmatite swarm, all folded into kilometer-scale moderately NE-plunging F2 Sockdologer antiform.	Upper amphibolite grade conditions of 600–720°C, 6 kbar were reached syn-D2 at 1.72–1.695 Ga; rapid cooling to 600–550°C by 1690 Ma; slower cooling to 550–500°C with continued fluid flux until 1660; cooling through 350°C at 1.4 Ga.
Vishnu shear zone	Mile 81	Narrow NE-striking subvertical fault zone with quartz veins marks juxtaposition of Grapevine Camp pluton abruptly against Vishnu Schist; kinematic history poorly known due to brittle overprint and likely multiple movements at high and low temperature.	Migmatites on the east are juxtaposed with amphibolite-facies rocks on west so zone marks a ~100°C temperature discontinuity; presence of sillimanite in schist near the fault may reflect heating during juxtaposition in D2 time.
Clear Creek block	Mile 81–88	Vishnu Schist with tectonic sliver of ultramafic rock at mile 83; Vishnu is structurally underlain by Brahma Schist then Zoroaster pluton in core of SW-plunging F2 Zoroaster antiform.	Except for higher-grade rocks against the Vishnu fault, peak metamorphic grade increases toward the west, from 500 to >600°C, all at 6 kbar. Volume of peraluminous granite also increases westward, and a 1698-Ma date on the granite is interpreted to be the time of peak metamorphism and D2 shortening, in agreement with U-Pb sphene dates of 1.7–1.695 Ga at mile 84. Late andalusite in Clear Creek Canyon (mile 84) and garnet zoning indicate decompression from 6 to 3 kbar during D2.
Bright Angel shear zone	Mile 88	2-km-wide zone of NE-striking subvertical S2 fabric that transposes NW-striking subvertical S1 fabric. The shear zone contains tectonically intermixed Rama, Brahma, and Vishnu schists, complexly injected and deformed peraluminous granite and pegmatite of the Cremation swarm, and the Bright Angel pluton that is elongated parallel to the zone.	Upper amphibolite grade, with ubiquitous sillimanite and migmatites showing evidence for partial melting of schist; parallel magmatic and solid-state fabrics suggests that movement was synchronous with granite emplacement.
Trinity block	Mile 88–96	Large refolded folds of Brahma and Vishnu schist, the Trinity orthogneiss, and an ultramafic sliver at mile 91.	Upper amphibolite grade, 650–725°C, 6 kbar; 1662 Phantom Creek Pluton is the youngest peraluminous granite in the Canyon. Metamorphic zircon, monazite, xenotime, and sphene give dates of 1660–1550 Ma with a systematic decrease in metamorphic age from east to west, apparently unrelated to the density of peraluminous dikes; the block may have been tilted 1.55–1.4 Ga; muscovite Ar–Ar dates of 1.4 Ga suggest cooling through 350°C at 1.4 Ga.
96-mile shear zone	Mile 96	300-m-wide NE-trending subvertical mylonite zone in Vishnu and Brahma schist and metadacite. Shear sense is west-side-up and dextral oblique slip.	Extreme grain size reduction and greenschist grade indicate low-T (~450°C) movement; 1.4-Ga muscovite Ar dates to east versus older dates to west suggest post-1.4-Ga movement.
Boucher block	Mile 96–98	Vishnu Schist intruded at the east and west by arc plutons; S2 dominates, but earlier fabrics evident in Boucher Canyon.	Greenschist to lower amphibolite amphibolite grade, $T \sim 500$°C, $P \sim 6$ kbar. U-Pb sphene dates yield the same age as the Boucher Pluton (1714 Ma), suggesting that the pluton cooled rapidly below ~600°C. Ar–Ar muscovite dates of >1.6 Ga indicate that rocks cooled through 350°C by 1.65 Ga and remained cold while rocks east of the 96-mile shear zone were at high grade from 1.66 to 1.55 Ga and >350°C until 1.4 Ga.
Crystal shear zone	Mile 98	1-km-wide, NE-striking subvertical zone of intense S2 fabric. Heterogeneous lithologies include various schists, granodiorite, ultramafics; anastomozing foliation and lenticuar boudins of diverse lithologies in chlorite schist form tectonic melange.	Movement sense and timing poorly constrained; presence of melange and abrupt change in Pb isotopes across the zone suggests it may be a paleosuture zone.
Ruby block	Mile 98–108	Vishnu Schist intruded by plutonic rocks of the Tuna Creek and Ruby plutons, all intruded at the east of the block by the Sapphire Pegmatite Complex. Moderately south-dipping S1 foliation folded into kilometer-scale F2 folds with S2 cleavage best developed in schist.	Metamorphic grade highest in area of pegmatite complex; about 600–700°C and 6–8 kbar. Metamorphic sphene from the ~1.72-Ga Tuna Pluton yields U-Pb dates of about 1.70 Ga and apatites from the pluton yield U-Pb dates of about 1.65 Ga. These data suggest that peak metamorphism occurred at about 1.70 Ga and the rocks cooled through about 500°C by 1.65 Ga, in agreement with >1600 Ma Ar–Ar dates on micas

(continued)

TABLE 2–2. Blocks and Shear Zones (*Continued*)

Name	*Mile*	*Rocks and Structures*	*Metamorphism and Cooling History*
Bass shear zone	Mile 107.8–108.2	0.5-km-wide ENE-striking subvertical zone within Vishnu Schist on west margin of Ruby Pluton; pluton itself is internally undeformed; anastomosing shear bands and foliation fish give dextral shear sense.	Shearing took place at high grade as shown by abundant sillimanite and K-spar and leucosomes that may be partial melt pods. Lower T shearing is indicated by chlorite and sericite alteration.
Walthenberg–Shinumo block	Mile 108–112 and in Shinumo Creek	Brahma and Rama schists overlain stratigraphically by Vishnu Schist near Walthenberg and in Shinumo Creek; kilometer-scale nappe-like F1 folds with moderately north-dipping axial planes are nearly E–W against Bass shear zone.	A 1697 U-Pb date on a syn-D2 pegmatite suggests D2 accompanied metamorphism (6–8 kbar, 550°C) at 1.7–1.69 Ga.
Contact Zone of Elves Pluton	Mile 112–113, Waltenberg Canyon area, in 115-mile Canyon, and Middle Granite Gorge	Concordant contact between the Elves Chasm Pluton and overlying Granite Gorge Metamorphic Suite is a 0.5-km-wide transitional contact marked by distinctive orthoamphibole-bearing gneisses and evidence for in situ partial melting. Contact is folded into large-scale F2 folds with a generally northwest-striking fold envelope. The contact may be a paleosol indicating deposition of Vishnu Schist on older basement, or a high-temperature shear zone, or both.	Peak metamorphic grade was approximately 600°C, 8 kbar, but garnet shows evidence for polyphase growth. Metamorphic monazite from pelitic schist in the contact zone yield U-Pb dates of 1700 Ma, which are interpreted as the time of peak metamorphism. Garnet with coronas of granitic leucosome suggest in situ partial melting of the Elves Chasm contact zone, but melting is interpreted to record channelization of heat and fluids in the contact zone at 1.7 Ga rather than contact metamorphism at 1.84 Ga. Kyanite is abundant, as is gedrite, with late-stage growth of sillimanite and cordierite indicating syn-S2 decompression to midcrustal levels.
Elves Chasm block	Mile 113–127	1.84-Ga Elves Chasm Pluton contains mafic and intermediate plutonic units, including tabular amphibolite bodies that are interpreted to be dikes; all units are strongly foliated and lineated by S1 and L1; voluminous dikes of the Garnet Canyon Pegmatite Complex cross-cut S1; contact zone is exposed near Blacktail Canyon.	Metamorphism was high grade as suggested by migmatitic gneisses in pluton margins; U-Pb data from sphene within the pluton suggest that rocks cooled through about 600°C 1675–1655 Ma.
Middle Granite Gorge block	Mile 127–139	Massive amphibolites are overlain by felsic volcanics, then turbidites near Bedrock rapid; section probably correlates with Walthenburg block, numerous granite dikes intrude metasedimentary schists.	Metamorphism amphibolite grade.
Granite Park block	Mile 209	Cumulate layered gabbro, anorthosite, amphibolite, and gabbroic pegmatite; rocks are strongly foliated and contain thrust sense top-to-west shear zones.	
Diamond Creek block	East of Hurricane fault, in Diamond Creek	Thick sections of intermediate volcanic schists interlayered with amphibolite (with local pillows); east side of the block cross-cut by peraluminous granite dikes; intensely developed NE-striking subvertical foliation may mark a ductile high strain zone parallel to the Hurricane fault.	Metamorphic grade is lower amphibolite, 500°C, 3 kbar.
Travertine block	Mile 212–234	Diamond Creek Pluton exposed from mile 212–228.5, mingled magma textures in area 227–228. Travertine Canyon area contains Vishnu Schist overlain on the west by pillow basalts, then intermediate volcanics. Area from 230–235 is Travertine Falls, 232-mile and 234-mile plutons, and related dikes and sills intruding metavolcanic rocks.	Metamorphic grade at Travertine Canyon is 550°C, 3 kbar; grade increases toward the plutons in the western part of the block; sphene in the Travertine area gives a U-Pb date of 1698 ± 7 Ma; hornblende in shear zones in the Diamond Creek Pluton gives an Ar age of 1.67 Ga. The former gives the time of peak metamorphism; the latter gives the time of cooling through about 500°C.

(continued)

TABLE 2–2. Blocks and Shear Zones (*Continued*)

Name	*Mile*	*Rocks and Structures*	*Metamorphism and Cooling History*
Gneiss Canyon shear zone	Mile 234–242.2	Migmatitic gneisses of mixed metasedimentary a metavolcanic protolith intimately injected with granites. Carbonate at mile 236.5 may mark important tectonic boundary. Structures consis of NE-striking subvertical S2 foliation zones separating domains of NW-striking folded S1 fabric; shear sense in the shear zones is NW-si up and dextral oblique slip. Discrete retrogra mylonites occur at mile 234, 235, 236, 239.5, and 242.2 and show dextral, west-side-up she 1.1-Ga diabase dikes are restricted to this zon	Metamorphism is upper amphibolite facies, $T = 650°C$, $P = 6$ kbar, with evidence for partial melting; monazite in syntectonic granite are 1701 Ma; monazites in late tectonic cross-cutting dikes are 1678 Ma.
Spencer Canyon block	Mile 242.2–247	Spencer Canyon metasedimentary and metavolca gneisses are intruded by the 246-mile mafic a pluton (1720 Ma) and by granite dikes of the Spencer pluton.	Metamorphic grade is upper amphibolite with evidence for partial melting, 600°C, 4–6 kbar in Spencer Canyon.
Surprise–ton; Quartermaster block	Mile 247–261	From mile 246–252, rock is mostly Surprise Plu- from 252–261, 1375-Ga Quartermaster pluton intrudes the Surprise Pluton. Foliation is mainl weak magmatic layering that is NW striking a moderately dipping.	Metamorphism of metasedimentary screens reaches partial melting within the granites.

Data from Ilg et al. 1996; Hawkins et al. 1996; Williams, unpublished data; Hawkins, 1996; Hawkins and Bowring, in press; Hawkins and Bowring, unpublished data.

ISOTOPIC BOUNDARIES

Although rocks are broadly similar across the transect, different isotopic "signatures" suggest that blocks were derived from crustal domains of different age and composition. On a regional scale, Bennett and DePaolo (1987) and Wooden and Dewitt (1991) showed that rocks of the Mojave province of northwestern Arizona and California had isotopic signatures that indicated that there was a larger percentage of older crustal material at depth than beneath the Yavapai province of central Arizona. The nature of older material is not well known but probably includes Archean basement fragments at depth, more voluminous 1.84-Ga arc basement like the Elves Chasm gneiss at depth, or an increase in Archean detrital zircon grains deposited in the schists.

In the Grand Canyon transect, there are two shear zone boundaries that mark changes in isotopic character of the crust. Rocks in the eastern Grand Canyon (east of the Crystal shear zone—mile 98) have Pb isotopes similar to the Yavapai province of central Arizona that suggest that arc rocks were derived from the mantle at ~1.75 Ga without contamination from older crustal rocks. In contrast, rocks from mile 98–115 show elevated Pb signatures indicative of older material similar to the Mojave province (Hawkins et al. 1996). This may mean that the two crusts are fundamentally different composition terranes and were juxtaposed across the Crystal shear zone. Alternatively, the 1.75- to 1.73-Ga Granite Gorge Metamorphic Suite and related arc plutons may have developed across a deep crustal transition from oceanic crust to 1.84-Ga Elves Chasm arc crust, just as the present-day Aleutian volcanic arc climbs from oceanic basement to older Alaskan crustal basement. The Crystal shear zone is narrow (500 m) but contains highly tectonized rock called melange (Table 2.2) that could mark the site where a whole ocean basin was subducted before the different sides collided and were sutured. Alternatively, this and other shear zones may mark simply a zone of relatively minor (kilometer-scale) slip that developed as crust was squeezed and thickened.

A similar, more subtle, isotopic transition occurs across the Gneiss Canyon shear zone in the Lower Granite Gorge. Here, Pb 207/204 values also increase to the west (J.L. Wooden, unpublished data). This shear zone is 10 km wide and contains a tectonic sliver of carbonate at mile 236.5 (Fig. 2.5), the only such outcrop of this rock type in the transect and hence an exotic lithology that may mark an important boundary. Like the Crystal shear zone, isotopic data show a similar though less dramatic increase in older (Mojave) crust on the west side. Structural and metamorphic studies suggest that the Gneiss Canyon shear zone is a thrust that brings deeper rocks up on the west (see below). If the Aleutian island analogue is valid, "Mojave province" may consist of older arc and Archean fragments at depth, on which the 1.74–1.71 arcs were built. The Yavapai arcs of the same age were built on oceanic crust, with a wide transitional boundary between the provinces, but an increase in the Mojave signature on the west side of major thrusts that mark the transition. The isotopic data do not prove that there are tectonic sutures at either shear zone, but they do indicate a change in bulk composition of the crust across the shear zones and suggest that these two boundaries are among the best candidates for suture zones.

METAMORPHISM

The Granite Gorge Metamorphic Suite contains a variety of metamorphic minerals that can be used to estimate the pressure (P) and temperature (T) history that rocks underwent. In particular, pelitic layers in the Vishnu Schist contain various combinations of garnet, sillimanite, staurolite, chloritoid, cordierite, kyanite, and andalusite (rare). Different combinations (assemblages) of minerals are indicative of specific P–T conditions. Similarly, garnet, biotite, and plagioclase change composition depending on the P–T conditions of their growth, so we can estimate P–T conditions by making compositional maps and calculating P and T (Fig. 2.8). Available P–T estimates for the different blocks are summarized in Table 2.2.

One remarkable finding is that rocks of the Upper Granite Gorge all record metamorphic pressures equivalent to depths of metamorphism of about 20 km; slightly greater depths are indicated between miles 98 and 115. This uniformity suggests either (1) that there are no structures (faults) that brought very deep rocks (e.g., 30–40 km) against shallow rocks, as one sees in continent–continent collision zones, or (2) that the metamorphic minerals that locked in the pressures grew after suturing such that any higher-pressure minerals were not preserved. Like the late intense contractional deformation, the peak 1.7- to 1.68-Ga metamorphism could have erased much of the evidence for earlier suturing.

In contrast, the Lower Granite Gorge records juxtaposition of different crustal levels. Rocks west of the Gneiss Canyon shear zone (Spencer Canyon block) record pressures that correspond to 15- to 20-km depths. These were thrust east onto rocks that were never deeper than about 10 km (Travertine block). This variation in exposed crustal depths, along with the large areas of uniform depth in the Upper Granite Gorge, allows us to make some generalizations about variations in the character of deformation and metamorphism with depth (Karlstrom and Williams 1998), but does not resolve our uncertainty about which boundaries may be suture zones.

Metamorphic temperature estimates tell another surprising story. Whereas depths remain relatively constant in the Upper Granite Gorge, minerals indicate that metamorphic temperatures varied from a low of 500°C (Boucher, Diamond Creek, and Travertine Canyons) to nearly 700°C (mile 78, near Horn Creek, 113–115 mile and Spencer Canyon). The hottest areas are generally associated with swarms of granite dikes and sills (Cottonwood, Cremation, and Garnet complexes), documenting the interaction of plutonism and metamorphism. These lateral temperature gradients of >200°C take place either abruptly across shear zones (e.g., 96-mile shear zone) or within distances of 5–10 km, as dike and granite complexes are approached (e.g., Clear Creek block—Table 2.2). We conclude that the temperature gradients were caused by additions of heat by the molten rock traveling through the dike complexes (Ilg et al. 1996). Similar temperature variability at 10 km in central Arizona (Karlstrom and Williams 1995) seem to suggest that the temperature distribution was laterally heterogeneous throughout the 10- to 20-km depths during 1.7- to 1.68-Ga tectonism (Karlstrom and Williams 1998).

In addition to the marked variation in temperature from place to place, metamorphic P–T data can also be used to document changes in the pressure and temperature of an individual rock through time. Figure 2.8 shows a compositionally zoned garnet mineral. These compositional changes, when compared to changes in plagioclase and biotite minerals in the same rock, suggest that this mineral decompressed from 20- to 10-km depths during its growth. Textures can document the relative timing of metamorphism and deformation. In Figure 2.8, the textures suggest that the core overgrew S1 and that the core and rim grew synchronously with progressive intensification of the S2 foliation seen in the matrix. Thus, we infer that this rock (and in fact most of the Upper Granite Gorge) passed from 20- to 10-km depths during the main 1.7- to 1.685-Ga deformational and metamorphic event. Similarly, the Spencer Canyon block of the Lower Granite Gorge was uplifted and thrust eastward along the Gneiss Canyon shear zone and stabilized at depths of about 10 km.

U-Pb dating of the metamorphic minerals monazite, xenotime, titanite, and apatite provides constraints on the thermal history of rocks during and immediately following the peak of metamorphism (Hawkins et al. 1996; Hawkins 1996; D.P. Hawkins and S.A. Bowring, in press; D.P. Hawkins and S.A. Bowring, unpublished data). The thermal history of the Mineral Canyon block (Fig. 2.4) is a model for most of the blocks in the Upper Granite Gorge. Vishnu Schist began to melt and was intruded by peraluminous granite melt at about 1710 Ma and continued to melt throughout the peak of metamorphism until about 1690 Ma. During this 20-million-year time interval, rocks decompressed from about 6 kbars to about 3 kbars and were rapidly cooled (at a rate of 15–30°C/ m.y.) to about 550°C. Then they cooled more slowly (at a rate of <2°C/ m.y.) for several hundred million years as the rocks resided at about 10- to 12-km depth.

The only block that does not seem to cool in this manner is the Trinity block, where metamorphic mineral growth took place between 1665 Ma and 1550 Ma (D.P. Hawkins and S.A. Bowring, in press). These ages become systematically younger from east to west for all of the dated minerals, and the minerals preserve little evidence for cooling. Perhaps this block represents a west-tilted crustal section for which the western end remained deeper and at higher temperatures and pressures than both the eastern end of the same block and the adjacent tectonic block to the west. This age pattern of the Trinity block emphasizes the fundamental importance of the block architecture for Proterozoic crust in the Grand Canyon. It also reveals how complicated the history of high-grade terrains can be and how difficult it is to correlate structures, fabrics, and metamorphic mineral reactions across such a terrain.

THE GREAT UNCONFORMITY

By 1.65 Ga, rocks throughout the Grand Canyon transect were at depths of about 10 km. For the Upper Granite Gorge, about 10 km of rock had been eroded during the 1.7- to 1.65-Ga deformation, reduced to particles, and transported and deposited elsewhere. At 1.35 Ga, rocks were probably still residing at near 10-km depths when they were intruded by granites like the Quartermaster granite. Thus, an additional 10 km of rock was eroded off the region between 1.35 and about 1.25 Ga to create a broad plain that was ready to receive sediments of the Grand Canyon Supergroup. The history of this unroofing of the middle crustal rocks is part of the development of the Great Unconformity (Powell 1876). Details are difficult to unravel because erosion has removed most of the record, and we do not know where the eroded sediment was deposited. Nevertheless, we are beginning to unravel parts of this history by examining the cooling history of the middle crustal rocks as they made their way toward the surface, as well as the sedimentary record of the Grand Canyon Supergroup (Chapter 5).

Ar isotope studies can date the last time a mineral underwent diffusion of ^{40}Ar. Diffusion ceases below about 500°C for hornblende and 300°C for muscovite. Data from across the transect show that micas become closed to diffusion at different times. Rocks between mile 96 and 115 cooled relatively quickly from 700°C at 1.7 Ga to below 300°C by 1.6 Ga, then stayed below 300°C. In contrast, rocks east of the Crystal shear zone did not cool through 300°C until approximately 1.4 Ga (Matt Heizler, unpublished data). This suggests differential cooling of the middle crust and continued importance of older shear zone boundaries during differential uplift at 1.4 Ga (Karlstrom and Humphreys 1998).

SYNTHESIS

Paleoproterozoic rocks of the Grand Canyon provide a rich laboratory for understanding the formation, stabilization, and reactivation of continental lithosphere. Crust formation took place in several steps: (1) differentiation of arc plutons at 1.84 and (mainly) at 1.75–1.71 above subduction zones, (2) deposition of 1.75- to 1.73-Ga volcanic and sedimentary rocks of the Granite Gorge Metamorphic Suite on the flanks of developing island arc volcanoes, and (3) collision of arcs with each other (1.74–1.71 Ga) and with the North American protocontinent at 1.7–1.68 Ga and resulting ductile deformation, melting of the lower crust and emplacement of syncollisional granites, and middle crustal metamorphism at depths of 10–20 km.

Stabilization of the new continental crust took place during and after the collisional orogeny as crust thickened and cooled. From 1.6 to 1.4 Ga, this part of North America remained little affected by tectonism, although Grand Canyon rocks remained in the middle crust and cooled fairly slowly. At about 1.4 Ga, renewed magmatism occurred; this may have driven differential uplift of blocks, erosion of about 10 km of crust, and development of a broad erosional surface that exposed middle crustal rocks and was to become the Great Unconformity. The northeast-trending basement shear zones and fabric of the Granite Gorge Metamorphic Suite were repeatedly reactivated in later Precambrian and Phanerozoic times to form dominant structures such as the Bright Angel and Hurricane faults.

ACKNOWLEDGMENTS

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• 3 •

GRAND CANYON SUPERGROUP: UNKAR GROUP

J. D. Hendricks and G. M. Stevenson

INTRODUCTION

In Arizona's Grand Canyon, a series of gently tilted sedimentary and igneous rocks are exposed in isolated outcrops along the Colorado River and its tributary canyons. The rocks overlie the schists and granites of the inner gorge and occur below the flat-lying Paleozoic sedimentary units. These wedges of rock are quite noticeable both from the rim and from the river because of the angular difference of the beds (compared to those above and below) and the striking color and topographic variations within the sequence. Major outcrops of these rocks occur in seven separate locations within the Grand Canyon National Park (Fig. 3.1).

NOMENCLATURE

John Wesley Powell was the first person to note the geology of these rocks during his historic traverses of the Grand Canyon in 1869 and 1872. In his reports, he described their stratigraphic position and ascribed a tentative Silurian age to the sequence.

Charles D. Walcott conducted extensive field studies in the eastern Grand Canyon in 1882–1883 and reported his findings some years later. In his 1894 report, Walcott divided this sequence of rocks into two terranes: the Chuar (upper) and Unkar (lower). The combined sequence was named the Grand Canyon Series. In the same report, Walcott provided the first geologic map of the eastern Grand Canyon and the measured stratigraphic thickness of the Grand Canyon Series. Walcott (1894) reported the series to be approximately 12,000 feet (3660 m) in thickness, with the Unkar terrane being 6800 feet (2073 m) and the Chuar terrane being 5200 feet (1587 m). A Precambrian age was assigned to the Grand Canyon Series by comparison with the "Keweenawan Series" of the north-central United States.

From studies conducted in the Shinumo area (Fig. 3.1), Noble (1914) subdivided the Unkar terrane into five formations and assigned group status to the Chuar and Unkar. The names applied to the formations of the Unkar Group, in ascending order, were (1) Hotauta Conglomerate, (2) Bass Limestone, (3) Hakatai

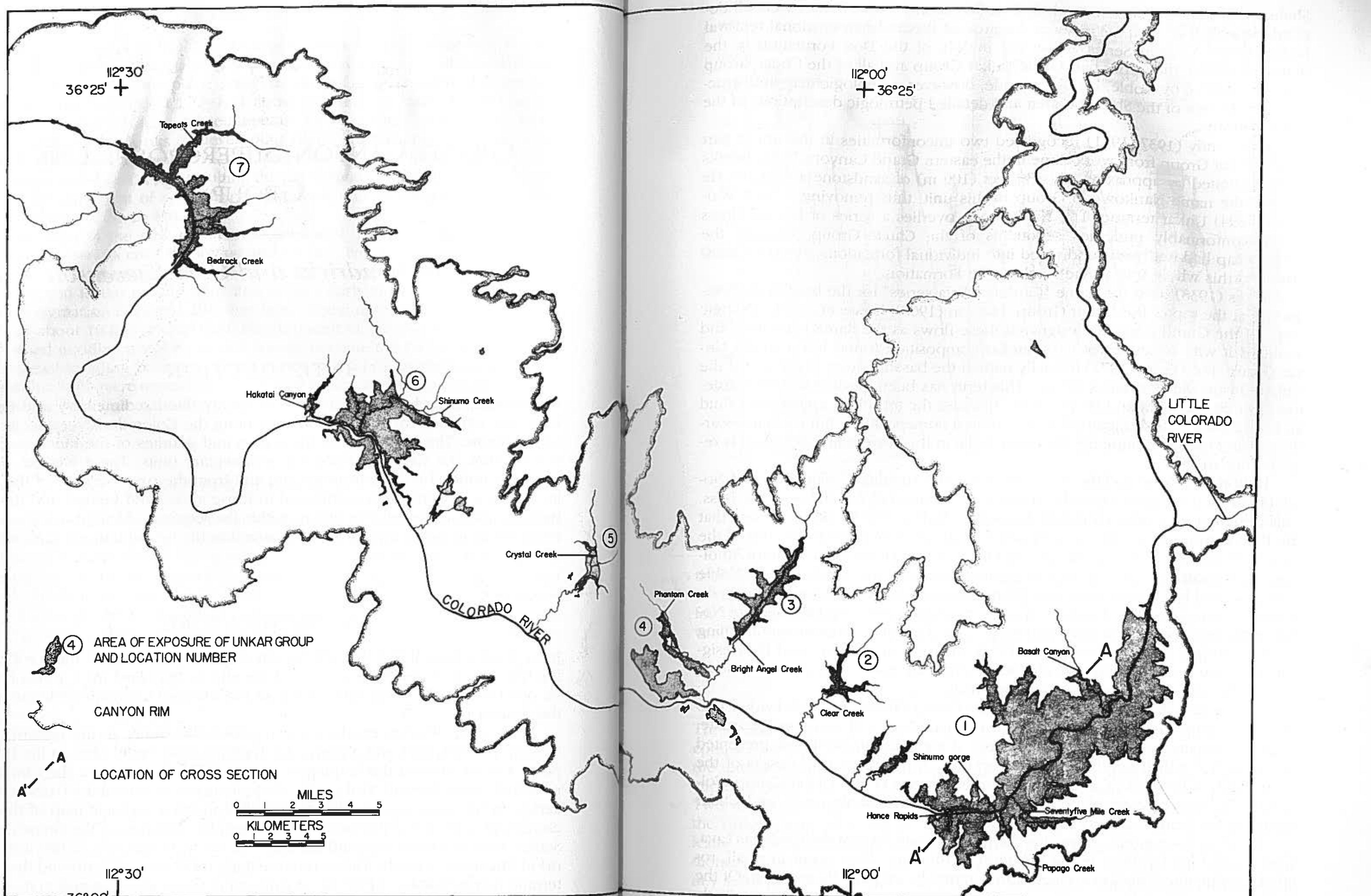

Figure 3.1. Location of exposures of the Unkar Group. (1) "Big Bend" region of the eastern Grand Canyon; (2) Clear Creek; (3) Bright Angel Creek; (4) Phantom Creek–Phantom Ranch; (5) Crystal Creek; (6) Shinumo Creek; (7) Tapeats Creek.

Shale, (4) Shinumo Quartzite, and (5) Dox Sandstone. These names were all derived from local geographic features. Because of Precambrian erosional removal of the Grand Canyon Series above the middle of the Dox Formation in the Shinumo region, the upper part of the Unkar Group and all of the Chuar Group were not named by Noble. He did provide, however, a geologic map and structural description of the Shinumo area and detailed petrologic descriptions of the Unkar Group.

Van Gundy (1937, 1951) recognized two unconformities in the upper part of the Unkar Group from work done in the eastern Grand Canyon. These breaks were separated by approximately 330 feet (100 m) of sandstone and shale. He applied the name Nankoweap Group to this unit, thus removing it from Walcott's (1894) Unkar terrane. The Nankoweap overlies a series of basaltic flows and unconformably underlies sediments of the Chuar Group. Because the Nankoweap had not been subdivided into individual formations, Maxson (1968) classified this whole unit as the Nankoweap Formation.

Keyes (1938) used the name "Cardenas lava series" for the basaltic flow sequence at the top of the Unkar Group. Maxson (1968), however, in his geologic maps of the Grand Canyon, designated these flows as the Rama Formation and included it with intrusive rocks of similar composition found lower in the Unkar Group. Ford et al. (1972) formally named the basaltic flows at the top of the Unkar Group the "Cardenas Lavas." This term has been modified to the Cardenas Lava by Lucchitta and Beus (1987). Because the term lava applies to a fluid and not a rock, this designation is not favored nomenclature, but the name Cardenas Lava is used commonly for these rocks in the current literature and is retained herein.

Dalton (1972) studied the Bass Limestone and Hotauta Conglomerate of Noble (1914) and suggested that the Hotauta be included as a member of the Bass. This designation is adopted in this discussion. Dalton (1972) also suggested that the Bass Limestone should be reclassified as the Bass Formation because of the variety of rock types within the unit and the fact that limestone is a minor lithology. Stevenson and Beus (1982) suggested that the Dox Sandstone of Noble (1914) should be renamed the Dox Formation also because of the lithologic diversity. For the purpose of continuity of nomenclature, the original names of Noble, with the exception of the re-ranking of the Hotauta Conglomerate, naming of the Cardenas Lava by Ford et al. (1972) and Lucchitta (1984), and the designation of the Dox Sandstone as the Dox Formation, will be used in the discussion of the individual formations of the Unkar.

As a result of these studies, the Unkar Group currently is subdivided into five formations: (1) Bass Limestone, (2) Hakatai Shale, (3) Shinumo Quartzite, (4) Dox Formation, and (5) Cardenas Lava. A stratigraphic section is presented as Figure 3.2, which depicts this nomenclature, the average thicknesses of the formations and members, and the general lithologies of the Unkar Group. Following the current code of nomenclature, the Unkar, Nankoweap, and Chuar comprise the Grand Canyon Supergroup.

All of the formations comprising the Unkar Group, except the Cardenas Lava, were named for localities in the Shinumo quadrangle. They occur in small, rotated, downfaulted blocks or slivers and commonly are partially exposed. Of the five recognized formations comprising the Unkar Group, only the Hakatai Shale is well-represented in the Shinumo Creek vicinity without further consideration for an "alternate" area. However, in the case of the three other sedimentary units, sections should be selected where they are best preserved. In the case of the Bass Limestone, a much more complete marine section is present to the west at Tapeats Creek (Fig. 3.1). Both the Shinumo and Dox formations have good sec-

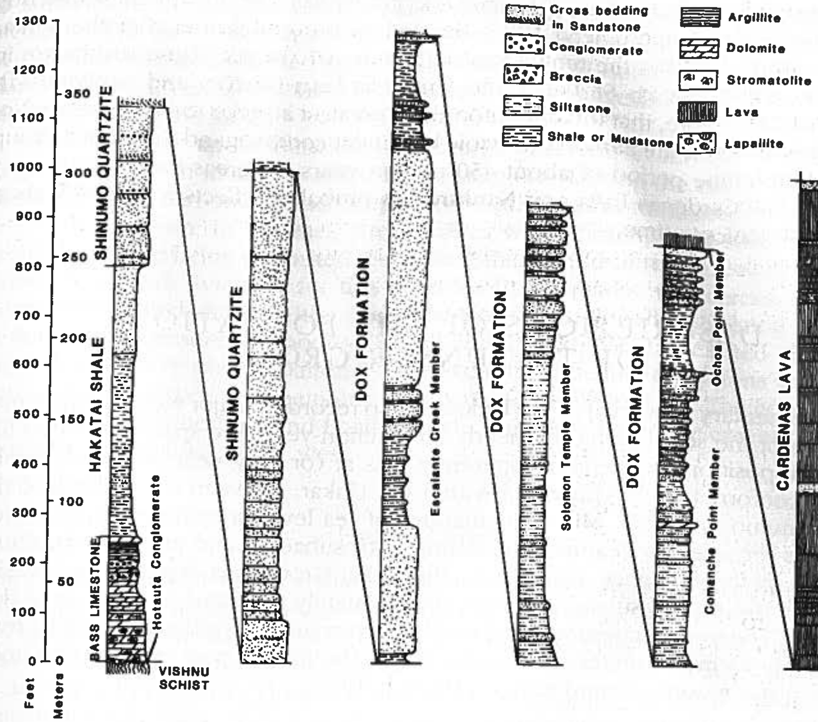

FIGURE 3.2. Columnar section of the Unkar Group.

tions eastward of Shinumo quadrangle in the Vishnu quadrangle where complete stratigraphic intervals are preserved and well-exposed. Perhaps the best locality for the Shinumo is in "Shinumo Gorge" near Hance Rapids, and the most revealing look at the Dox can be found in the "Big Bend" area (area 1 of Fig. 3.1).

AGE OF THE UNKAR

The Grand Canyon Supergroup overlies the metamorphic and granitic basement complex of Early Proterozoic age (1700 million years ago [m.y.a.] and underlies the middle Cambrian Tapeats Sandstone (550 m.y.a.). Thus, the supergroup occupies a portion of the 1150-million-year interval between 550 and 1700 m.y.a. Two methods of age determination have been applied to establish the time of the Unkar Group's formation. The Cardenas Lava provides the only stratigraphically controlled lithology of the supergroup, discovered to date, suitable for radiometric age determinations. McKee and Noble (1974) obtained an age of 1100 m.y.a. for the Cardenas Lava using the Rb–Sr method. The K–Ar method has produced Cardenas ages that are considerably younger than 1100 m.y.a. Ford et al. (1972) presented a single K–Ar age of 845 ± 20 m.y.a. for samples of the Cardenas Lava, and McKee and Noble (1974) obtained ages of 810 ± 20, 790 ± 20, and 781 ± 20 m.y.a. (K–Ar) for samples of the Cardenas Lava. The lower ages obtained using the K–Ar method may reflect an episode of heating and resetting of the K–Ar clock about 800 million years ago (McKee and Noble 1974).

Extensive study of the paleomagnetic pole positions and polar wandering paths by Elston (summarized 1986) has led to the conclusion that the Unkar Group accumulated in the time interval 1250 to 1070 m.y.a. These results are in agreement with the Rb–Sr age of the Cardenas Lava (McKee and Noble 1974). It appears, therefore, that the unconformity ("greatest angular unconformity"; Noble 1914) between the Early Proterozoic basement complex and the Unkar Group represents a time period of about 450 million years, whereas the unconformity between the Cardenas Lava and Nankoweap probably reflects a relatively short period of geologic time.

DESCRIPTIONS OF THE FORMATIONS OF THE UNKAR GROUP

The sedimentary sequence of the Unkar Group records a major west-to-east transgression of the sea. During the nearly 250-million-year time span postulated for Unkar deposition, the region apparently was at (or very near) sea level. Only one unconformity is documented within the Unkar: between the Hakatai Shale and Shinumo Quartzite. Minor fluctuations of sea level or sediment surface elevation is recorded by features suggesting both subaerial and marine deposition throughout the sequence. Apparently, the Unkar Group was deposited in a basin in which the rate of subsidence was approximately the same as the rate of deposition. The only suggestion of relatively deep-water deposition is noted by textural features in dolomites and mudstones in the middle part of the Bass Limestone in the western Grand Canyon (Dalton 1972).

The Bass Limestone and Hotauta Conglomerate Member

The Vishnu surface, over which the Unkar sea advanced from the west, was smooth with a local relief of probably no more than 150 feet (45 m). The Hotauta Conglomerate Member, the lowermost unit of the Bass Limestone, was deposited in low areas of the Vishnu terrane. This conglomerate consists of rounded, gravel-sized clasts of chert, granite, quartz, plagioclase crystals, and micropegmatites in a quartz sand matrix. It is found in the eastern Grand Canyon. In the Unkar exposures of the western Grand Canyon, the lowermost part of the Bass contains intraformational breccias and small pebble conglomerates, suggesting that the source of these clasts was toward the east.

The lithology of the Bass Limestone is predominantly dolomite with subordinate amounts of arkose and sandy dolomite with intercalated shale and argillite. Intraformational breccias and conglomerates also are found throughout the sequence (Dalton 1972). One feature of note within the Bass is the presence of biscuit-form and biohermal stromatolite beds (Nitecki 1971). The thickness of the Bass Limestone shows a general increase to the northwest ranging from 330 feet (100 m) at Phantom Creek (Fig. 3.1) to 187 feet (57 m) at Crystal Creek. The anomalously thin section at Crystal Creek probably reflects the presence of a Vishnu topographic high in this area during deposition. The Bass generally forms cliffs or stair-stepped cliffs: The more resistant dolomites make the risers, and the shale and argillite form steep treads.

Sedimentary features common to all exposures of the Bass Limestone include symmetrical ripple marks, desiccation cracks, interformational breccias/conglomerates, and both normal and reversed small-scale, graded beds (associated with stromatolites). All of these features suggest a relatively low-energy intertidal to supratidal environment of deposition. Although no evaporites presently are recognized in the Bass, some of the interformational breccias may be the result of collapse of earlier-formed gypsum. Dalton (1972) noted monoclinic crystal clasts in chert layers of the Bass Limestone. This is suggestive of a dolomitic replacement of gypsum.

The lithology and sedimentary structures observed in the Bass Limestone suggest deposition in an easterly transgressing sea. During the maximum incursion of the sea, carbonates and deep-water mudstones accumulated in the western Grand Canyon. In the east, stromatolites were forming, and shallow-water mudstones were being deposited. Following this period of transgression, the sea slowly regressed. Evidence for this includes ripple marks, mudcracks, and deposits of oxidized shales in the upper part of the Bass—all suggesting periods of subaerial exposure. Evaporite-forming conditions probably existed also during this regressive phase (Dalton 1972). Eventually, deltaic conditions predominated, which marked the beginning of Hakatai Shale deposition. The contact between the Bass Limestone and Hakatai Shale is gradational in the east and sharp, though conformable, in the west.

Hakatai Shale

The Hakatai Shale probably is the most colorful formation in the Grand Canyon, with colors that vary from purple to red to brilliant orange on outcrop. The colors result from the oxidation state of the iron-bearing minerals in the formation. The Hakatai is subdivided into three informal members based on lithology (Beus et al. 1974; Reed 1974. The lower two units are highly fractured argillaceous mudstones and shales that weather to gentle-to-moderate, granular slopes. The upper unit consists of cliff-forming beds of medium-grained quartz sandstone (Fig. 3.2). The Hakatai varies in thickness from about 445 feet (135 m) at Hance Rapids to nearly 985 feet (300 m) at the type section in Hakatai Canyon in the Shinumo Creek area.

Sedimentary structures, such as mudcracks, ripple marks, and tabular-planar cross bedding, suggest that the Hakatai was deposited in a marginal marine environment. Mudstones and shales of the lower two members probably were deposited in a low-energy, mud-flat environment. The upper sandstones suggest a higher-energy, shallow-marine environment (Reed 1974).

Although the contact between the Hakatai Shale and Bass Limestone is gradational in the eastern Grand Canyon and sharp and easily located in the western exposures, the contact between the Hakatai and the overlying Shinumo Quartzite is evident in all exposures. It is marked by an unconformity that truncates cross beds and channel deposits of the Hakatai. From observations made in the canyon of Bright Angel Creek, Sears (1973) indicates that Hakatai deposition in the area ended in conjunction with tectonic activity along a series of northwest-trending, high-angle, reverse faults.

Shinumo Quartzite

In contrast to the slope-forming argillaceous beds of the Hakatai Shale below and the Dox Formation above, the Shinumo Quartzite is a series of massive, cliff-forming sandstones and quartzites. The color of the Shinumo ranges from muted red, brown, and purple to white.

Four or possibly five poorly defined members have been recognized within the Shinumo Quartzite (Elston 1986; Daneker 1974). The lower units, in ascending order, consist of conglomeratic subarkose and submature quartz sandstone; to mature quartz sandstone; to brown quartz sandstone with abundant cross beds, clay galls, and mudcracks.

Near Shinumo Creek, the uppermost unit is the thickest member. It consists of fine-grained, well-sorted, and rounded quartz grains in a siliceous cement. Beds in the upper part of the upper member are contorted by fluid evulsion, which suggests that there might have been tectonic activity during this period.

The thickness of the Shinumo Quartzite shows a general increase to the west and ranges from 1132 feet (345 m) at Papago Creek in the east to 1328 feet (405 m) (Noble 1914) in Shinumo Creek. Because the Shinumo is such a resistant unit, it formed hills where exposed during the pre-Tapeats erosional event.

This feature, the pinching out of the Tapeats Sandstone against Shinumo Quartzite highs, can be seen today in exposures near the bottom of the Grand Canyon. Analysis of the lithology and the sedimentary structures of the Shinumo suggest that the environment of deposition was near-shore, very shallow, marginal marine, and part fluvial, part deltaic (Daneker 1974). The contact between the Shinumo and the overlying Dox Formation appears to be conformable in most locations and is marked by the lowermost shaley interval of Dox lithology.

Dox Formation

The Dox Formation is the thickest unit of the Unkar Group. The only complete section presently exposed, however, is in the eastern Grand Canyon (area 1 of Fig. 3.3), with thicknesses variously reported to be 3020 feet (921 m), 3115 feet

(950 m), and 3230 feet (985 m). The Dox consists of four members: (in ascending order) Escalante Creek, Solomon Temple, Comanche Point, and Ochoa Point. West of 75-mile Creek (western part of area 1, Fig. 3.1), only the Escalante Creek and Solomon Temple members are preserved; the Comanche Point and Ochoa Point have been removed by pre-Tapeats erosion.

The Escalante Creek Member is reported by Stevenson (1973) to be 1280 feet (390 m) thick where exposed in the eastern Grand Canyon. Here, it is a light-tan to greenish brown, siliceous quartz sandstone and calcareous lithic and arkosic sandstone that is 800+ feet (244+ m) thick, combined with an overlying 400-feet (122-m)-thick sequence of dark-brown-to-green shale and mudstone. The tan to brownish color of this lower member is in marked contrast to the characteristic red and red-brown color of the rest of the Dox Formation.

Sedimentary structures observed in the sandstones of the Escalante Creek included contorted bedding (within 100 feet [30 m] of the base), small-scale, tabular-planar cross beds, and graded beds (with shale interclasts at the base). The contacts between members of the Dox Formation are gradational and are based mainly on topographic expression, depositional environments, and color changes. The Escalante Creek is tan to brown and forms a cliff-slope topography, as opposed to the more red-orange overlying Solomon Temple Member, which forms rounded-hill topography that is more characteristic of the remainder of the Dox.

The Solomon Temple Member is a cyclical sequence of red mudstone, siltstone, and quartz sandstone. It is 920 feet (280 m) thick in the eastern Grand Canyon. The lower 700 feet (213 m) is a slope-forming series of predominantly

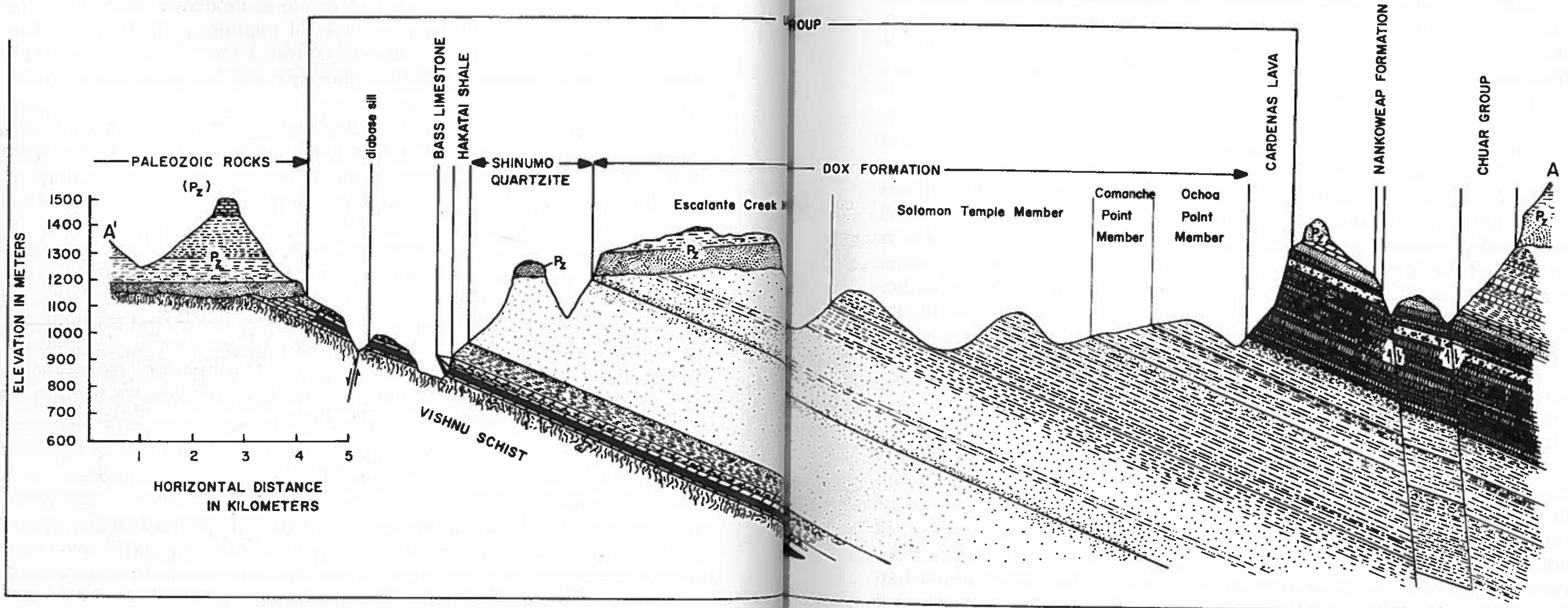

FIGURE 3.3. Cross section A-A' (eastern Grand Canyon). See Fig. 1.1 for location of section.

red-to-maroon shaley siltstone and mudstone with subordinate quartz sandstone. The upper 220 feet (67 m) of the member is primarily maroon quartz sandstone with numerous channel features. These channels, along with common low-angle, tabular cross beds, suggest a floodplain environment.

The Comanche Point Member occupies more than half of the Dox outcrop area and is distinguished from enclosing members by its slope-forming and color-variegated character. This member varies in thickness from 425 to 617 feet (130 to 188 m) in the eastern Grand Canyon. The lithology is primarily shaley siltstone and mudstone with minor amounts of sandstone. Five pale green-to-white, leached red beds, some up to 40 feet (12 m) in thickness, provide the variegated appearance of this unit. Stromatolitic dolomite layers are found within or directly adjacent to the leached beds. Sedimentary features found in this member include ripple marks; mudcracks and curls; salt casts; and wavy, irregular bedding.

The Ochoa Point Member is 175 to 300 feet (53 to 92 m) thick and forms steep slopes and cliffs below the Cardenas Lava. It consists of micaceous mudstone that grades upward into a predominantly red quartzose, silty sandstone. Sedimentary structures of this member also include salt crystal casts in the mudstone and asymmetrical ripple marks and small-scale cross beds in the sandstones.

The lithology and sedimentary structures found within the Dox Formation suggest, in ascending order, a subaqueous delta, floodplain, and tidal flat environment during deposition. According to Stevenson and Beus (1982), features of the lowermost member, the Escalante Creek, record a rapid transgression of the Dox sea followed by gradual basin filling.

This basin was filled by the close of Escalante Creek time, and the region was at or very near sea level for the remainder of Dox time. Stevenson and Beus (1982) also suggest a possible westerly source for some of the sediments of the Dox, which is opposite to the inferred source direction for other formations of the Unkar Group.

The Cardenas Lava

Cardenas Lava is the name given to a series of basalt and basaltic andesite flows and sandstone interbeds that are stratigraphically above the Dox Formation and below the Nankoweap Formation (Ford et al. 1972). This sequence of rocks is exposed only in the eastern Grand Canyon, where the thickness of the formation ranges from about 785 (240 m) to nearly 985 feet (300 m). The contact between the Cardenas Lava and the overlying Nankoweap Formation is unconformable, with an unknown amount of the Cardenas being removed prior to Nankoweap deposition.

The contact between the Dox Formation and the Cardenas is conformable and interfingering. This is highlighted by the presence of a thin, discontinuous basaltic flow in the upper Dox a few meters below the top of the formation. The Dox near the contact is mildly baked and locally displays small folds and convolutions that are suggestive of soft sediment deformation. At one location, the basalt of the lowermost flow sequence contains rounded masses of igneous rock, up to 3.3 feet (1 m) in diameter, completely surrounded by a thin layer of siltstone of Dox lithology. The uppermost Dox is fine-grained sandstone and siltstone deposited in a tidal flat environment. The region at the time of initial Cardenas eruption was at or very near sea level (Stevenson and Beus 1982). One interpretation is that the lava flowed over unconsolidated sandy and silty Dox sediments that were wet at the time. Whether these sediments were slightly above or slightly below water level is unknown.

Strictly on the basis of topography, the Cardenas Lava can be divided into two units (Hendricks and Lucchitta 1974). The lower unit forms granular slopes and is from 245 to 295 feet (75 to 90 m) thick. Although this unit is highly altered and weathered, many of the primary features are preserved. This "bottle-green member" (Lucchitta and Hendricks 1983) is a composite of many thin, discontinuous flows and sandstone interbeds. The basalt of this unit is broken and weathers into nodules typically 3 to 10 inches (10 to 30 cm) in diameter. Near the top of this unit, some 230 feet (70 m) above the base, the basalt is more massive and less altered. Petrographically, the lower unit is an olivine-rich basalt with a subophitic texture. Lower in the bottle-green member, the rock is highly altered but has a texture that suggests it may have been quite glassy originally.

Chemically, rocks from the lower unit are high in sodium and magnesium and depleted in potassium, suggesting spilitic alteration. The nodular character, glassy texture, and anomalous chemistry of rocks of the lower unit suggest rapid quenching in sea or brackish water. The thin, discontinuous sandstone interbeds also indicate that water flowed over or was ponded on the lavas during periods of volcanic quiescence.

Approximately 328 feet (100 m) above the base of the Cardenas is a continuous sandstone bed some 16 feet (5 m) in thickness that overlies the bottle-green member. This sandstone is laminated and forms vertical cliffs. In a number of locations, the sandstone occupies channels in the upper surface of the bottle-green flow member. The petrology of this sandstone suggests quiet-water deposition; the channels, therefore, probably are lava channels that were left as extrusion temporarily ceased and basin subsidence continued, thereby lowering the lava surface below sea level.

The upper unit of the Cardenas is a series of cliff-forming basalt and basaltic andesite flows along with additional sandstone interbeds. Features preserved in the individual flow units suggest that the volcanic pile accumulated at a slightly greater rate than basin subsidence. In ascending order, these are: an autoclastic breccia directly above the 328-foot (100-m) sandstone, a fan-jointed unit, ropy lava, and, finally, a lapillite unit at the 754-foot (230-m) level. Following eruption of the lapillite, volcanic activity ceased for a short period of time. The pyroclastic surface was smoothed as subsidence continued until the surface again was lowered below sea level. This is noted by a planar upper surface on the lapillite unit with a continuous sandstone layer directly above. At least two more eruptive events followed deposition of this sandstone layer.

Following cessation of volcanic activity, the sediments and igneous rocks of the Unkar were tilted gently toward the northeast. An unknown amount of Cardenas Lava was eroded prior to Nankoweap deposition. Elston and Scott (1976) suggested that major tectonic movement occurred along the Butte fault in the eastern Grand Canyon during the post-Cardenas, pre-Nankoweap interval.

Intrusive Rocks of the Unkar Group

Diabase sills and dikes intrude all formations of the Unkar Group below the Cardenas Lava. Sills are restricted to the Bass Limestone and Hakatai Shale. Dikes intrude the Hakatai Shale, Shinumo Quartzite, and Dox Formation above the sills. Feeder dikes or vents for the sills are not exposed. Above the sills, dikes can be traced, discontinuously, to within a few meters of the base of the Cardenas Lava.

Unkar sills range in thickness from 75 feet (23 m) at Hance Rapids in the eastern Grand Canyon to 985 feet (300 m) in Hakatai Canyon (Shinumo Creek area). Fine-grained, chilled margins suggest that the magma was highly fluid at the time of intrusion. Only the sills of the Shinumo Creek area show extensive differentiation and segregation products. Here, a picritic layer of unknown thickness is present near the base of the sill, while the top of the intrusion is marked by a 20-foot (6-m)-thick granophyre layer.

Contact metamorphism caused by intrusion of the diabase resulted in the formation of chrysotile asbestos above the sills where the magma intruded the Bass Limestone. Adjacent to the sills, the Hakatai Shale is a knotted hornfels containing porphyroblasts of andalusite and cordierite that have been replaced by muscovite and green chlorite, respectively.

The relation among the sills, dikes, and Cardenas flows is not self-evident. The mineralogy of the sills is uniform throughout the region and is identical to that of the unaltered parts of the bottle-green member of the Cardenas. Chemical variation diagrams (Hendricks and Lucchitta 1974) indicate that the flows are more felsic than the sills, but the evidence does not preclude the possibility of a common parentage. Paleomagnetic evidence (Elston 1986) suggests that the majority of the sills were intruded at the same time that the Comanche Point Member of the Dox Formation was being deposited. The Hance sill, the easternmost of the Unkar sills, produces an anomalous paleomagnetic pole position that Elston (1986) interpreted as an indication of a slightly greater age for this intrusion. If the majority of the sills were emplaced during Comanche Point time, approximately 330 feet (100 m) of sediment of the Comanche Point and Ochoa Point members of the Dox was deposited during the time interval between intrusion of the sills and initial eruption of the Cardenas Lava. Some of the dikes in the eastern Grand Canyon have paleomagnetic pole positions similar to the Cardenas and may represent feeders. These dikes cannot be seen connecting to the nearby Hance sill.

The sills, dikes, and flows of the Unkar Group may represent a single volcanic episode. If so, the earliest phases were intrusions of diabase sills followed by a period of igneous quiescence long enough for accumulation of approximately 300 feet (100 m) of Dox sediment. Later phases were eruptions of basalt and basaltic andesite flows via a network of thin dikes. Elston and McKee (1982) indicated that initial strontium-87/strontium-86 ratios are slightly different for a sill in the Shinumo Creek area and the Cardenas Lava and concluded that either the sill and flows are not comagmatic or the magma acquired ^{87}Sr from passing solutions. These ratios [0.7042 ± 0.0007 (sill) and 0.7065 ± 0.0015 (flows)], do not preclude a common parentage for the intrusive and extrusive rocks. This might occur if the magma ponded in the crust long enough for derivation of the basaltic andesite of the Cardenas Lava from the basaltic magma of the sills, provided that the intruded crust was assimilated and that it had an $^{87}Sr/^{86}Sr$ ratio greater than 0.7065.

OVERVIEW OF GEOLOGIC HISTORY

Prior to the beginning of deposition of the Unkar sediments, the region of the Grand Canyon was along the southwestern margin of the North American craton. The surface was a smooth terrain composed of Vishnu granitic and metamorphic rocks. Pre-Unkar relief on this surface was undulating with a relief on the order 30 to 500+ feet (a few tens to a few hundreds of meters). Relatively

high terrain may have existed to the east and northeast, with probable open oceans to the west. As the land surface subsided, the ocean advanced eastward, marking the beginning of Unkar deposition about 1250 million years ago.

The Hotauta Conglomerate Member of the Bass Limestone, lowermost unit of the Unkar Group, was deposited in relatively low areas of the Vishnu terrain. The clasts size suggests an easterly source area. During this initial transgression, the sea advanced at least as far as Hance Rapids in the eastern Grand Canyon—and probably much farther. Sediments in the middle of the Bass Limestone in western Grand Canyon suggest that this area was below wave base and away from any strong currents. At the same time, indicators in the east suggest an intertidal-to-supratidal environment.

The initial transgression of the sea was followed by a gradual regression, which resulted from sediment accumulation. Sedimentary features found in the upper part of the Bass Limestone and Hakatai Shale indicate that conditions varied from subaerial to subaqueous throughout the region during this of time. Tectonic activity along northeast-trending, high-angle reverse faults marked the end of deposition of the Hakatai Shale and resulted in a period of erosion prior to the accumulation of sands of the Shinumo Quartzite.

Near-shore fluvial and deltaic conditions predominated in the region during Shinumo and early Dox time, marking a further subsidence of the region and the second transgression of the sea; marine conditions returned by the close of this period. Following deposition of about 165 feet (50 m) of Dox sediments, rapid subsidence caused the sea to advance further eastward followed by a sustained period of basin filling. Subaerial conditions returned by the close of Solomon Temple time. The contact between the Solomon Temple and Comanche Point members of the Dox Formation marks another transition to marine conditions with the remainder of the Dox environment fluctuating between marine and nonmarine conditions.

Features found in the lowermost Cardenas Lava suggest the outpouring of the basalt onto wet, probably shallow-water Dox sediments. Sporadic accumulation of the lava pile and continued basin subsidence resulted in conditions that varied between marine and nonmarine, but, generally, the flows accumulated at a greater rate than that of regional subsidence. Following extrusion of more than 985 feet (300 m) of Cardenas lava, tectonic uplift raised and tilted gently the region of the eastern Grand Canyon toward the northeast. Subsequent erosion removed an unknown amount of the lavas prior to deposition of sediments of the Nankoweap Formation.

INTER-REGIONAL CORRELATIONS

The only other unmetamorphosed sequence of rocks that is younger than the 1700-million-year basement complex and older than Cambrian in the region is found in central Arizona. These rocks, the Apache Group and Troy Quartzite, are similar to those of the Unkar Group.

Shride (1967) has suggested a possible correlation of the Unkar Group with the Apache Group. This is based on similarities in the age and lithology of the two units. Elston (1986) reviewed the correlation of various Middle and Late Proterozoic sequences on a paleomagnetic basis. He concluded that the Mescal Limestone of the Apache Group correlates with the middle part of the Dox Formation and that sills in the Unkar Group are similar paleomagnetically to sills that intrude both the Apache Group and Troy Quartzite. Elston (1986) also provided possible correlations of the Unkar Group with other Middle Proterozoic sequences of North America. These include the Uinta Mountain Group of Utah and Colorado, the Belt Supergroup of Montana and Idaho, the Sibley Series of Ontario, and the Keweenawan Supergroup of the Lake Superior Region.

SUMMARY

Sediments and lavas of the Unkar Group of the Grand Canyon Supergroup were deposited in a basin along the western edge of the North American craton during the period 1250 to 1070 million years ago. Features preserved in the Unkar record a west-to-east transgression of the sea with minor sea-level variations resulting from basin filling and subsidence. About 5800 feet (1770 m) of sediment was deposited in the Unkar basin before the onset of volcanic activity in the area. Unkar sedimentary rocks presently are exposed in isolated outcrops in the Grand Canyon and have been subdivided into formations; in ascending order they are, the Bass Limestone, Hakatai Shale, Shinumo Quartzite, and Dox Formation. A volcanic episode marked the end of sedimentation in the Unkar basin. During this volcanic period, igneous material formed sills in the lower parts of the Unkar and dikes in the upper parts. Lava was erupted onto sediments of the Dox Formation and accumulated to a thickness of nearly 1000 feet (305 m). These extrusive rocks have been named the Cardenas Lava. Following cessation of volcanic activity, the Unkar Group rocks were tilted slightly and the top of the Cardenas Lava eroded before the onset of further sedimentation in the region.

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GRAND CANYON SUPERGROUP: NANKOWEAP FORMATION, CHUAR GROUP, AND SIXTYMILE FORMATION

Trevor D. Ford and Carol M. Dehler

INTRODUCTION

The upper half of the younger Precambrian strata of the Grand Canyon presents a sequence of approximately 6800 feet (2100 m) of rocks not exposed anywhere else in the southwestern United States. They have not been metamorphosed, and the sedimentary rocks include an unparalleled assemblage of late Precambrian fossils. For these reasons, the Nankoweap Formation, Chuar Group, and Sixtymile Formation deserve a chapter devoted to themselves. Furthermore, they provide unique evidence of sedimentation during tectonic activity, with examples of intraformational faults and unconformities in the Nankoweap Formation and the Chuar Group, as well as slump features in the Sixtymile Formation.

The younger Precambrian rocks of the Grand Canyon first were recognized as "Algonkian" by Powell (1876), who named them the Grand Canyon Series. Now redesignated as the Grand Canyon Supergroup (Elston and Scott 1976), the upper half comprises the Nankoweap Formation, overlain by a thick Chuar Group and a thin Sixtymile Formation. Walcott (1894) presented the first detailed descriptions, but divided the Supergroup into the Unkar and Chuar "terranes." Because the Unkar Group is treated elsewhere in this volume, only the younger part of the younger Precambrian is covered in this chapter.

The Nankoweap Formation and Chuar Group are exposed in the wide eastern part of the Grand Canyon, clearly visible from the Desert View Tower overlook (Fig. 4.1). The former is present in cliffs overlooking Basalt, Tanner, and Comanche Canyons and is visible from the river. The Chuar Group is only visible from the river immediately north of Basalt Canyon and crops out dominantly in the upper parts of several right-bank tributaries—notably Nankoweap, Kwagunt, Carbon, Chuar, and Basalt Canyons. The outcrops of the Chuar Group are bounded on the east by the Butte fault, and on all other sides by Powell's Great Unconformity and the overlying Cambrian Tapeats Sandstone (Fig. 4.1). The Sixtymile Formation is only exposed in the infrequently visited Sixtymile and Awatubi canyons and also caps Nankoweap Butte, which makes up the divide between Nankoweap Creek to the north and Kwagunt Creek to the south (Fig. 4.1).

The north rim overlooks at Point Imperial and Cape Royal provide fine views of the Chuar Group outcrops in these tributary canyons. Observers at Cape Royal can clearly see that the Chuar Group has been folded into a prominent

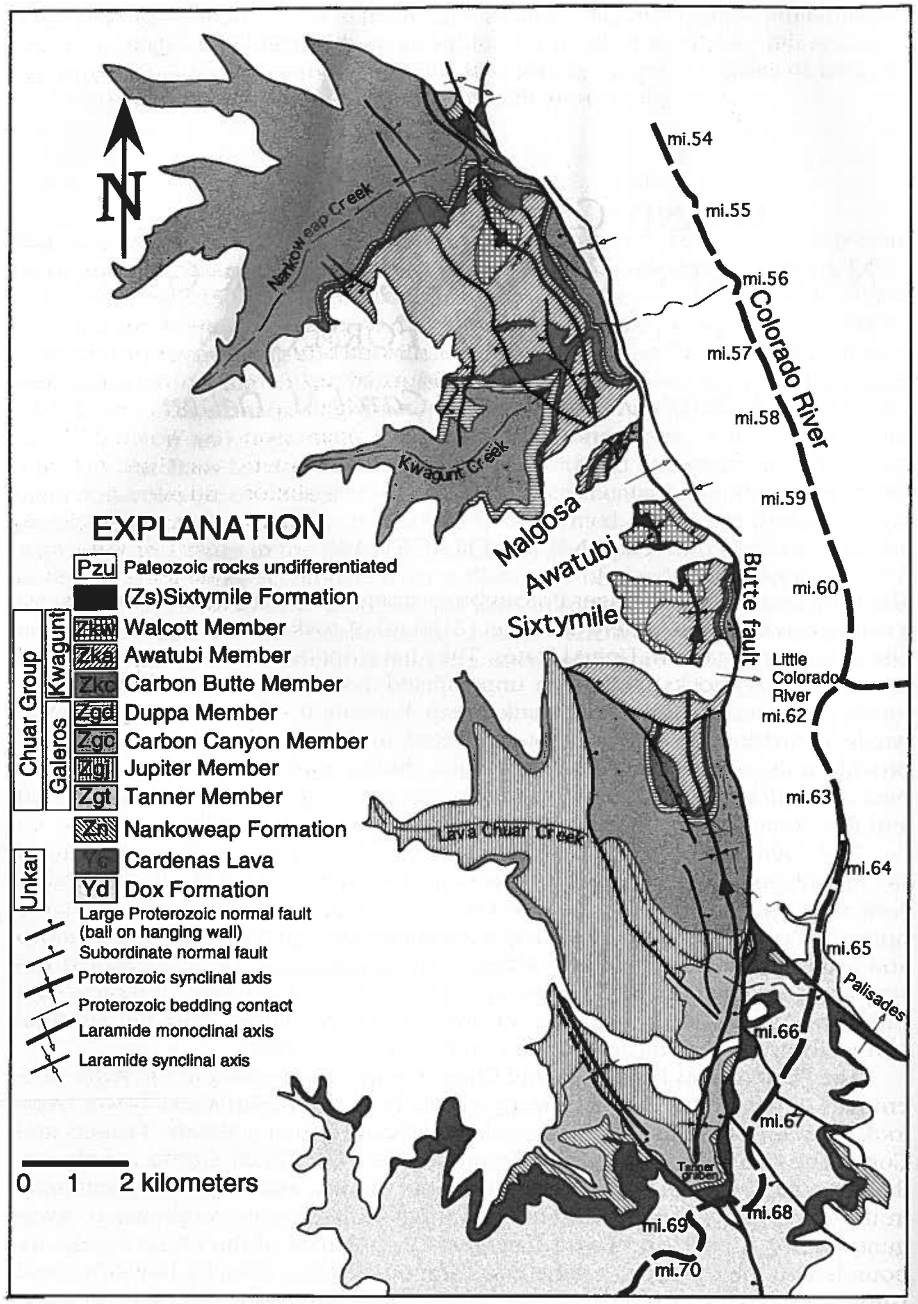

Figure 4.1. Geologic map of the upper Unkar Group, Nankoweap Formation, Chuar Group, and Sixtymile Formation from Nankoweap Canyon southward to Basalt Canyon. (Timmons, unpublished data, 1998.)

north–south-trending syncline, west of and parallel to the Butte fault (Fig. 4.1). Displacement on the Butte fault was of the order of 10,500 feet (3200 m) down to the west in late Precambrian time, but this has been canceled out partly by west-side-up reactivation in Post-Paleozoic times of about 2660 feet (810 m).

THE NANKOWEAP FORMATION

In the middle part of the younger Precambrian sequence of the eastern Grand Canyon, a distinctive group of red-brown and tan sandstones, and subordinate siltstones and mudrocks, lies unconformably on top of the Cardenas Lava. The rocks crop out from just south of Carbon Canyon to Basalt Canyon on the west bank of the river and around Comanche Creek and Tanner canyon on the east bank. Splays of the Butte fault offset the Nankoweap Formation on both sides of the river. These sedimentary deposits were originally included partly in the top of the Unkar Group and partly in the basal Chuar Group by Walcott (1894), but were separated as a new unit, the Nankoweap Group, by Van Gundy (1937). Later, in 1951, Van Gundy gave a more complete description and noted the presence of unconformities at both the upper and lower contacts. Maxson (1968) introduced the term "Nankoweap Formation" on his geologic maps of the Bright Angel Quadrangle and eastern Grand Canyon, but made no comment on the unconformities. The name is taken from a small, fault-bounded block of the Nankoweap Formation in Nankoweap Canyon, although more extensive outcrops exist above and adjacent to the Cardenas Lava in Basalt Canyon (Fig. 4.3), Comanche Creek, and Tanner Canyon.

The Nankoweap Formation is divided into two informal members (Elston and Scott, 1976). The lower (ferruginous) member is 40 feet (13 m) thick and rests disconformably on the Cardenas Lava. It is composed of red, fine-grained quartzitic sandstones and siltstones with hematite laminae and lenses of volcanic detritus derived from the Cardenas Lava. The upper member, 330 feet (100 m) thick, disconformably overlies the lower member and is composed mainly of fine-grained quartzitic sandstones that are shaley and silty towards the top. These sandstones are thin-to-medium bedded, with cross-beds, ripplemarks, mudcracks, soft-sediment deformation, and rare salt pseudomorphs. Elston and Scott (1976) interpreted the lower member to have been deposited in quiet shallow water in a structurally controlled lake or pond. Sedimentary structures in the upper member suggest a moderate to low energy, shallow water, marine or lake environment.

Although exposures in the cliffs immediately north of Tanner Rapids show well-bedded, red-brown sandstone dipping evenly at about five degrees to the northeast, the upper parts of Basalt and Tanner Canyons reveal more complex stories, once post-Nankoweap faulting is removed. Substantial paleotopography cut into the Cardenas Lava can be observed in Tanner Canyon and Tanner Graben, as well as the disconformity that separates the lower and upper members. Some of this paleotopography is controlled by growth faults, which offset the Cardenas Lava and the lower member, yet not the overlying upper member (Elston and Scott, 1976). A growth fault was recently discovered within the upper member on the south-facing side of Basalt Canyon, signifying that syntectonic deposition did occur during upper member time.

No direct dating is possible on the Nankoweap Formation. The age of this formation can be bracketed between the 1070 ± 70-million-year age of the underlying Cardenas Lava and associated diabase dikes and sills (Elston and McKee 1982 and references therein) and the overlying Chuar Group, which Elston

(1979) has argued was terminated by the "Grand Canyon orogeny" approximately 823 million years ago. Limited paleomagnetic data (Elston and Grommé, 1974; Elston and McKee 1982; Elston 1989a) are consistent with an age of ~1070–1050 Ma for deposition of the lower member and ~950 Ma for deposition of the upper member.

Paleontology

A structure found in a sandstone bed of the Nankoweap Formation in Basalt Canyon was identified as a trace fossil impression of a stranded jellyfish (Van Gundy 1937, 1951). It comprises a series of lobes rounded at the extremities. Some lobes have a median groove radiating from a small, irregular hollow, and the whole structure is approximately five inches (12 cm) in diameter. Hinds (1938) and Bassler (1941) also considered this to be a jellyfish impression. It was named *Brooksella canyonensis* by Bassler as a new species of a genus well known in the Paleozoic Era, though the interpretation of the genus as a jellyfish is still in considerable doubt.

Cloud (1960, 1968) subsequently obtained a partial second specimen but claimed that the structures were of inorganic origin formed by "compaction of fine sands deposited over a compressible but otherwise unidentifiable structure, possibly a small gas blister." Glaessner (1969) was unconvinced by Cloud's explanation and drew comparisons with a Mesozoic stellate trace fossil, *Asterosoma*, deducing that it was of organic origin and that its "possible originator was a sediment feeder able to burrow into the sediment . . . worm-like in shape . . . probably an annelid."

Accordingly, Glaessner (1984) renamed the structure *Asterosoma? canyonensis* (Bassler, 1941) and apparently still accepts the trace fossil interpretation in spite of its age. In a 1981 study, Kauffman and Steidtmann (1981) supported Glaessner's interpretation as a burrow made by a sediment-feeding, worm-like organism.

An examination of both specimens revealed a similarity to small "sand-volcanoes" formed by the upward expulsion of gas or fluid from sediments as more sediment is loaded on top or as the sediment is shaken during seismic activity. These soft-sediment deformation features are ubiquitous in sandstone beds of the Grand Canyon Supergroup. In view of this, it is difficult to support Glaessner's interpretation of the phenomenon as trace fossils. Additionally, sinuous, high-relief ridges on the soles (undersides) of sandstone beds resemble infilled burrows, yet are more likely the partial casts of mudcracks, another ubiquitous sedimentary feature in the Grand Canyon Supergroup.

If the age of the Nankoweap Formation is ~1070–950 Ma, and if these specimens are truly fossils, whether burrows or jellyfish impressions, these specimens could be the earliest record of complex life on earth. A more intensive search of Nankoweap Formation outcrops for comparable structures obviously is needed.

THE CHUAR GROUP AND SIXTYMILE FORMATION

The Chuar Group has been subdivided by Ford and Breed (Ford et al. 1972a; Ford and Breed 1973b) into two formations and seven members. The Sixtymile Formation was originally included as the top member of the Chuar Group until it was recognized as a distinct coarse-grained unit and moved to formation status. The stratigraphic nomenclature is summarized in Table 4.1 and shown graphically in Figure 4.2.

TABLE 4–1. Stratigraphic Subdivisions of the Chuar Group and Sixtymile Formation Including Thickness Measurements by Various Workers[a]

			Thickness [in feet (meters)]			
Group	*Formation*	*Member*	*Ford and Breed (1973b)*	*Elston (1989) and references therein*[b]	*Cook (1991)*	*Dehler and Timmons (unpublished data, 1998)*
Chuar Group	Sixtymile Formation		118 (36)	194–210 (59–64)		
	Kwagunt Formation	Walcott Member	836 (255)	922 (281)	758–804 (231–245)	823 (251)[c]
		Awatubi Member	1128 (344)	987 (301)		653 (199)[c]
		Carbon Butte Member	249 (76)	164 (50)		112–223 (34–68)[c]
	Galeros Formation	Duppa Member	571 (174)	341 (104)		2050 (625)[c]
		Carbon Canyon Member	1545 (471)	1148 (350)		2918 (889.5)[c]
		Jupiter Member	1515 (462)	1424 (434)		868 (264.5)
		Tanner Member	640 (195)	512 (156)		604 (184)

[a]For locations of measured sections, see individual publications.

[b]Sixtymile Formation thickness from Elston (1979) and Elston and McKee (1982), and Chuar Group (member) thicknesses modified by Reynolds and Elston (1986) and Reynolds, written communication to Elston (1988).

[c]Member measured sections or intramember partial measured sections that show lateral thickness variations. Thickness variations observed by Dehler and Timmons (unpublished data, 1998) correspond with structural trends across the Chuar syncline that may explain the thickness variability amongst workers.

Reynolds and Elston (1986) have revised the thickness because they found the Galeros Formation to be only 3000 feet (915 m) thick and the Kwagunt Formation to be 2083 feet (635 m). Ongoing work by Dehler and Timmons (unpublished data, 1998) has revealed thickness variations across the Chuar syncline (west–east) in stratigraphic intervals within the middle and upper Chuar Group (Table 4.1). These thickness variations may be attributed to, in part, lateral facies changes, along with changes in accommodation space controlled by movement of the syncline during deposition. The thickness variations have been documented in the upper Carbon Canyon Member and younger members, implying that the Chuar syncline was developing as these sediments were being deposited.

The Chuar Group shows an overall pattern of repeating carbonate–shale cycles on the order of hundreds of meters thick. In the middle of the Chuar Group,

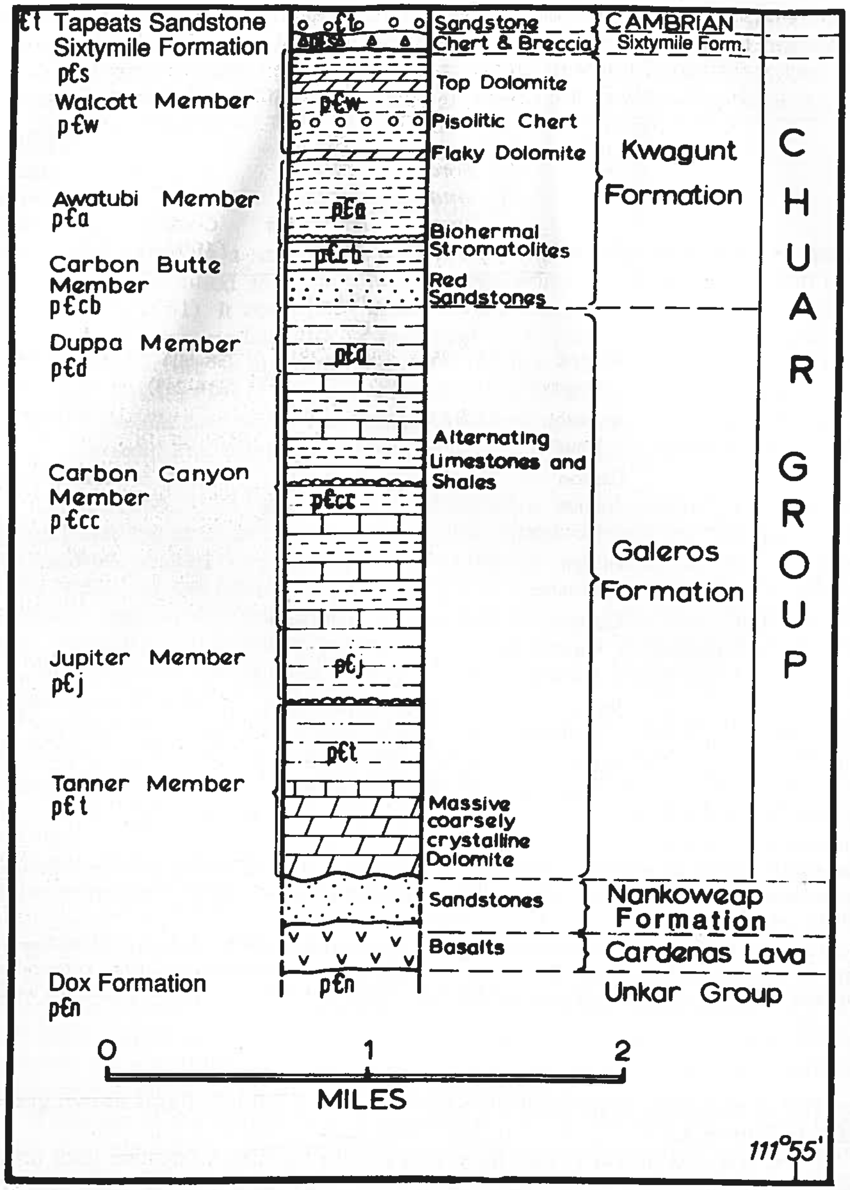

FIGURE 4.2. Schematic stratigraphic column of the Nankoweap Formation and Chuar Group.

the carbonate–shale cycles are interrupted by a distinctive red sandstone unit, the Carbon Butte Member, which marks the boundary between the Galeros Formation below and the Kwagunt Formation above. The Galeros Formation comprises the Tanner, Jupiter, Carbon Canyon, and Duppa Members; and the overlying Kwagunt Formation comprises the Carbon Butte, Awatubi, and Walcott Members. Research thus far suggests that contacts between members are gradational. These gradational contacts are also an attribute to the different thickness measurements attained by different workers (Table 4.1). Specifically, the contacts between the Jupiter, Carbon Canyon, and Duppa Members are very difficult to distinguish because there are no distinctive marker units to delineate these contacts, rather they are defined by the presence or absence of carbonate beds that are known to change lithologic character laterally.

GALEROS FORMATION

Tanner Member

The Tanner Member, which overlooks the Tanner Rapids on the Colorado River, consists of 20 to 80 feet (6 to 24 m) of thickly bedded, coarsely to finely crystalline dolomite at the base, and 580 feet (177 m) of almost entirely shales above. The basal Tanner dolomite disconformably fills in paleotopography cut into the upper member of the Nankoweap Formation. Sedimentary structures in the dolomite include parallel horizontal laminations and intraclast horizons. It forms a massive ledge capping the cliffs around Basalt Canyon (Fig. 4.3) and also crops out on the southern flank of Chuar Canyon. A fault-bounded sliver of Tanner dolomite exists on the north side of Nankoweap Canyon. The Tanner dolomite was included in the Unkar Group by Walcott, but it was transferred to the Chuar Group by Van Gundy (1951).

The overlying 177 meters of the Tanner Member is predominantly composed of shales, along with subordinate siltstones, sandstones, and dolomites. These strata are exposed throughout much of Basalt and Chuar Canyons. The shales are finely laminated to massive, are predominantly black, and weather ocherous yellow, orange, red-purple, pale green, and gray. The shales commonly have very thin to thin lenses and tabular beds of white to green siltstone and fine-grained sandstone. *Chuaria circularis* has been found in the upper 50 meters of black shale. Hematitic cement is common within the shales and sometimes weathers to goethite box-stones. Thicker sandstone and dolomite beds are present toward the top of the member. The sandstones are thinly to thickly bedded, green, and fine-grained, and they exhibit rare ripplemarks and mudcrack casts. The dolomite beds are massive and medium-bedded, and they are only found in the upper 2 meters of the member.

Jupiter Member

The Jupiter Member also consists of carbonates below and shales above. The basal division is about 40 feet (12 m) of stromatolitic limestones and dolomites, including undulating and broad-domed forms of algal laminate within a mass of dolomitized tufa-like rock, with flat-pebble conglomerates at the base. The upper part of the carbonate member has layers with abundant casts of gypsum crystals and a few, poorly defined and solitary stromatolite columns. These are similar to the form *Inzeria*, and to undulating stromatolites of the form *Stratifera*. The remainder of the 1516-foot (462-m)-thick member is predominantly shale, but commonly includes thin beds of sandstones and siltstones. The shales are variable in color, from red-purple to ocherous yellow to pale green to blue-black, and are commonly micaceous. Rare *Chuaria circularis* has been found in the black shales. The sandstones and siltstones are rarely more than a few inches thick and have abundant symmetric ripplemarks and mudcrack casts, as well as

Figure 4.3. Upper Basalt Canyon showing cliffs of Cardenas Lava in the bottom left corner, overlain by the flat-topped terrace of Nankoweap sandstones and Tanner Member dolomite, with shales up to the basal stromatolitic layers of the Jupiter Member, extending upwards toward the Great Unconformity beneath the Cambrian Tapeats Sandstone. Photograph by Parker Hamilton.

soft-sediment deformation features, ripple laminations, and rare raindrop prints and salt pseudomorphs. Hematitic cement is common in patches and sometimes weathers to goethite box-stones.

Carbon Canyon Member

The Carbon Canyon Member consists of alternating carbonate, shale, and sandstone beds [each a few feet thick (1 m)], attaining a total thickness of 1546 feet (471 m) (Fig. 4.4). The carbonates commonly are 3 to 6 feet (1–2 m) thick, and almost all are dolomicrite to dolosiltite with local chert nodules and common

Figure 4.4. The Chuar syncline looking north along its axis. Chuar Canyon in lower left; Carbon Canyon in the right center. The ridge in the middle foreground is of alternating dolomites, sandstones, and shales of the Carbon Canyon Member. The synclinal cuesta in the middle distance is made up of the Carbon Butte sandstone. The Butte fault trends north through the shadows on the right. (Photograph by John Shelton.)

laminations of quartz siltstone. In places, the carbonates grade into calcareous siltstones. Many of the carbonates have irregular (crinkly) laminations, which are interpreted to be of algal origin. The tops and bottoms of the carbonate beds commonly have symmetric ripplemarks and mudcracks. A dolomite marker bed in the upper part of this member exhibits a distinctive horizon of possible large-scale mudcracks (<0.5 m deep) that have undergone soft sediment deformation. These possible mudcracks form polygonal patterns on bedding planes, and probably indicate prolonged periods of exposure. Many carbonate beds grade from yellowish-tan crinkly laminated dolomites into reddish-orange massive dolomites, probably representing shallowing-upward cycles.

The interbedded fine-grained siliciclastics vary from blue-black, micaceous shale to red and green mudstones. Sandstones are fewer in number and rarely more than 2 feet thick (61 cm). They generally are green to gray to tan in color and have subangular to rounded quartz grains set in carbonate, silica, hematite, or chlorite cement, or clay matrix. Many of the fine-grained sandstones contain medium- to coarse-grained, well-rounded quartz sand that is randomly oriented in the fine-grained matrix. Mudcrack casts are common in the sandstones, and some laminae show truncated, incipient cracks that look misleadingly like worm tracks on weathered joint faces. Other sedimentary structures in the sandstones include symmetric ripplemarks, interference ripples, ripple laminations, soft-sediment deformation features, low-angle planar crossbeds, and trough crossbeds.

The stromatolites, located toward the top of the member, take the form of sharply widening, irregularly branching columns that show strongly convex laminae. These columns are clustered into domes, usually about 1.6 feet (0.5 m) in diameter and 1 foot tall (0.35 m), and taper at the base. Some clusters have been observed that are much as 6 feet (1.8 m) in diameter, yet still maintaining a height of 1 foot (0.5 m). The domes are commonly closely spaced, sometimes growing into one another. This stromatolitic horizon shows little lateral variation and is commonly interbedded with black shales and intraclastic dolomite. Dolomitization has destroyed internal detail, but the stromatolites appear to fall within the form *Baicalia* (probably *B.* aff. *rara*) Semikhatov.

Duppa Member

The Duppa Member is over 570 feet (174 m) thick and marks the return to a shaley character with only a few thin beds of dolomite. Calcareous siltstone beds up to 3 feet (1 m) thick are found throughout the member and contain well-rounded silt grains. The rest of the member is shale (generally micaceous) that grades into red mudstones and thinly bedded sandstones and siltstones toward the top. The contact with the overlying Kwagunt Formation is gradational.

KWAGUNT FORMATION

Carbon Butte Member

Because the Carbon Butte Member has at its base the only thick sandstone in the Chuar Group, it provides a distinctive marker for the base of the Kwagunt Formation. The member is 252 feet (76 m) thick at its type locality yet is found to vary in thickness across the Chuar syncline (Table 4.1). The basal 80 feet (24 m) comprises thickly bedded, tan to red, fine- to medium-grained sandstones

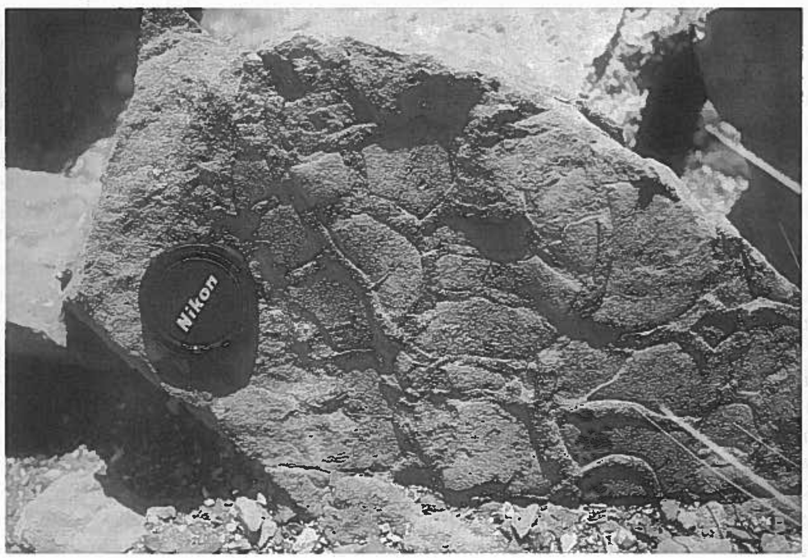

Figure 4.5. Mudcrack casts in the basal red sandstone of the Carbon Butte Member, Carbon Butte. Lens cap for scale. (Photo by Mike Timmons.)

with interbedded shales and siltstones which have weathered into a prominent cliff that surrounds Carbon Butte (Fig. 4.4). Sedimentary structures include 3- to 6-feet thick (1- to 2-m thick) cross-bed sets (epsilon?), symmetric ripplemarks, trough cross-beds, ripple laminations, soft-sediment deformation features, and mudcrack casts (Figs. 4.5 to 4.7). The overlying 170 feet (52 m) consists of predominantly red to purple mudstones and shales with subordinate thin to medium beds of fine- to medium-grained sandstone and siltstone. Some of these beds have mudcrack casts. In the upper part of this member is a 9-foot (3-m)-thick unit of white sandstone that exhibits very well preserved symmetric ripplemarks, interference ripplemarks, soft sediment deformation features, and trough cross-beds. Preliminary results from paleocurrent analyses indicate a dominant symmetric-ripple-crest trend of about north–south, similar to the trend of the Butte fault.

Figure 4.6. Symmetric ripplemarks in the basal red sandstone of the Carbon Butte Member, Carbon Butte. Preliminary paleocurrent data show that symmetric ripple crests are oriented north–south and parallel the Butte fault. Hammer for scale. (Photo by Mike Timmons).

FIGURE 4.7. Cross-bedding (epsilon?), and soft-sediment deformation in the basal red sandstone of the Carbon Butte Member, Carbon Butte. (Photo by Mike Timmons.)

Awatubi Member

The Awatubi Member comprises a basal stromatolitic carbonate unit overlain by dominantly shales and mudstones of varying colors. The basal 12 feet (3.7 m) of this 1128-foot (344-m)-thick member consists of biohermal domes, each 8–10 feet (2.5–3 m) in average height and width. These domes are made up of a complex of columns 2–3 inches (5–7.5 cm) in diameter, and are interbedded with confluent domes (Figs. 4.8 and 4.9). The columns show nearly flat laminae and generally are parallel-sided with rare bifurcation. Dolomitization has destroyed most of the internal detail, but a clearly defined wall is present, indicative of the form *Boxonia* Koroljuk. The matrix between columns generally is crystalline dolomite, whereas that between bioherms is coarsely granular dolomite of a highly porous nature. Flat-pebble conglomerates are present at the base of some bioherms.

The overlying variegated shales are interbedded with thin to very thin beds of sandstone and siltstone. These beds commonly exhibit ripple laminations, symmetric ripplemarks, interference ripplemarks, horizontal and low-angle planar laminations, and mudcrack casts. Approximately 30 feet (9 m) from the top of the member, black, finely fissile shales yield abundant *Chuaria circularis* on both the eastern and western slopes of Nankoweap Butte. The former is thought to be Walcott's type locality. These fossils have also been found in the same horizon in Sixtymile Canyon.

Walcott Member

The Walcott Member forms the topmost subdivision of the Kwagunt Formation and is much more diverse in character than those below it. It is 838 feet (255 m) thick and forms the upper part of Nankoweap Butte (Fig. 4.10). At the base

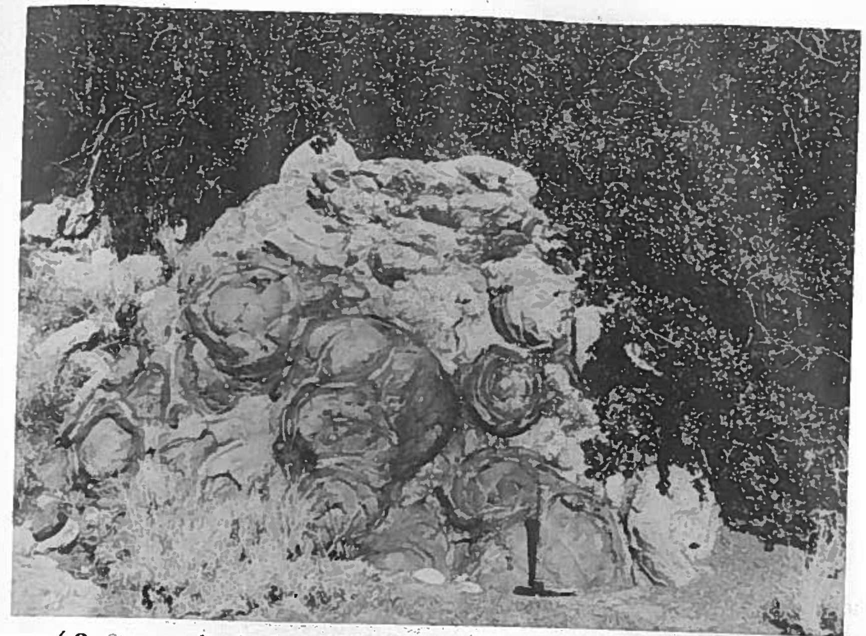

FIGURE 4.8. Stromatolite bioherm from the Awatubi Member, Kwagunt Canyon. Hammer for scale.

is a remarkable "flaky dolomite" that ranges in thickness from 12 to 31.5 feet (3.75 to 9.6 m) (Fig. 4.11; Cook 1991). This unit is composed of oolitic and intraclastic dolomite; folded and broken, crinkly, silicified and dolomitic laminations (stromatolite horizon); and wavy to horizontally laminated, silty intraclastic dolomicrite (Cook 1991).

Above the flaky dolomite are *Chuaria*-bearing black shales, silicified oolite and pisolite beds, and three distinctive carbonate units. The oolite and pisolite beds generally range in thickness from 6 inches to 1 foot (15 to 30 cm) and are found interbeddded with the black shale (Cook 1991). An exception is the white silicified oolitic bed, formerly classified as a pisolite by Ford and Breed (1973a), which is 4–5 feet (1.2–1.5 m) thick (Cook 1991). The pisolite beds are jet black and contain ooids and pisoids that are completely replaced by chert. The outer surfaces of some pisoliths contain a mat of algal filaments of at least two types, as well as spheroidal bodies (Schopf et al. 1973). Toward the top of the member are three dolomite units, the lower two are known as the "dolomite couplet," and the upper unit is known as the "karsted dolomite" (Cook 1991).

The lower dolomite couplet is 8.5 to 22 feet (3.5 to 7 m) thick and is composed of wavy to horizontally laminated dolomite (some contorted), trough–cross-bedded oolitic and intraclastic dolomite, a crinkly laminated, "cornflaky" algal stromatolite, and pink medium-grained quartz sandstone (Cook 1991). The upper dolomite couplet is 31–38 feet (9–12 m) thick and is a massive micrite to dolomicrite with rare mud chips, interbeds of black shale, and carbonate breccia zones (Cook 1991). The karsted dolomite is 40 feet (12 m) thick and is only present in Sixtymile Canyon. This unit comprises crystalline dolomite with vugs, cavities, dissolution features, and brecciated dolomite and sandstone clasts in some cavities (Cook 1991).

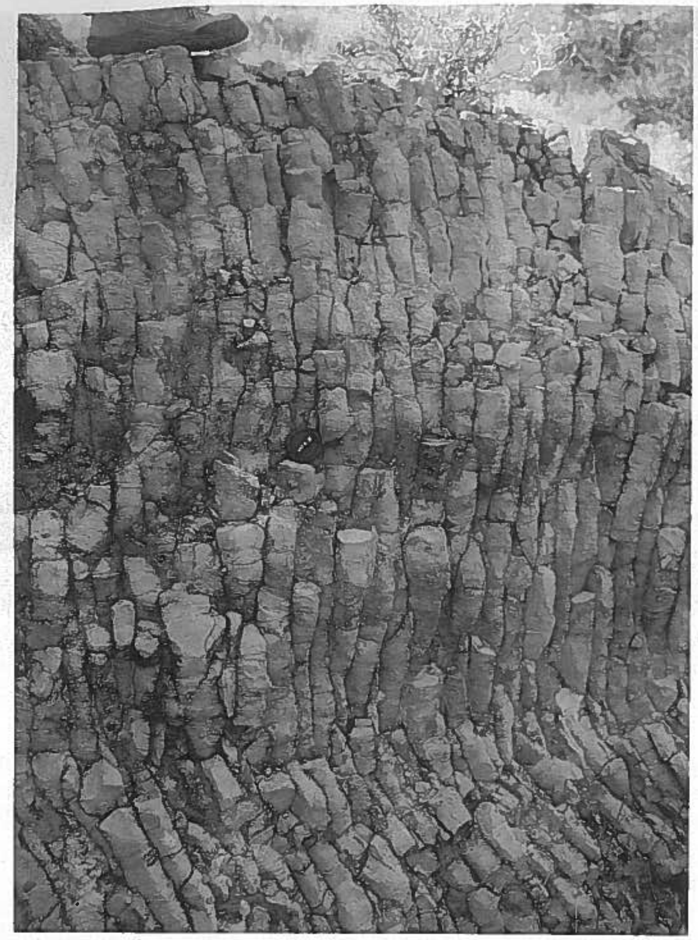

Figure 4.9. Internal columnar structure of a bioherm from the Awatubi Member, Carbon Canyon. Lens cap for scale. (Photo by Mike Timmons.)

Paleontology of the Chuar Group

During the past century, a number of fossils or possible fossils have been described from the Precambrian rocks of the Grand Canyon. In recent years, most of these either have been reassigned to biological groups other than those in which they first were placed, or their biologic affinities have been questioned. With the increasing interest in the early stages of life on our planet, several reviews of Precambrian life have been published, and these have presented conflicting interpretations of some of the fossils of the Grand Canyon. A review is presented here of the state of knowledge concerning fossils from the Chuar Group.

Chuaria circularis This small, disc-like, carbonaceous fossil first was found by Walcott during his pioneer work on the Grand Canyon Series (Precambrian) in

Figure 4.10. Nankoweap Butte, showing the upper Awatubi Member, the overlying Walcott Member with its ledges of pisolitic chert and flaky dolomite, capped by the Sixtymile Formation. Seen looking northeast toward the lower part of Nankoweap Canyon. The axis of the Chuar Syncline runs through Nankoweap Butte, and the east limb of the Chuar Syncline is in the upper right of center. The Butte fault is in the shadows in the upper part of the photo and runs right to left. (Photograph by John Shelton.)

1882–1883. It was formally named in 1899 and described as a primitive brachiopod allied with *Orbiculoidea* or *Discina*. As such, it was the first Precambrian brachiopod to be described, and it is somewhat surprising that it was largely overlooked by subsequent writers. The exceptions include Wenz (1938), who assigned *Chuaria* to the gastropods without giving very clear reason for so doing; Hantzschel (1962), who regarded *Chuaria* as inorganic; and both Glaessner (1966, 1984) and Cloud (1968), who tentatively regarded it as algal in nature, without going into details. A full description was given by Ford and Breed (Ford et al. 1972b; Ford and Breed 1973a).

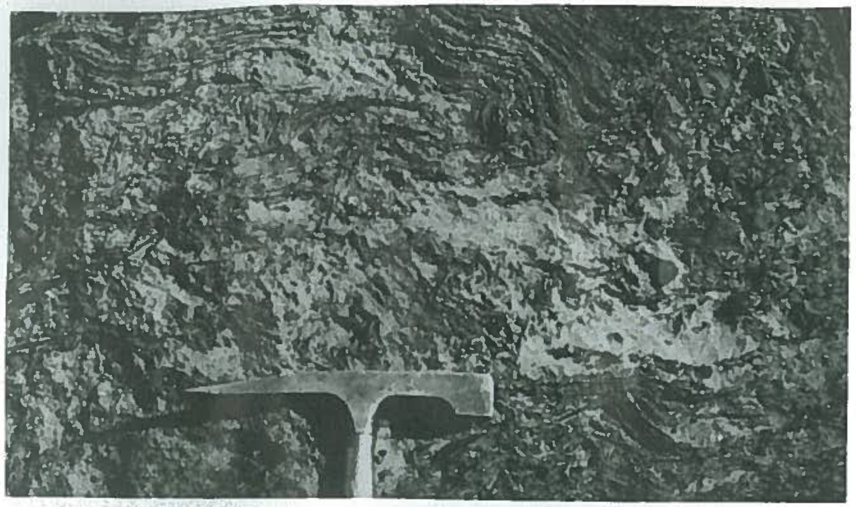

FIGURE 4.11. The "Flaky dolomite" bed in the Walcott Member, Nankoweap Butte. Hammer for scale.

Others, including Vidal and Ford (1985), reexamined the problem and showed that this smooth, organic-walled, spheroidal microfossil was best classified with the Acritarcha (algal cysts of unknown affinity). It has been found at many horizons in the Chuar Group, where it is associated with assemblages of other late Riphean to early Vendian acritarchs.

The size range of *Chuaria circularis* revealed by palynological techniques was 70–712 μm; taken together with the megascopic specimens, this gives a total range of 70 μm to about 5 mm. The large forms are most common at what is presumed to be Walcott's-type locality, high in the Awatubi Member about 30 feet (9 m) below a prominent ledge of pisolitic chert on the eastern flank of Nankoweap Butte.

The large forms of *C. circularis* are crushed, carbonized, (originally) spheroidal objects commonly 2–3 mm in diameter, lying either alone or in small clusters (they are never seen to lie on top of another), indicating their globular shape at the time of deposition (Fig. 4.12). Specimens extracted from the shale with acids are hollow with a narrow marginal thickening. The central area is wrinkled like a crushed prune. No apertures have been seen and no ornamentation is recognized. It clearly is not a brachiopod or gastropod. Other tentative suggestions included the possibilities that *Chuaria* might be a trilobite egg or a noncalcareous foraminiferan; however, no evidence to support such suggestions can be found. The size range down to 70 μm, together with the composition and features noted above, makes it clear that *Chuaria* is an acritarch.

Microfossils Samples of the black shale beds throughout the Chuar Group have yielded organic detritus, and most horizons have yielded acritarchs. These are spheromorphic, cyst-like objects presumed to be the resting stage of some form of algae. Among the genera recovered are *Stictosphaeridium, Trachysphaeridium, Leiosphaeridia, Kildinosphaera, Cymatiosphaeroidies, Tasmanites,* and small *Chuaria.* A new form, *Vandalosphaeridium walcotti,* also was discovered.

FIGURE 4.12. *Chuaria circularis*—a mega-acritarch from the Walcott Member. Each carbonized disc is about 2 mm in diameter.

Assemblages with either megascopic *Chuaria* or the above microplankton have been recovered from many late Precambrian strata around the world, and they can be used for stratigraphic correlation. On these micropaleontological grounds, the Chuar Group can be correlated with sequences in Utah, northwest Canada, Greenland, Scandinavia, Svalbard, Russia, Iran, India, China, and Australia. In both Scandinavia and Russia, the sequences are earlier than the Vendian glacial deposits and are placed in the latest Riphean and earliest Vendian divisions of the Proterozoic—that is, around 800–700 million years ago. This is slightly younger that the date of ca. 823 million years deducted for the Grand Canyon disturbance by Elston and McKee (1982), but dating of Precambrian sedimentary deposits still has many uncertainties.

Vase-Shaped Microfossils A shale horizon in the Walcott Member, about 3 feet (1 m) below the prominent pisolitic chert, has yielded an assemblage of vase-shaped, organic-walled microfossils (from 32–170 μm in length) that Bloeser (1985) has named *Melanocyrillium.* Up to 10,000 specimens per cubic centimeter of shale have been found. The teardrop-shaped, vase-like, noncolonial microfossils have an ornamented aperture at the narrow end, and three species have been erected on the basis of ornamental detail. They are of uncertain biological affinity and are not chitinozoa, as first thought (Bloeser et al. 1977). Comparable forms have been found in rocks of late Precambrian age in Sweden, Greenland, and Brazil.

An assemblage of well-preserved filamentous and spheroidal plant microfossils have been found in a cherty pisolite bed of the Walcott Member on the slopes of Nankoweap Butte in eastern Grand Canyon (Schopf et al. 1973). The spheroidal forms probably are related to coccoid blue-green algae; the filaments bear a striking resemblance to *Eomycetopsis,* a thallophyte from bedded cherts in the Bitter Springs Formation of central Australia. Preliminary studies have revealed similar filamentous forms at other horizons of the Chuar Group.

Stromatolites Stromatolites are common in the Precambrian of the Chuar Group, where they first were noted as "concretionary limestones" by Walcott (1894). They were given the now obsolete name *Cryptozoon occidentale* by Dawson (1897), though their nature was not then understood. Ford and Breed (Ford et al. 1969, 1972a; Ford and Breed 1973a) have recorded three well-developed stromatolite horizons in the Chuar Group, as well as a large number of other carbonate beds with crinkly laminations of probable algal origin. They generally have shapes such as gentle undulations or low domes with confluent laminae and only rarely show the more distinctive columnar forms that Russian workers regard as diagnostic fossils. Forms regarded as *Inzeria, Baicalia,* aff. *B. rara,* and *Boxonia* have been recognized by comparing the external form with those illustrated by Cloud and Semikhatov (1969), though diagenetic alteration has obscured internal detail and made more specific identification impossible. Recognition of these genera allows correlation of the Chuar Group with the Russian Upper Riphean—that is, late Precambrian.

The three beds of stromatolites in the Chuar Group are well-exposed. The lowest horizon forms a prominent ledge and waterfall about a mile up Chuar Canyon from its mouth on the north side. It also is seen as a prominent ledge in the middle reaches of Basalt Canyon on the north side. The middle horizon is a single bed of limestone and dolomite within the upper part of Carbon Canyon Member—outcropping low on the flanks of Carbon Butte and also about 100 feet (33 m) above the creek in Nankoweap Canyon at the foot of Nankoweap Butte. The highest forms a bed of stromatolite "reefs," or bioherms, cropping out around the foot of Nankoweap Butte in Nankoweap and Kwagunt canyons (Figs. 4.9 and 4.10) and just above the prominent sandstone bench of Carbon Butte.

The paleoenvironment of stromatolites in the Chuar Group, judged by associated sedimentary rocks, is one of relatively low energy, shallow waters. Ripplemarks and mudcracks are common, suggesting a gentle current and intermittent desiccation. Intraclastic breccias associated with the stromatolites suggests periods of relatively higher energy current associated with possible dessication. The three major stromatolites and many other crinkly laminated carbonate beds are interbedded with black shales and maroon mudstones. These relationships suggest that algal growth may have been the most successful during times of less siliciclastic input, and when more siliciclastics came into the system, the water became too clouded for algae to survive.

Chuar Group Paleoenvironments

In the first and only study of the entire Chuar Group, Ford and Breed (Ford et al. 1972b; Ford and Breed 1973a) stated that the presence of megaplankton (*Chuaria*) and abundant microplankton, stromatolites, and filamentous algae indicate at least partial marine influences, and that sedimentologic evidence suggests a marine shoreline setting. Subsequent studies have resulted in a range of depositional interpretations—from deep and shallow marine, to alluvial and lacustrine (Horodyski 1986; Reynolds and Elston 1986; Cook 1991). Ongoing work on the sedimentology and stratigraphy of the Chuar Group is in general agreement with Ford and Breed's original interpretation (Dehler 1998). Knowing whether the Chuar Group sediments were deposited in a lake or in a coastal setting can reveal information about the paleogeography of the region at this

time in the late Precambrian. If the Chuar Group is marine in origin, then it suggests that there was a continental margin in what is now northern Arizona during late Precambrian time. If the Chuar Group is lacustrine in origin, the Chuar basin was inboard of the continental edge.

Galeros Formation The basal Tanner dolomite is probably very similar to other overlying carbonate units, other than the fact that it has been severely diagenetically altered. The faintly laminated Tanner dolomite is likely similar to laminated dolomites in other parts of the Chuar Group that suggest a shallow subtidal or intertidal environment. The overlying black shales, sometimes *Chuaria*-bearing, represent a deeper water environment that occasionally received silts and sands during storms. Reynolds and Elston (1986) interpreted the Tanner Member to represent a sediment-starved basin, rich in organic material. The overlying Jupiter Member contains the basal *Inzeria* bed that also represents shallow subtidal to intertidal deposition. The overlying shales in the Jupiter are variegated and commonly display mudcracks, raindrop impressions, and salt crystal casts. These shales probably represent an intertidal to supratidal environment. Reynolds and Elston (1986) interpreted the Jupiter Member to represent a coastal or alluvial plain. The Carbon Canyon Member has a mixture of characteristics similar to the Tanner and Jupiter Members, and it represents fluctuating deeper subtidal to intertidal to supratidal environments. Mudcracked wave ripplemarks suggest fluctuations in currents that were tide- and wave-affected. Reynolds and Elston (1986) interpreted this member to represent a mixed coastal or paludal swamp. The only difference between the Duppa Member and the underlying Carbon Canyon Member is the lack of carbonate and sandstone beds. Therefore, the Duppa Member may also represent a deeper subtidal to supratidal environment. The Duppa Member is interpreted by Reynolds and Elston (1986) to represent an alluvial plain.

Kwagunt Formation The sedimentary structures in the Carbon Butte Member, such as mudcrack casts, wave ripplemarks, interference ripplemarkss, bidirectional trough crossbeds, and possibly epsilon crossbeds, indicate a tide- and wave-affected shoreline. However, some of these sedimentary structures could represent fluvial conditions. The Carbon Butte Member sandstones represent a dominance of relatively coarse, clastic sediments indicating an increase in sediment supply, a shift in environments (closer to the shoreline), or a significant change in energy conditions. The basal Awatubi Member bioherm suggests shallow subtidal to intertidal conditions. The upper Awatabi Member has symmetric ripplemarked sandstones with mudcrack casts similar to those in the Carbon Butte Member, and the uppermost Awatubi Member has *Chuaria*-bearing black shales. The Awatubi Member represents an intertidal to shallow subtidal environment that deepened through time. The Walcott Member has been interpreted by Cook (1991) to represent a carbonate ramp: The black shales are the deepest water environment, the oolite and pisolite beds are the shallow subtidal environments, and the carbonate beds represent shallow subtidal, intertidal, and supratidal environments.

The sedimentary rocks of the Chuar Group suggest that, in general, they were deposited in a quiet, nonturbulent embayment on a tide- and wave-affected marine platform fringing a continental mass. The outcrops are too limited, however, to indicate the direction of the coastline relative to this embayment, although ongoing basin analysis will result in more data regarding this problem (Dehler 1998; Timmons et al. 1998).

SIXTYMILE FORMATION

In contrast to the underlying Chuar Group, the Sixtymile Formation is composed largely of breccias and sandstones, with subordinate siltstones and mudstones. Slump folds are common, as are large boulders interpreted to be landslide blocks derived from the east limb of the Chuar syncline. About 200 feet (60 m) thick, the Sixtymile Formation caps Nankoweap Butte, where it has been misidentified in the past as basal Cambrian Tapeats Sandstone (Van Gundy 1951). However, magnificent exposures of this unusual formation are present in the little-visited and remote Awatubi and Sixtymile Canyons. The formation is named from the latter canyon, and its northwestern cliffs are well worthy of detailed study (Fig. 4.13). The Sixtymile Formation is beveled by the Great Unconformity and overlain by the Cambrian Tapeats Sandstone in both Sixtymile and Awatubi Canyons.

Elston (1979) divided the Sixtymile Formation in Sixtymile Canyon into three unnamed members. He also provided a comprehensive and detailed description of the rock types, together with an analysis of the events resulting in this unusual formation. The lower member, only present in Sixtymile Canyon, is as thick as 90 feet (27 m) and consists of coarse breccias full of many different types of clasts, including chert and dolomite derived from the underlying Chuar Group (Fig. 4.14). Among these are large blocks, the largest one measuring about 33 × 20 × 20 feet (10 × 6 × 6 m). A dolomitic limestone unit, 26 × 130 feet (8 × 40 m), was interpreted by Elston (1979) to be one of these landslide blocks. Cook (1991) studied this limestone unit in detail and concluded that it was *in situ*, and he named it the karsted upper dolomite (see Walcott Member discussion). Also in the lower member are red sandstones, some of which are thinly bedded and laminated.

FIGURE 4–13. Sixtymile Canyon: The Sixtymile Formation is the dark part of the cliffs on the right.

FIGURE 4–14. Intraformational breccia in the Sixtymile Formation, Nankoweap Butte. Hammer for scale.

The middle member is a thinly bedded and laminated, fine-grained quartzitic sandstone and siltstone 80 feet (25 m) thick, with common chert lenses, parting lineations, and massive white thin beds that may be tuffaceous. Slump fold axes in this member parallel the axis of the Chuar syncline.

The upper member is 40 feet (12 m) thick and comprises locally derived, fine- to coarse-grained sandstone, siltstone, conglomerate, and breccia. This member fills in paleotopography cut into the middle member, and the clasts are all derived from the middle member. These cut-and-fill features are as deep as 16 feet (5 m), and channel axes have northwest–southeast trends (parallel to the Butte fault). Other sedimentary features include contorted bedding, adhesion ripples, and small cut-and-fill structures filled with trough–cross-bedded sandstone showing paleoflow to the east–northeast.

Sixtymile Formation Paleoenvironments

A drastic change in environment is recorded in the Sixtymile Formation, yielding very coarse clastic sediments, probably under largely terrestrial(?) and syntectonic conditions. Landsliding and slumping in early Sixtymile time was followed by colluvial(?) and alluvial(?) sedimentation in later Sixtymile time, indicating that there was significant local relief throughout Sixtymile deposition.

Large boulders, breccias, and contorted bedding characterize the lower member and indicate mass-wasting processes. Elston (1979) interpreted these mass-wasting events to be controlled by folding of the Chuar syncline and uplift along the Butte fault. The lower member red sandstones were deposited in subaerial environments at or near sea level, and the red sandy siltstone represents mud-flat deposition Elston (1979). The thinly bedded siltstones and sandstones of the middle member are probably representative of a low-energy fluvial environment such as a floodplain. Elston (1979) suggested that the middle member was deposited in a localized lake in the synclinal axis. Paleovalleys cut into the middle member and filled with coarse breccias and sandstones of locally derived, middle member clasts potentially indicate exhumation and deposition by debris flow, and fluvial processes. Elston (1979) interpreted these upper member deposits to be of fluvial origin that were eroding off the high-standing footwall of the Butte fault. Ongoing work on the Sixtymile Formation will contribute to our understanding of the depositional environment(s) of this enigmatic unit (Karl Karlstrom, personal communication, 1998).

CORRELATION OF THE NANKOWEAP FM, CHUAR GROUP, AND SIXTYMILE FM

Younger Precambrian strata are well- and widely exposed across central and southern Arizona, where they are known as the Apache Group. The Apache Group is correlated with the Unkar Group based on (a) lithologic similarities (Troy Quartzite with the Shinumo Quartzite, respectively) and (b) diabase sills of similar ages which cross-cut both groups (Link et al. 1993). There is no equivalent to the Nankoweap Formation, Chuar Group, or Sixtymile Formation in Arizona, yet comparisons can be made with strata in other parts of western North America.

In the Death Valley region of California, a thick infra-Cambrian sequence lies beneath the equivalent of the Tapeats Sandstone and rests on the Pahrump Group. This lower part of the Pahrump Group (Crystal Spring formation) is correlated with Unkar Group and the Apache Group based on similar texture and chemistry of crosscutting diabase sills (Link et al. 1993). The Chuar Group likely correlates with the Beck Springs Dolomite of the upper Pahrump Group, based on lithologic (shales, stromatolites, carbonate rocks), paleoenvironmental (fluvial-marine), and acritarch assemblage similarities. Unconformably above the Beck Springs Dolomite lies the Kingston Peak Formation, which contains tillite (Link et al. 1993). No glacial deposits have been identified in the Chuar Group, and based on microfossil global correlations, the Chuar Group predates the late Precambrian glacial events (Vidal and Ford 1985). The Red Pine Shale of the Uinta Mountain Group in Utah has yielded microfossils almost identical with those in the Chuar Group. Several geologists have proposed a correlation between these two groups (Hoffmann 1977; Vidal and Ford 1985).

The Belt Supergroup, predominantly located in Montana, was previously thought to be coeval with the Apache and Unkar Groups. Recent U-Pb ages of about 1400 Ma or older for the Belt Supergroup make this correlation unlikely (Aleinikoff 1996). The northern Canadian Windermere Supergroup unconformably overlies a dike of about 778 Ma and is cross-cut by a sill of about 723 Ma (Link et al. 1993). The presence of *Chuaria* in the Hector Formation of the Windermere Group provides support for a correlation with the Chuar Group (Ford and Breed 1973a, 1973b). The Little Dal Group of the Mackenzie Mountain Supergroup in northwest Canada is cross-cut by a diabase that is about 778 Ma, and it has yielded microplankton fossils similar to those in the Chuar Group (Hoffmann and Aitken 1979; Link et al. 1993). As noted earlier, the age of the Grand Canyon disturbance has been dated at about 823 Ma, which correlates reasonably well with the structural disturbance that terminated the Little Dal Group in northwestern Canada at about 778 Ma.

Support for these correlations comes from paleomagnetic studies of several of the above-mentioned sequences. Comparisons of paleomagnetic poles from Keweenawan igneous rocks and the Cardenas Lava are consistent with an age of about 1100 Ma (Link et al. 1993 and references therein). Limited paleomagnetic data (Elston and Grommé 1974; Elston and McKee 1982; Elston 1989a) are consistent with an age of less than about 1050 Ma for deposition of the upper member of the Nankoweap Formation. Paleomagnetic results from sills in the Little Dal Group yield paleomagnetic poles compatible with those for the Chuar Group strata (Don Elston, personal communication, 1997). There are strong indications that many of the Supergroup and Sixtymile Formation rocks contain primary magnetizations, but detailed and necessary field-based tests remain to be conducted to test this hypothesis (John Geissman, personal communication, 1998).

Although the ages of some of these sequences cannot yet be ascertained by isotopic dating, the microfossil evidence provided by acritarch assemblages strongly suggests intercontinental correlation with the Visingso Group of Sweden and, thus, with the latest Riphean of Russia. Isotopic age determinations in Sweden and Svalbard indicate the age of these beds to be around 800–700 million years (Vidal and Ford 1985). Applying a correlation by long-distance lithologic comparison, micropaleontologic assemblages, paleomagnetism, and very limited isotopic dating, it seems that the Nankoweap Formation, Chuar Group, and Sixtymile Formation were deposited within the period ~1000–700 million years ago.

SUMMARY

The Chuar Group represents deposition associated with a fluctuating shoreline in a tectonically active marine basin. Preliminary paleocurrent data on symmetric ripplemark crests show trends that parallel the Butte fault, suggesting the shoreline orientation was also north–south. Black shales throughout the Chuar Group, many *Chuaria*-bearing, represent the deepest subtidal environment. Maroon mudstones, pisolite and oolite beds, and carbonate beds (including stromatolites) were deposited in subtidal environments. Many sediments were exposed intermittently or for prolonged periods in intertidal and supratidal zones. Sandstones represent wave- and tide-affected shallow subtidal and estuarine(?) conditions.

The late Precambrian paleontological record from the Chuar Group consists of a variety of microfossils from filamentous algal sheets, coccoid algae, acritarchs (including the megascopic *Chuaria*), vase-shaped microfossils of unknown affinity, and stromatolites. The lobate markings from the Nankoweap Formation can be regarded as a dubio-fossil and may not be of organic origin at all.

The Nankoweap Formation has no known correlative sequences. The Chuar Group and Sixtymile Formation(?) are likely correlative with the Beck Springs Dolomite in Death Valley, the Red Pine Shale in Utah, and the Little Dal Group or the Windermere Group in northern Canada, based on lithologic, paleontologic, and geochronologic data. Global correlations can be made with 700- to 800-Ma successions in Svalbard and Sweden using archritarch assemblages.