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Tonto Group

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INTRODUCTION

The Cambrian of the Grand Canyon is, without question, one of the classic sequences of sedimentary rocks exposed in North America. These strata crop out along a prominent, essentially horizontal surface known as the Tonto Platform in the central part of the canyon and near the banks of the Colorado River in western areas of the canyon. The surface of the Tonto Platform roughly coincides with the top of the oldest Cambrian formation, the Tapeats Sandstone. Above the Tapeats, a series of small cliffs are separated by thicker intervals of slopes composed of finer-grained deposits of the Bright Angel Shale. These, in turn, are overlain by cliffs of resistant carbonate of the Muav Limestone, the youngest Cambrian formation of the Tonto Group.

This three-part system of sandstone, mudstone, and limestone is well known to most geologists and to a great number of Grand Canyon hikers and tourists. Despite this, research into the origin of these rocks has not kept pace with the developments in the last twenty years concerning the dynamics of nearshore and shelf depositional systems. The classic work of McKee and Resser (1945) endures as the most comprehensive study of the Cambrian system in the Grand Canyon.

To date, only a few studies have attempted to document carefully the lateral and vertical facies associations recorded in these strata (McKee and Resser 1945; Wanless 1973; Hereford 1977; Martin 1985; Middleton 1988). A major objective of this chapter is to present new data to enable recreation of the depositional systems that existed during the Cambrian in northern Arizona. Both sedimentologic and ichnologic data will be used to examine the depositional history of the Tonto Group.

Cambrian deposits in the Grand Canyon and throughout the Rocky Mountains long have been cited as representing a classic transgressive sequence of sandstone, mudstone, and limestone that accumulated on the slowly subsiding Cordilleran miogeocline and adjacent craton (McKee and Resser 1945; Lochman-Balk 1970, 1971; Stewart 1972; Stewart and Suczek 1977). During Early and Middle Cambrian time, a north–south trending strandline migrated progressively eastward across the craton. This shoreline was characterized by numerous embayments and offshore islands that affected sedimentation in nearshore areas. Shoreline migration for the most part was eastward, resulting in deposition of coarse clastics in shallow water areas to the east and finer clastics and carbonates in more offshore areas to the west. As will be discussed, numerous regressive phases interrupted this overall eastward transgression resulting in complicated facies interactions.

Cambrian strata in Arizona and along the entire western margin of North America thicken to the west, presumably reflecting more subsidence in the miogeoclinal or offshore shelf areas. Continued subsidence and/or sea-level rise, interrupted by a number of sea-level retreats or regressions, resulted in the complete submergence of the western cratonic margin by the Late Cambrian.

REGIONAL STRATIGRAPHIC RELATIONSHIPS

The Tonto Group comprises three formations (Fig. 6.1) that are, in ascending order: Tapeats Sandstone, Bright Angel Shale, and the Muav Limestone. The term "Tonto Group" was used first by G.K. Gilbert (1874) to describe this sandstone–shale–limestone sequence, though he considered these rocks to be of Silurian age. Subsequent stratigraphic and paleontologic work by Walcott (1890) established a Cambrian age for the Tonto Group, and Noble introduced the now-accepted formation names in 1914 during his mapping of the Shinumo Quadrangle in the Grand Canyon.

Strata of the Tonto Group also crop out along the Grand Wash Cliffs in western Arizona and to the east in the Juniper Mountains and the Black Hills in west-central Arizona. In these areas, the Tapeats Sandstone is overlain disconformably by the Devonian Martin Formation or the Chino Valley Formation, the age of which is uncertain (Hereford 1975). Presumably, the Bright Angle Shale and the Muav Limestone were removed during extensive pre-Devonian erosion. In central Arizona, scattered outcrops of the Tapeats Sandstone occur along the East Verde River and in the Sierra Ancha Range north of Young, Arizona. Tonto Group equivalents in southeastern Arizona include the Bolsa Quartzite and part of the overlying Abrigo Formation (Hayes and Cone 1975; Middleton 1988).

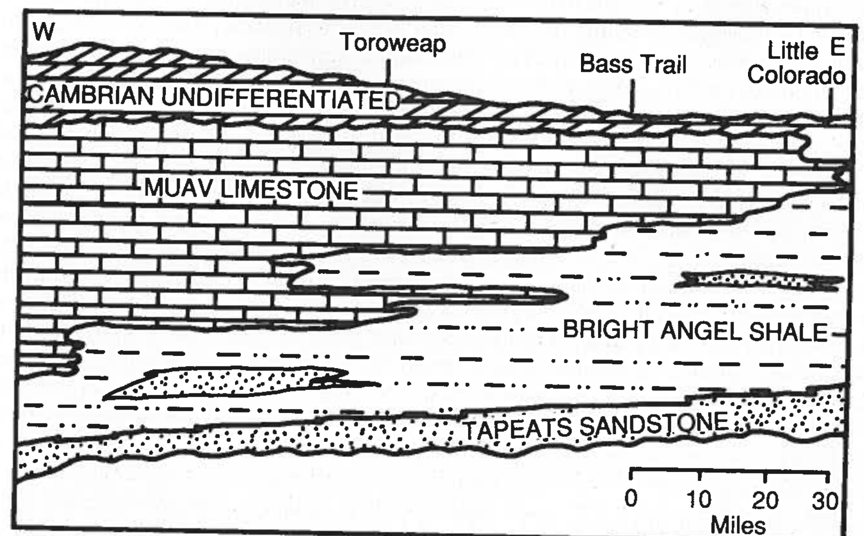

Figure 6.1. West-to-east cross section of Tonto Group in Grand Canyon illustrating stratigraphic relationships and eastward younging of the group. (From McKee and Resser 1945.)

Cambrian strata overlie a variety of Precambrian lithologies throughout the Grand Canyon. In the eastern part, the Tonto Group rests on tilted beds of the 1.4- to 1.1-billion-year-old Grand Canyon Supergroup, whereas in the western areas the Tonto Group nonconformably overlies older Precambrian (circa 1.7- to 1.6-billion-year-old) rocks of the Vishnu Group and other metamorphic units. This major unconformity between Precambrian and Tonto Group rocks, which has been recognized for a long time, obviously represents a considerable period of time during which the region was subjected to episodes of mountain building and extensive erosion. Walcott (1910) applied the name "Lipalian interval" to the period of time represented by this unconformity. Although dating of the igneous rocks within younger Precambrian strata establishes a minimum age, it is impossible to measure precisely the time from cessation of uplift to production of the nearly flat erosion surface onto which sediments of the Tapeats Sandstone were deposited. Clearly, we are dealing with a long period of time.

The surface upon which the Tonto Group accumulated was quite irregular. It was characterized by a rolling topography of resistant bedrock "hills" and lowlands. The Precambrian bedrock was weathered extensively in places and eroded during prolonged periods of subaerial exposure. Walcott (1880) and Noble (1914) were among the first to recognize that the Precambrian surface represented paleotopography and that sedimentation patterns were influenced by the relief and lithologies of these "hills." Other workers likewise have documented the influence of Precambrian topography on Cambrian sedimentation in other areas of the Rocky Mountains and in the midcontinent. There are numerous places in the canyon where the Tapeats Sandstone thins across or pinches out against these Precambrian highs. Where the Tapeats pinches out, the Bright Angel Shale overlies the Precambrian surface. The influence of these Precambrian highs will be discussed as the depositional environments of the Tonto Group are reconstructed.

A highly weathered horizon occurs on top of the Precambrian surface in several places in the canyon. The only effort to understand the genesis of this potentially very significant unit is that of Sharp (1940). His study suggested that extensive chemical weathering of Precambrian rocks occurred prior to deposition of Cambrian sediments. In places, this highly weathered surface, or regolith, is up to 50 feet (15.3 m) thick but generally is less than 10 feet (3.1 m) thick. Sharp speculated that where the Tapeats rests on unaltered basement, the regolith probably was removed by wave erosion associated with the initial Cambrian transgression. Sharp (1940) and McKee and Resser (1945) have suggested that the presence of such a thick, weathered horizon indicates that dominantly humid conditions existed during the early Paleozoic prior to deposition of the Tapeats Sandstone. Unfortunately, there have been no petrologic and geochemical studies that could substantiate this hypothesis. Considering that the time represented by the unconformity was likely several hundred million years, that the climate could have changed numerous times during this period, that this horizon was buried and exhumed numerous times prior to deposition of the Tapeats Sandstone, and that in the absence of terrestrial vegetation, weathering processes in soils would have been different (Basu 1981), a humid climate interpretation is quite tenuous. Obviously, considerable research needs to be done in this area.

STRATIGRAPHY OF THE TONTO GROUP

Tapeats Sandstone

The Tapeats Sandstone was named for exposure along Tapeats Creek in the western part of Grand Canyon National Park. For the most part, the formation

is a medium- to coarse-grained feldspar and quartz-rich sandstone with granule and pebble-size, quartz-rich conglomerate present locally near the base. The percentage of feldspar is highest at the base and decreases upwards through the formation. The composition of the basal Tapeats reflects to varying degrees the mineralogy of the underlying Precambrian rocks. To date, however, there have been no petrologic studies of the Tapeats aimed at documenting the changes in mineralogy with respect to facies changes or evaluating the influence of basement lithology and paleotopography on the composition of the Tapeats.

The formation can be divided into two generalized packages. The majority of the Tapeats crops out as a cliff consisting of beds typically less than 3 feet (1 m) thick. Sedimentary structures include planar and trough cross-stratification and crudely developed horizontal stratification. Both the scale of the bedding and the cross-stratification decrease upwards. Overlying the main cliff is a thinner zone of interbedded fine- to medium-grained sandstone and mudstone. Stratification is of a smaller scale in these beds and is largely trough and ripple cross-stratification and horizontal stratification.

The significance of the upper unit is that it marks a major facies transition into the upper Bright Angel Shale (Fig. 6.2). An increase in fine-grained material indicates a reduction in the bedload to suspension load ratio. The concomitant changes in bedding thickness and scale of sedimentary structures are consistent with the above interpretation. The contact between the two formations, therefore, is arbitrary and probably should be placed at the top of the thickest sandstone bed within the transitional interval.

The Tapeats varies considerably in thickness throughout the Grand Canyon and also in areas to the south and west. A thickness of 393 feet (120 m) was reported by Noble (1922) along the Bass Trail. It is possible that this represents a maximum thickness in the canyon. Typically, the formation is between 100

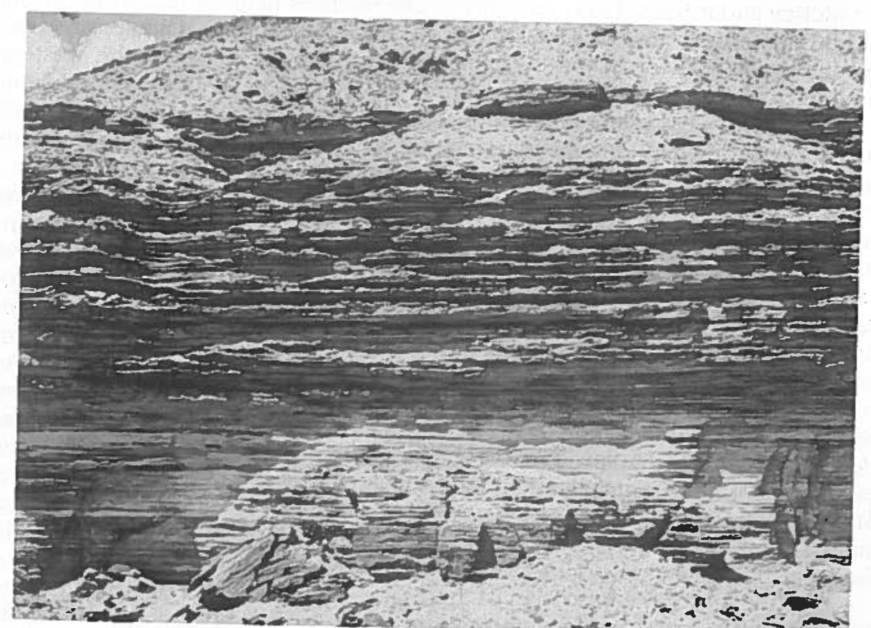

FIGURE 6.2. Thin-bedded, horizontally and ripple-laminated upper portion of the Tapeats Sandstone. Overlying slope marks gradational contact with the Bright Angel Shale.

and 325 feet (30 and 100 m) thick. The thickness of the Tapeats clearly is controlled by relief of the underlying Precambrian surface. As previously stated, there are areas where the Tapeats thins across and/or pinches out against these Precambrian highs.

Except for trace fossils, which in places are quite common, body fossils are rare and only occur within the transition zone (McKee and Resser 1945). However, these fossils establish a late Early Cambrian age for the upper parts of the Tapeats Sandstone in the Grand Wash Cliffs in the western part of the canyon and an early Middle Cambrian age for the formation in the eastern canyon. These ages are based on trilobite assemblages (*Olenellus-Antagmus*) in the overlying Bright Angel Shale. This diachroneity is a reflection of the west-to-east sense of strandline migration.

Bright Angel Shale

The Bright Angel Shale is perhaps the least-studied formation in the Grand Canyon. It was named by Noble (1914) for exposures of slope-forming, interbedded, fine-grained sandstone, siltstone, and shale just above the Tonto Platform along Bright Angel Creek. Conglomerates and coarse-grained sandstones of the Bright Angel Shale contain quartz, minor amounts of potassium feldspars and sedimentary rock fragments, and glauconite. The latter is responsible for imparting the green color to many of the siltstones and sandstones. A number of the sandstones and siltstones contain a high percentage of hematitic ooids and iron oxide cements imparting a reddish brown coloration. The dominant lithology is greenish shale composed largely of illitic clay with varying amounts of chlorite and kaolinite. Inarticulate brachiopods, trilobites, and *Hyolithes* are locally abundant. Trace fossils are extremely abundant and varied.

McKee and Resser (1945) recognized one member in the Bright Angel Shale, which they termed the Flour Sack Member. This unit consists of shale, siltstone, and limestone and forms the uppermost part of the formation in the western canyon. Limestone decreases in abundance toward the east until the entire member is shale at its easternmost outcrop near Quartermaster Canyon. A number of rusty-brown dolomite tongues that occur in the Bright Angel represent carbonate extensions of the Muav Limestone (McKee and Resser 1945).

Sedimentary structures are numerous in the coarse-grained lithologies in the Bright Angel Shale throughout the Grand Canyon. These include horizontal laminations, small- to large-scale planar tabular and trough cross-stratification, and wavy and lenticular bedding (Fig. 6.3). Locally, structureless and crudely stratified conglomeratic sandstones typically overlie a scoured surface. Martin (1985) documented a number of coarsening-upward and fining-upward sequences in the Bright Angel in the central canyon.

The Bright Angel is over 450 feet (137 m) thick in the western Grand Canyon, only 270 feet (82 m) at Toroweap in the central canyon, and 325 feet (99 m) along Bright Angel Creek (McKee and Resser 1945). This variability in thickness is due to the complex intertonguing relationships with the Muav Limestone. The Bright Angel Shale thins toward the south and is only a few feet thick in the Juniper Mountains north of Prescott, Arizona. South and east of the Black Hills, the formation is absent—presumably the result of extensive erosion.

Like the rest of the Cambrian of the Grand Canyon, the Bright Angel crosses time lines, becoming younger toward the east. In the western part of Grand Canyon, the base of the formation lies below the *Olenellus-Antagmus* assemblage zone. It is, therefore, late Early Cambrian, whereas in the eastern part of

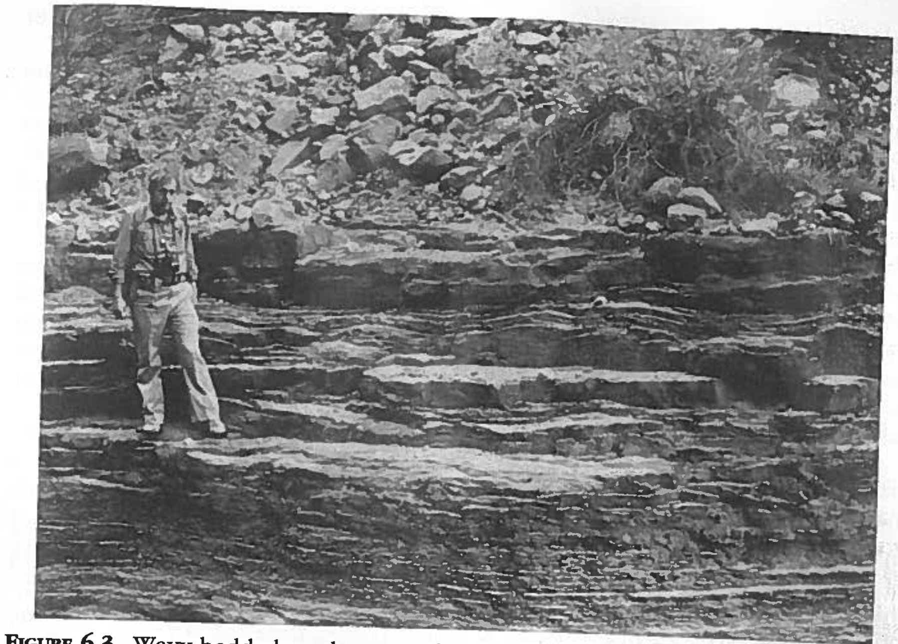

Figure 6.3. Wavy-bedded sandstone in the Bright Angel Shale along Pipe Creek. Internal structures suggest deposition by storm-enhanced currents.

the canyon, the lower third of the Bright Angel lies below the Middle Cambrian *Alokistocare-Glossopleura* assemblage zone.

Muav Limestone

The Muav Limestone is the youngest formation of the Tonto Group. It forms resistant cliffs above the Bright Angel Shale throughout the Grand Canyon. Noble (1914) named the formation for exposures in Muav Canyon. Contact with the Bright Angel Shale is gradational and characterized by complex intertonguing of the two formations. McKee and Resser (1945) defined seven members within the Muav. The members are differentiated on the basis of key marker horizons defined by fauna, intraformational conglomerate, or persistent beds of shale and/or thin-bedded limestone. The upper three members can be correlated throughout the entire canyon, whereas the lower four members are confined to areas west of Fossil Rapids.

The Muav consists of thin- to thick-bedded, commonly mottled, dolomitic, and calcareous mudstone and packstone, as well as beds of intraformational and flat-pebble conglomerate. Thin beds of micaceous shale and siltstone, minor amounts of fine-grained sandstone, and silty limestone occur at numerous horizons in the Muav, where they form small recesses and/or benches in the cliff-forming carbonate. The amount of siliciclastics increases toward the east, concomitant with a decrease in carbonates. Bedding thickness, in general, increases toward the west.

Most of the Muav comprises beds of structureless or horizontally laminated carbonate. Small-scale (less than 5 cm thick) trough, planar tabular, and low-angle cross-stratification occurs at many localities. Fenestral fabrics and desiccation cracks also are reported (Wanless 1975). Trace fossils, though not as abundant as in the Bright Angel Shale, are present throughout the canyon.

As a result of the intertonguing relationships between the Muav Limestone and the Bright Angel Shale, thickness trends within the Muav are variable. The unit as a whole thickens toward the west. McKee and Resser (1945) reported that the Muav is 827 feet (252 m) thick in the Grand Wash Cliffs near Lake Mead, 439 feet (134 m) thick at Toroweap in the central canyon, and only 136 feet (42 m) thick at the confluence of the Little Colorado and the Colorado rivers at the eastern end of the canyon.

In the western part of the canyon, the Muav lies above the *Alokistocare-Glossopleura* assemblage zone and is Middle Cambrian in age. In eastern Grand Canyon, the upper part of the Muav contains the *Bathyuriscus-Elrathina* zone and is late Middle Cambrian. The decrease in age of the Muav toward the east parallels the age trends of the Tapeats Sandstone and the Bright Angel Shale and reflects the west-to-east nature of the Cambrian transgression.

Cambrian Undifferentiated

In the western part of Grand Canyon, a thick (up to 426 feet [131 m]) sequence of dolostone overlies the Muav Limestone. McKee and Resser (1945) referred to this unit as the undifferentiated dolomites and considered it to be Upper Cambrian, though there is no paleontological evidence. Wood (1956) proposed the term "supra-Muav" for this unit, and Brathovde (1986) has suggested that these dolostones be named the Grand Wash Dolomite because the best exposures occur along the Grand Wash Cliffs in Western Arizona.

McKee and Resser (1945) recognized three lithofacies: white-to-buff massive dolomite; white-to-yellow, very fine-grained, thick-bedded dolomite; and gray, fine-grained, thick-bedded dolomite. Brathovde (1986) reported thick beds of oolitic grainstones and stromatolites that are interbedded with the fine-grained dolostones. Sedimentary structures include wavy and asymmetric ripple laminations and small-scale cross-stratification. Horizontal burrows and tracks are the dominant trace fossils.

PALEONTOLOGY

Since the early work of Resser (1946), there has been virtually no work done on the taxonomy and biostratigraphy of Cambrian strata in the Grand Canyon. Therefore, the systematics of the invertebrate fauna remain the same and will be reviewed only briefly here. Fossils have been described from the transition interval of the Tapeats Sandstone and from the Bright Angel Shale and Muav Limestone. There have been a few trace fossil studies of these strata, however, that have added to our understanding of the paleoecologic and environmental conditions during periods of deposition. These will be used in conjunction with the depositional environmental reconstructions to characterize the depositional systems.

Invertebrate Fossils

Despite the paucity of well-preserved invertebrate fossils, analysis of the fauna has provided some information concerning the paleoecology and certainly has facilitated the biostratigraphic zonation of the Tonto Group. Brachiopods and trilobites are the most common invertebrates reported from the Tonto Group, though preservation typically is poor. Fragments of sponges, primitive mollusks, echinoderms, and algae occur in the Bright Angel Shale and the Muav Limestone; however, these fossils are not very abundant. In addition to establishing the time-transgressive nature of these deposits, the biostratigraphic reconstructions have aided in documenting the numerous transgressive and regressive cycles that characterize the Cambrian of the Colorado Plateau and the Rocky Mountain regions (Lochman-Balk 1971; Aitken 1978).

Trilobites are the most abundant fossils in the Tonto Group, and common genera include *Olenellus, Antagmus, Zacanthoides, Albertella, Kootenia, Glossopleura,* and *Bolaspis*. Most specimens are poorly preserved and occur in the coarser-grained sandstones of the Bright Angel Shale and also in the mudstones of the Bright Angel Shale and the Muav Limestone. Resser (1946) reported 47 species of trilobites from the Tonto Group and suggested that these arthropods were, to some degree, facies-specific. More research needs to be done to evaluate this interesting relationship between depositional environment and trilobite distribution.

Brachiopods are locally abundant in the coarse-grained sandstones of the Bright Angel Shale. They also occur in some of the mixed siliciclastic–carbonate facies of the Muav Limestone. The most common genera in the Tonto Group are *Lingulella, Paterina,* and *Nisusia*. In general, the brachiopods tend to occur in beds containing few other invertebrate taxa.

Paleontologists have reported a number of species of primitive mollusks (Conchostraca) from the coarser-grained, hematitic sandstone of the Bright Angel Shale. The association of these fossils in coarse-grained sandstones led Resser (1946) to speculate that these mollusks occupied shallow-water habitats. Documentation of this environmental zonation, however, is unsubstantiated.

Resser (1946) reported sponge spicules from the Muav Limestone in the western Grand Canyon. These consist of thick, six-rayed spicules that Resser suggested were similar to purported sponge spicules of *Tholiastrella? hindei* that Walcott (1920) reported from the Cambrian of British Columbia. Elliott and Martin (1987) described six-rayed sclerites, which they assigned to the genus *Chancelloria*, from the Bright Angel Shale along Horn Creek in the Grand Canyon. Although Walcott (1920) considered *Chancelloria* to be a sponge (Phylum Porifera), Rigby (1976) and Elliott and Martin (1987) have questioned the assignment of this genus to this phylum and have suggested that *Chancelloria* represents a separate, yet unknown, phylum.

Algae, echinoderms, and gastropods have been described from the Tonto Group, though they certainly are the rarest taxa reported. Algal structures in the Muav Limestone consist of convex-upward laminae of calcite and/or dolomite and also small nodules composed of concentric laminations that have been termed *Girvanella* (McKee and Resser 1945). The environmental significance of the algae has yet to be established. Two well-preserved specimens of *Eocrinus* have been reported from the Bright Angel Shale. The excellent preservation of these echinoderms suggests relatively quiet water environments. Gastropods are represented by one species of *Scenella* from the Muav Limestone and several well-preserved species of *Hyolithes* from the Bright Angel Shale.

ICHNOLOGY

Trace fossils are common in all formations of the Tonto Group, particularly the Bright Angel Shale, and include a diverse array of tracks, trails, and burrows (Fig. 6.4 and Table 6.1). Despite this, the ichnofauna has been described in only a few studies (McKee 1932; McKee and Resser 1945; Seilacher 1970; Hereford 1977;

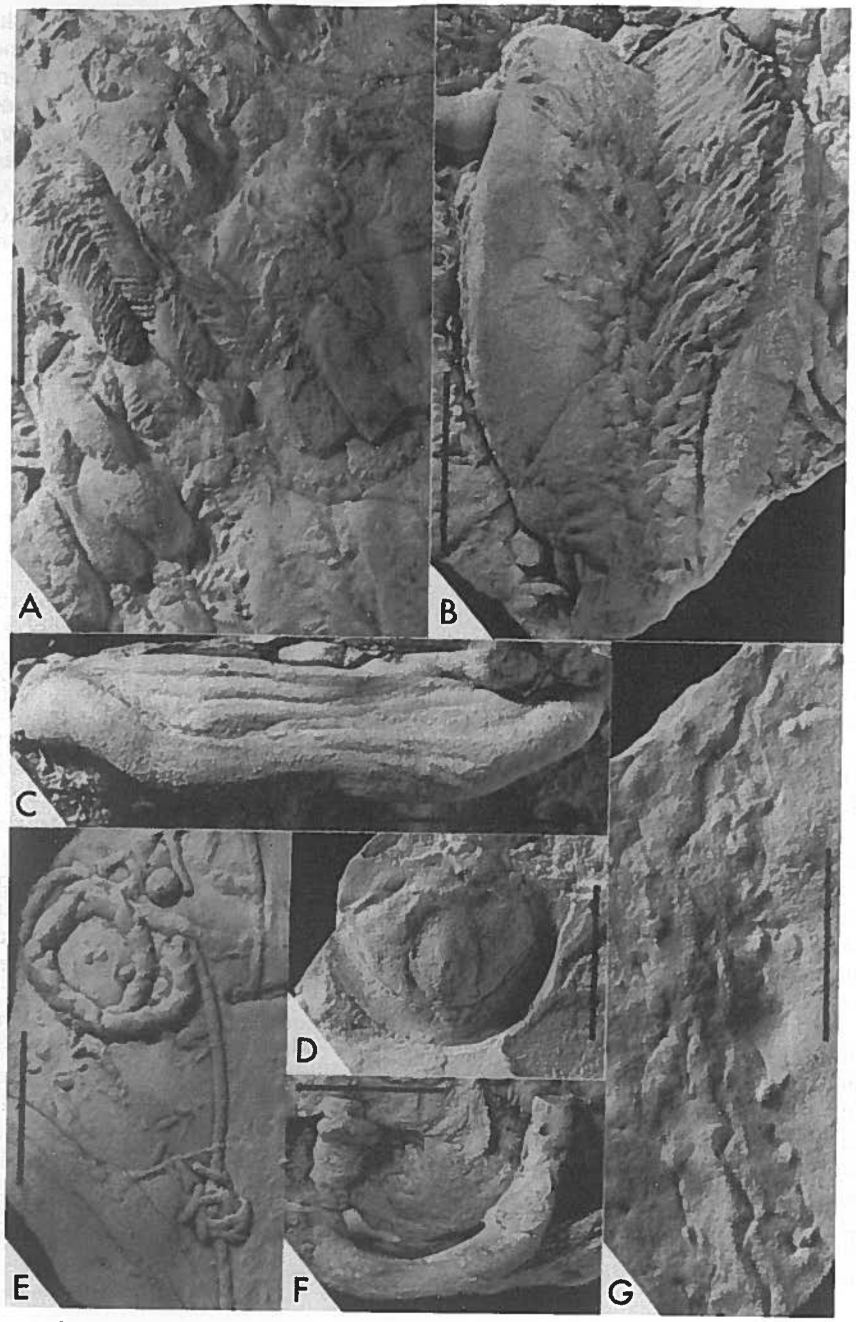

Figure 6.4. Trace fossils from the Bright Angel Shale in the central Grand Canyon. (a) *Cruziana* and *Rusophycus;* (b) *Rusophycus;* (c) *Teichichnus;* (d) *Glossopleura;* (e) *Phycodes pedum;* (f) *Diplocraterion;* (g) *Angulichnus alternipes.*

Table 6–1. Common Invertebrate Trace Fossils

Ichnogenus	*Formation*
Angulichnus	Bright Angel Shale
Arenicoloides	Tapeats Sandstone
Corophioides	Tapeats Sandstone
Cruziana	Bright Angel Shale
Diplichnites	Bright Angel Shale
Diplocraterion	Bright Angel Shale
Palaeophycus	Bright Angel Shale
Phycodes	Bright Angel Shale
Rusophycus	Bright Angel Shale
Scalarituba	Bright Angel Shale
Scolicia	Bright Angel Shale
Skolithos	Tapeats Sandstone
	Bright Angel Shale
Teichichnus	Bright Angel Shale

Elliott and Martin 1987). Consequently, much remains to be done to establish the taxonomic affinities and the relationships between certain physical processes such as current strengths and substrate stability and the mode of infaunal and epifaunal behavior. Studies of other ancient shelf sequences (e.g., Crimes 1970), have demonstrated the benefits of integrating ichnologic and sedimentologic data.

Trace fossils are more abundant in the upper half of the Tapeats Sandstone, particularly in the transition interval into the Bright Angel Shale. These consist of single and paired vertical tubes and several types of horizontal traces.

Unbranched, straight vertical burrows assigned to the ichnogenus *Skolithos* are common at many localities. These sand-filled burrows occur near the top of beds. Burrows of this type probably functioned as dwellings and/or temporary resting structures of suspension-feeding organisms. Their occurrence in fine- to coarse-grained sandstones suggests an environment characterized by currents capable of active bedload transport. This is further substantiated by their occurrence in cross-bedded sandstones. Similar structures are common in many modern nearshore settings.

U-shaped burrows perpendicular to bedding also occur in the fine- to coarse-grained sandstones of the Tapeats Sandstone and the Bright Angel Shale. These tubes appear as paired holes on bedding planes or as concave-upward scours, where they have been eroded to the base of the burrow. These abundant traces, assigned to the ichnogenus *Corophioides*, occur in shallow-water deposits (Hereford 1977). The traces probably represent dwelling structures of suspension-feeding organisms, such as certain groups of annelids, and are common in many modern nearshore deposits.

Horizontal traces first were reported by Walcott (1918) from green shales in the Tonto Group and by McKee (1932) from the Tapeats Sandstone. These so-called "fucoides" are smooth-sided curving traces several inches in length that typically occur in large numbers covering entire bedding surfaces. Presumably, they were formed by the detritus-ingesting annelids moving through the sediment.

Trilobite crawling (*Cruziana*) and resting (*Rusophycus*) traces occur in the transition interval and throughout the Bright Angel Shale (Fig. 6.4a, b). Seilacher

(1970) provided the first detailed description of *Cruziana arizonensis* from the Tapeats Sandstone, and Martin (1985) reported trilobite trace fossils from the Bright Angel Shale. Martin (1985) noted the common occurrence of these traces in interbedded sandstones and mudstones in the Bright Angel Shale. Elliott and Martin (1987) suggested that the *Cruziana* traces were formed during fair-weather periods as these arthropods moved across the muddy shelf sediments. They also suggested that *Rusophycus* marks formed during storms.

Trace fossils are relatively uncommon in the coarser-grained, cross-stratified sandstones of the Bright Angel Shale (Martin 1985). Only the U-shaped trace *Diplocraterion* (Fig. 6.4f) is common, attesting to a relatively mobile substrate where the infauna frequently had to relocate their burrows (Martin 1985). *Diplocraterion* also occurs in upward-fining sequences where preservation of the complete burrow is common. In these sequences, the animals evidently recolonized the substrate and then evacuated the sediments as silt and clay were deposited from suspension following the passage of storms (Elliott and Martin 1987).

Interbedded sandstones and mudstones constitute the most abundant facies sequence in the Bright Angel Shale and also contain the most diverse trace fossil assemblage. Horizontal traces, which dominate these beds, include *Cruziana, Rusophycus, Palaeophycus, Diplichnites, Scalarituba, Scolicia, Angulichnus, Teichichnus,* and *Phycodes* (Elliott and Martin 1987). These traces were produced by organisms burrowing in mud, crawling and feeding on the sediment surface, or moving across the top of a sand bed covered with a thin layer of mud (Elliott and Martin 1987).

Trace fossils are not common in the Muav Limestone. Wanless (1975) described several horizons of burrowed, fine-grained carbonate, yet there has been no attempt to provide systematic descriptions of these trace fossils. Nor have we examined their potential environmental significance. Horizontal burrows, which appear to be the most abundant variety in the Muav, consist of relatively thin, sinuous traces.

DEPOSITIONAL SETTINGS OF THE TONTO GROUP

Cambrian strata in the Grand Canyon accumulated in environments ranging from braided streams to mid-shelf to offshore carbonate buildups. Environments range widely to include beach and intertidal flats; shallow, subtidal sand wave complexes (Tapeats Sandstone); offshore sand sheet deposits; open-shelf, fine-grained sandstones and mudstones (Bright Angel Shale); and subtidal and possibly intertidal carbonate buildups (Muav Limestone and Grand Wash Dolomite). The purpose of this section is to provide detail on the specifics of these depositional systems and to provide a brief review of previous sedimentologic and stratigraphic studies.

Tapeats Sandstone

Deposition of the Tapeats Sandstone was influenced by a variety of geomorphic factors, as well as processes inherent in fluvial and shallow marine depositional environments. The basal sediments of the Tapeats were deposited on a Precambrian surface that had been exposed for long periods of time. There was considerable relief on this surface. These "hills" have relief as great as 800 feet (244 m) (McKee and Resser 1945). Generally, however, the relief is considerably less. Large blocks of the younger Precambrian Shinumo Quartzite occur in the basal deposits of the Tapeats Sandstone near Bright Angel Canyon. Walcott (1883) reported large basement blocks mantling the sides of the areas of high relief. Dott (1974) has shown that similar deposits in Cambrian strata in Wisconsin almost certainly were eroded by storm waves.

McKee and Resser (1945) attempted to reconstruct the depositional environments of the Tapeats Sandstone based on texture of the sediments, paleocurrent trends, and, to some degree, the types of sedimentary structures. These authors concluded that most sedimentation occurred below the beach zone and seaward for tens of miles from the coast in water depths up to 100 feet (33 m). The prevalent west-to-southwest dip of the cross-bedding indicates net offshore transport of sediment. McKee and Resser believed that wide channels filled with cross-stratification and other large-scale scour-and-fill structures represented rip channels oriented perpendicular to the coast.

McKee and Resser (1945) also concluded that the monadnocks, or islands, had relatively little impact on sedimentation—other than serving as a local source of coarse, clastic sediment. They based their conclusions on the relatively consistent dip directions of the cross-stratification. The major influence of these islands, according to the authors, was in modifying sedimentation patterns in the inter-island embayments. They reported southeast-directed current trends in the major embayment that developed between the Shinumo Quartzite islands in Bright Angel Canyon.

A far more detailed facies analysis of the Tapeats Sandstone is that of Hereford (1977). This study was concentrated in Chino Valley and the Black Hills of north-central Arizona. Hereford recognized six environmentally specific lithofacies that he related to physical and biological processes operative on modern tidal flats, beaches, and braided river systems.

This study documented a continuum of tidal flat deposits ranging from lower tidal flats to upper tidal flats dissected by channels. Lower tidal flat sandstones are characterized by complex cross-stratification, reactivation surfaces and herringbone cross-stratification. The complex cross-bedding reflects the passage of smaller bedforms across the surface of larger dunes or sand waves. Reactivation surfaces are common in tidal systems where reversals of flow and/or erosion during periods of bedform immobility result in the scouring of lee-side avalanche deposits. The herringbone cross-stratification reflects bimodal–bipolar flow of the tidal currents. Despite the polymodality of current directions indicated in this study, there is a dominantly southwestern component of sediment transport that is in agreement with the data of McKee and Resser. Tidal systems typically are characterized by asymmetry in flow velocities and durations. In the case of the Tapeats, it is apparent that the ebb phase was the strongest and that it resulted in a preservational bias toward the structure produced during offshore flow.

Deposits of the high intertidal flats are characterized by interbedded sandstones and mudstones exhibiting a variety of features that attest to exposure and late-stage emergent runoff. Additionally, Hereford (1977) was able to document the presence of tidal channels that drained the flats.

Large channels occur near the top of the Tapeats at several localities in the central and western canyon (Fig. 6.5). The channels are up to 13 feet (4 m) deep and 60 feet (18 m) wide. At several localities, up to three laterally contiguous channels form a complex of southwest-oriented channel systems. Channel fill is variable and consists of thick sets of planar tabular cross-stratification, co-sets of planar tabular and trough cross-stratification, and/or simple vertical fills that conform to the shape of the channel. Flow tends to parallel the southwestern strike of the channel axis, and in some instances there is a well-developed bi-

FIGURE 6.5. Large subtidal channel complex near top of the Tapeats Sandstone. Transition zone is indicated by covered slope below first Bright Angel cliff.

modal–bipolar orientation to the cross-bed dip directions. Vertical trace fossils occur in the upper parts of the channel fills.

Although the geometry of these channels is similar to that of both fluvial and tidal flat channels, the internal stratification differs from that found in fluvial sequences in the Tapeats Sandstone (Middleton and Hereford 1981). The presence of trace fossils and bimodal–bipolar foreset dips, along with the absence of exposure features, suggests a subtidal channel complex dominated by offshore flow, with minor preservation of flood-oriented structures. Although subtidal channels occur on modern tide-dominated coasts, comparatively little is known of their sedimentologic characteristics. In the lower intertidal and shallow subtidal zones, bedload transport and erosion can be intense, particularly along meso- and macrotidal coasts, because of the concentration of flow in these areas.

Johnson (1977) documented similar deposits in late Precambrian shallow-marine, quartz arenites in Norway. He demonstrated that these subtidal channels were oriented perpendicular to the coast and separated (dissected) subtidal sand bodies. To date, we have not gathered enough data to identify definitively the sandstone bodies that surround these channels as subtidal ridges or sandwaves. Nor have we established the fair-weather or storm-generated origin of these channels.

Two other facies reported by Hereford (1977) were not documented in the study of McKee and Resser (1945). One comprises low-angle, cross-laminated sandstone that likely formed on beaches. These tend to occur most frequently around Precambrian highs, where beach and upper foreshore sediments should have been common.

The second facies association represents braided stream deposits that grade into the marine units. Fluvial deposits in the Tapeats occur in the basal portions of the formation. Typically, they are less mature texturally and mineralogically than the associated marine deposits, which reflects a lack of extensive reworking that is common in the high-energy nearshore. These deposits are characterized by broad, shallow channels filled by horizontally stratified, coarse-grained sandstone and conglomerate that alternate with thick sets of planar-tabular and trough cross-stratified sandstone (Middleton and Hereford 1981). This sequence of structures, which is consistent with processes operative in coarse-grained, braided fluvial systems, has been reported from other pre-vegetation fluvial systems (Cotter 1978; Middleton et al. 1980; Cudzil and Driese 1987). Depositional occurred in wide, shallow streams where in-channel transport of sediment was accomplished by the movement of sheets of coarse-grained sediment along the bed and by migration of dunes and slightly sinuous transverse bars.

Bright Angel Shale

The Bright Angel Shale comprises a variety of lithologies and sedimentary and biogenic structures that indicate deposition in open shelf environments. McKee and Resser (1945) concluded that the Bright Angel Shale accumulated in waters below wave base at depths intermediate between the shallow water represented by the Tapeats Sandstone and the deeper waters of the Muav Limestone. More recent work by Wanless (1973), Martin et al. (1986), Elliott and Martin (1987), and Rose et al. (1998) have provided new data that permit more precise environmental reconstructions. Although generally supporting the conclusions of McKee and Resser, these workers have documented shallow-water deposits in the Bright Angel Shale, as well as providing important information concerning the roles of fair-weather and storm-related processes in controlling depositional patterns in the Bright Angel Shale.

Martin (1985) recognized eight facies in the Bright Angel Shale and grouped these into three genetically significant facies sequences. These include cross-bedded, upward-coarsening, and upward-fining sequences and a heterolithic sequence consisting of interbedded sandstone and mudstone. These facies sequences reflect deposition in subtidal areas influenced by tidal and meterologic processes. They also aid in tracking transgressive and regressive strandline movements.

Upward-coarsening sequences are up to 25 feet (8 m) thick and typically can be traced for several tens of kilometers. The lower parts of these sequences are characterized by laminated, bioturbated mudstones that were deposited during fair-weather suspension settling of silt and clay. The coarser-grained portions accumulated as sand waves, dunes, and ripples that migrated over sand sheets. These portions are characterized by thick sets of planer tabular cross-stratification (Middleton 1988). The presence of reactivation surfaces and abrupt changes in the dip of many foresets indicates periodic movement of these large bedforms, many of which are palimpset or relict. It also may indicate lee-side erosion during tidal reversals and/or storms. Deposition was entirely in subtidal areas, though the upper portions of many of these sequences were deposited in relatively shallow waters, as evidenced by eroded burrows of *Diplocraterion* (Elliott and Martin 1987).

Sequences that fine upwards are common and consist of a lower, normally graded small-pebble conglomerate or sandstone overlying an erosive base. This, in turn, is overlain by interbedded and fine-grained sandstone and mudstone. The basal coarse-grained facies represent deposition from high-energy, storm-induced currents that transported coarse materials from nearshore areas. The tops of these sequences contain symmetrical ripples, as well as appreciable amounts of laminated mudstone deposited from waning flows following the passage of storms (Fig. 6.3). These beds are very similar to those reported from both modern and ancient storm deposits. Complete vertical and horizontal traces occur at the top of many beds, indicating that the substrate was recolonized soon after deposition (Elliott and Martin 1987).

Lenticular beds of interbedded sandstone and mudstone constitute the majority of the Bright Angel Shale. This association comprises very fine-grained sandstone lenses and micaceous shale. Most beds are graded normally and contain an abundant and diverse trace fossil assemblage (Elliott and Martin 1987). These deposits represent post-storm suspension settling of muds and sands—and, possibly, remobilization during fair-weather periods. *Cruziana* and *Rusophycus* indicate a substrate inhabited by trilobites. Other trace fossils also indicate a relatively stable substrate colonized by a variety of infaunal and epifaunal organisms.

Muav Limestone

McKee and Resser (1945) considered the Muav Limestone to have been deposited in subtidal environments. The subtidal origin of much of the Muav is based on faunal and textural characteristics. These include an open-marine fauna, the very fine-grained nature of the mottled limestone and dolostone facies, and the fact that the Muav grades eastward into a shallow-water facies of the Bright Angel Shale. These authors also indicated that many of the flat-pebble conglomerates occurring throughout the formation were deposited in relatively deep water.

Intraformational or flat-pebble conglomerates are an extremely important facies in the Muav Limestone. These deposits, which are abundant from the Bass Trail eastward, consist of disc-like clasts of micrite and, occasionally, silt-size quartz and glauconite grains. The orientation of these clasts is variable. Some are oriented parallel with the bedding; some clasts are imbricated, and in some instances the clasts are vertical.

McKee and Resser (1945) described two associations of these conglomeratic beds. One variety consists of intraformational conglomerates that occur as scattered, discontinuous lenses within thinly bedded limestones. The other variety consists of one to several thin, conglomeratic beds that extend up to 45 miles. The great lateral persistence of these beds makes them ideal stratigraphic markers, and McKee and Resser used them to correlate over great distances in the canyon. These workers considered the widespread conglomerates to represent subtidal deposits formed during regressions.

The origin of the clasts obviously requires early lithification by cementation and/or compaction because they are derived from sediments within the basin. Where this induration takes place is controversial. Opinions range from ripups of carbonate muds exposed on tidal flats by storms and/or tidal channels to submarine lithification and subsequent erosion during storms.

Dew (1985) documented the occurrence of intraformational conglomerates similar to those reported from the Muav Limestone in the Upper Cambrian DuNoir Limestone in Wyoming. Based on facies associations, her study showed that both intertidal and subtidal limestone conglomerates can occur over a short stratigraphic interval. Sepkoski (1982) demonstrated that storm-induced currents were mostly responsible for the widespread distribution of flat-pebble conglomerates in Montana's Cambrian strata. In this case, the conglomerates are interbedded with shales that lack any evidence of subaerial exposure. Considering the stratigraphic importance that has been made of these conglomerates, it is clear that

they represent key lithofacies and that documentation of their mode of origin (i.e., subaerial or subtidal) needs to be determined.

Regardless of their mode of origin, it is particularly interesting that they are abundant only in Cambrian and Ordovician rocks. Sepkoski (1982) speculated that with the proliferation of organisms living within the sediment during and following the Ordovician, the potential for early submarine cementation of carbonate shelf deposits was reduced substantially due to bioturbation processes. This hypothesis has yet to be tested and, of course, assumes a subtidal origin for these deposits.

Although many of the limestone and dolomite beds in the Muav are subtidal, Wanless (1973, 1975) reported intertidal and supratidal facies from outcrops in the western Grand Canyon. Many of the textures and structures reported by Wanless are similar to those found in modern tidal flats on Andros Island in the Bahamas. In particular, the laminated dolostones in the Muav have many characteristics in common with laminated dolomites that occur on supratidal levees adjacent to tidal channels on Andros Island. In these areas, fine-grained carbonate sediment is deposited during periods of overbank flooding following storms. Algae that inhabit the levees trap the sediment, resulting in the generation of continuous laminae of carbonate mud and pellets. Aitken (1967) referred to these laminated horizons as cryptalgal laminations because the evidence of algal binding had to be inferred. Discontinuous laminae also occur and are produced by traction transport of pellets and other grains over the algal-bound sediment. Wanless (1975) reported that these units are up to 66 feet (20 m) thick in the Muav Limestone. This suggests that there were prolonged periods of supratidal sedimentation far offshore from the Cambrian strandline.

It is clear that the Muav Limestone records episodes of both subtidal and peritidal deposition. A reasonable depositional model, therefore, might be one of offshore shoals surrounded by deeper water areas. Pratt and James (1986) proposed a tidal flat island model for Lower Ordovician shelf carbonates of Newfoundland (Fig. 6.6). In this model, small, localized carbonate islands occurred far offshore and were separated by subtidal areas. Middleton et al. (1980) documented similar facies distributions from Cambrian strata in Wyoming.

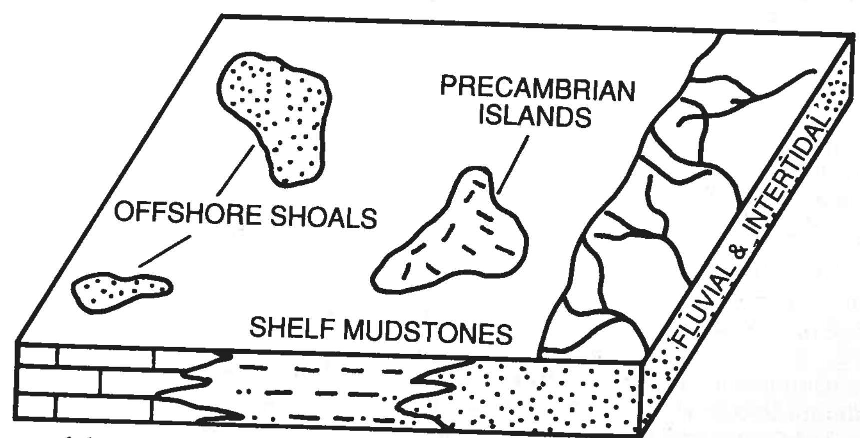

Figure 6.6. Block diagram illustrating the distribution of depositional environments represented by the Tapeats Sandstone, Bright Angel Shale, and Muav Limestone.

Undifferentiated Dolomites

The undifferentiated dolomites that overlie the Muav are not well understood in terms of their temporal and environmental significance. Brathovde (1986), however, has documented thick beds of oolitic grainstones and stromatolites interbedded with fine-grained carbonates. This association clearly indicates shallow subtidal and, possibly, intertidal environments.

SUMMARY

Facies analyses of the Tonto Group indicate deposition in a variety of fluvial, nearshore, and shallow shelf environments. Braided stream and intertidal-to-shallow subtidal deposits of the Tapeats Sandstone grade seaward into a complex array of shallow shelf sands and muds of the Bright Angel Shale. Shelf sedimentation was influenced by both tidal and storm currents. Sand ridges, sand waves, and broad areas where fine-grained siliciclastics were deposited from suspension settling—following storms and during fair-weather periods—characterized the shelf. Farther offshore, carbonate islands dotted the shelf. Here, the carbonate buildups were characterized by intertidal and possible supratidal zones separated by deeper water areas where tidal currents were active and where finer-grained carbonate sediments were deposited.

A number of transgressions and regressions resulted not only in the intertonguing of the formations of the Tonto Group but also in the vertical juxtaposition of facies belts that probably were not laterally adjacent. As Curray pointed out in his 1964 study, rapid migration of the strandline can result in the overstepping of offshore facies over nearshore deposits with no record of intervening environments. Numerous examples of such transitions occur in the Tonto Group, but most are poorly documented.

More detailed facies mapping and, in particular, documentation of lateral facies changes are needed. Only through such studies can the nature of the transgressive and regressive stratigraphies preserved in the strata of the Tonto Group be reevaluated and related to regional paleogeography.

• 7 •

Temple Butte Formation

Stanley S. Beus

INTRODUCTION

Strata of the Temple Butte Formation of early Late Devonian and possibly late Middle Devonian age are exposed through most of the Grand Canyon but are relatively inconspicuous. Outcrops in the east are thin, discontinuous lenses, and in central and western Grand Canyon the exposures, though continuous, tend to merge with cliffs of the much thicker overlying Redwall Limestone. Temple Butte lithology is predominantly dolomite or sandy dolomite with minor sandstone and limestone beds.

NOMENCLATURE

Rocks of Devonian age in the Grand Canyon were first reported by Walcott (1880, 1883), who recognized Devonian strata beneath the Redwall Limestone in Kanab and Nankoweap canyons. He applied the name Temple Butte Limestone (Walcott 1889) to a thin band of dolomite along Temple Butte (on the west side of the Colorado River in eastern Grand Canyon). Although no specific type section has been designated, the Temple Butte site (Fig. 7.1) has gained universal acceptance. McKee (1939) and others have extended the name, as Temple Butte Formation, to the much thicker and more extensive outcrops in western Grand Canyon. West and north of the Grand Canyon, Devonian strata equivalent to the Temple Butte are designated as the Muddy Peak Limestone (Longwell et al. 1965), a name derived from the Muddy Mountains of southern Nevada (Longwell 1921). South of the Grand Canyon, along the Mogollon Rim of central Arizona, equivalent Devonian rocks are recognized as the Martin Formation (Teichert 1965), a name originally applied to Devonian strata in the Bisbee area of southeastern Arizona (Ransome 1904).

DISTRIBUTION

In eastern Grand Canyon and upstream in Marble Canyon the Temple Butte crops out as scattered, lens-shaped exposures that fill channels eroded into the upper surface of the Muav Limestone or the Cambrian undifferentiated dolomite. These channel-fill lenses commonly are less than 100 feet (30 m) thick but may

Figure 7.1. Approximate location of the type section of the Temple Butte Formation, on west side near south end of Temple Butte. Cu, Cambrian undifferentiated dolomite; Dtb, Temple Butte Formation; Mr, Mississippian Redwall Limestone.

be up to 400 feet (120 m) wide. Numerous exposures of these lenses are visible near river level in Marble Canyon beginning just below mile 37 (see Fig. 7.2) and also in the Little Colorado gorge. From Hermit Creek (about mile 95) westward throughout central and western Grand Canyon, the Temple Butte forms a continuous band of dolomite above local channel-fill deposits at the

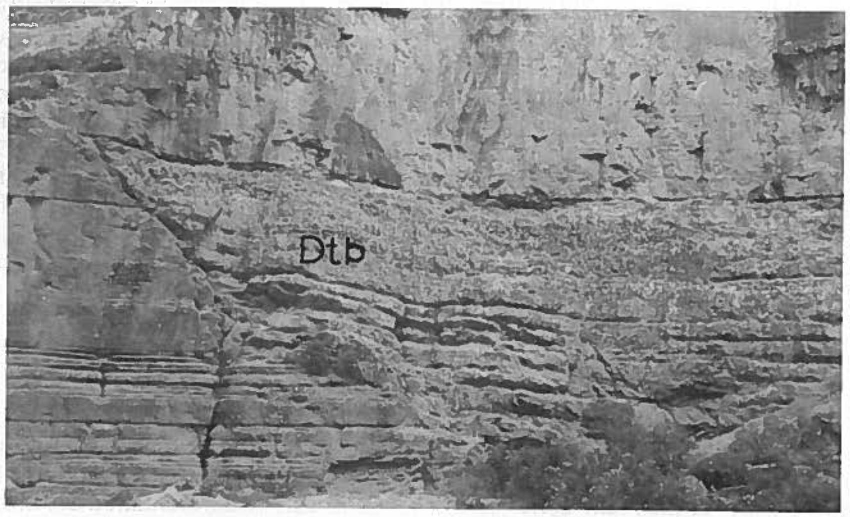

Figure 7.2. Temple Butte channel-fill outcrop on right bank of Marble Canyon at approximately mile 38.4. Dtb, Temple Butte Formation.

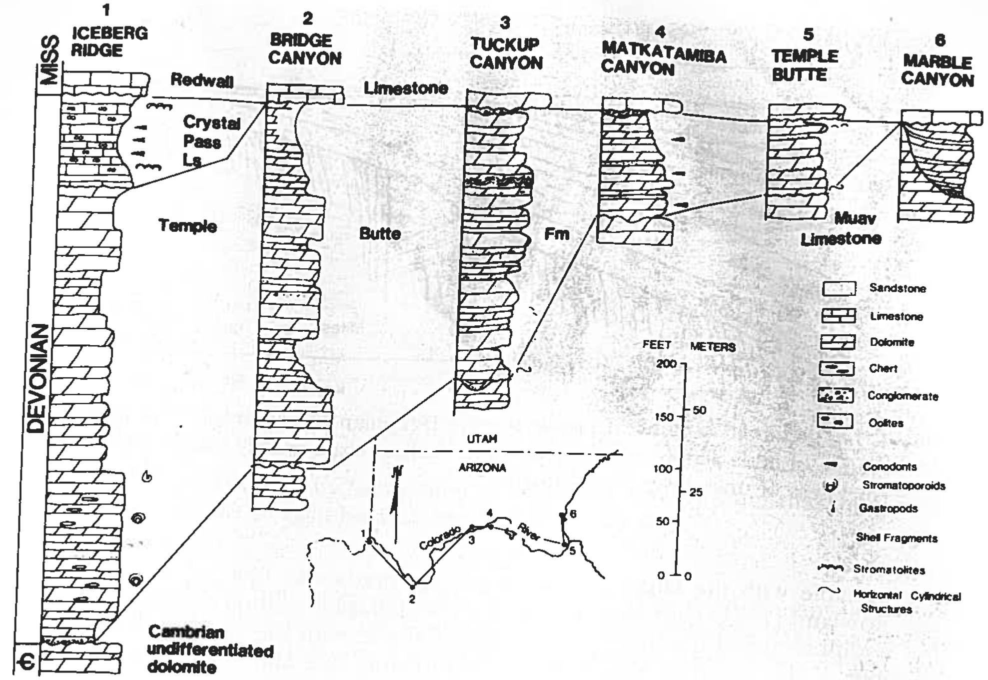

Figure 7.3. Selected stratigraphic sections of the Temple Butte Formation.

base. It gradually thickens to more than 450 feet (220 m) at Iceberg Ridge, five miles west of the mouth of the Grand Canyon (Fig. 7.3). Earlier descriptions of an Iceberg Ridge Devonian section more than 1200 feet (365 m) mistakenly included several hundred feet (220 m) of unnamed Cambrian dolomite beds [the undifferentiated Cambrian beds of McKee and Resser (1945)] as part of the Temple Butte.

CAMBRIAN–DEVONIAN UNCONFORMITY

In a broad, regional sense, Devonian strata truncate successively older rocks from west to east across northern Arizona (Fig. 7.4). The unconfomity at the base of the Temple Butte is one of the major stratigraphic breaks in the Paleozoic sequence of the Grand Canyon. It probably represents latest Cambrian, all of the Ordovician and Silurian, and most of Early and Middle Devonian time. The time gap involved is probably somewhat less in western Grand Canyon because the uppermost Cambrian strata were deposited in a sea regressing westward and the Temple Butte was formed in a sea transgressing eastward.

In western Grand Canyon a sequence of light-gray dolomite beds up to 500 feet (150 m) thick overlies, and appears conformable with, the Muav Limestone. These strata were referred to as undifferentiated Cambrian dolomite by McKee and Resser (1945). The beds are truncated gradually eastward by the Temple Butte but are present near Lava Canyon in easternmost Grand Canyon, where McKee and Resser (p. 141) reported 163 feet (50 m) of dolomite and siltstone. Although no diagnostic fossils are known from these unnamed beds, they are

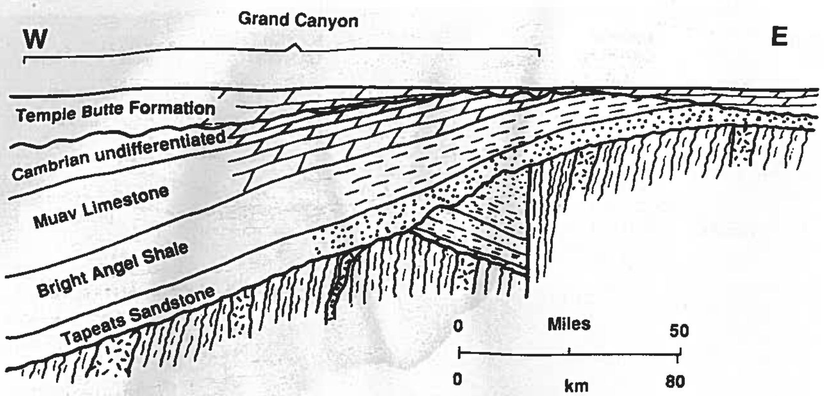

Figure 7.4. Regional pattern of Devonian–pre-Devonian unconformity and angular discordance of beds across northern Arizona. Vertical scale greatly exaggerated, relative thickness of units only approximate.

conformable with the Muav Limestone below (Brathovde 1986, p. 1). Korolev and Rowland (1993) have demonstrated on stratigraphic criteria a correlation of the undifferentiated dolomite in the Grand Canyon with the Banded Mountain Member of the Bonanza King Formation (probable late Middle Cambrian age) of Nevada.

The unconformable surface at the base of the Temple Butte is marked locally by considerable relief in the form of channels and depressions cut into the underlying Cambrian strata. These channels, up to 100 feet (30 m) deep in Marble Canyon, occur throughout most of the Grand Canyon. Where present, they clearly mark the Devonian base. They were first observed in Kanab Canyon by Walcott (1880) and were studied and described in detail between Garnet and Cottonwood Creek in eastern Grand Canyon by Noble (1922, pp. 49–51). In parts of the Grand Canyon, including the type section on Temple Butte (where the channels are absent), the Cambrian–Devonian strata appear in local exposures to be without angular discordance, and the contact is planar, with gray dolomite beds below and above. Here, the unconformity, even though representing more than 100 million years, may be difficult to locate.

STRATIGRAPHY

The strata forming the basal channel-fill part of the Temple Butte are commonly a distinct pale, reddish purple dolomite or sandy dolomite. The bedding generally is irregular, and it is gnarly in places. At some localities the beds are horizontal, but elsewhere they may conform to the walls of the channel they fill and be truncated with angular discordance by the overlying beds (Fig. 7.5). Rarely, there are basal conglomerate beds composed of subrounded dolomite pebbles. The upper beds in the channel-fill illustrated in Figure 7.2 contain about 20 percent insoluble residue consisting of detrital quartz grains, clay, and hematite. They also exhibit a peculiar columnar pattern of pale gray and purple dolomite—perhaps the result of leaching or weathering.

The more continuous strata above the basal channel-fill beds of the Temple Butte are exposed as uniformly medium to thick blocky ledges outlined by

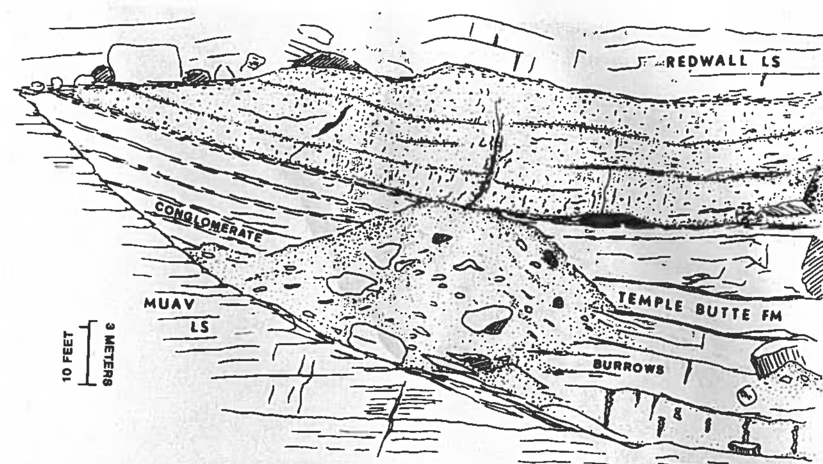

Figure 7.5. Field sketch by James Obey of portion of Temple Butte channel-fill lens in Marble Canyon on right bank at approximately mile 38.4.

parting planes or thin recesses (Figs. 7.6 and 7.7). The predominant lithology is dark-to-light olive gray, fine to medium crystalline dolomite that commonly weathers to a sugary texture. A subordinate, though extensive, second carbonate lithology is that of thin-bedded, very fine grained or aphanitic dolomite that is grayish-orange-pink. It weathers to a very light gray or yellow, exhibits a concoidal fracture, and commonly appears porcelaneous. Rounded, frosted quartz sand grains occur either scattered or as thin lenses or laminae in some of the

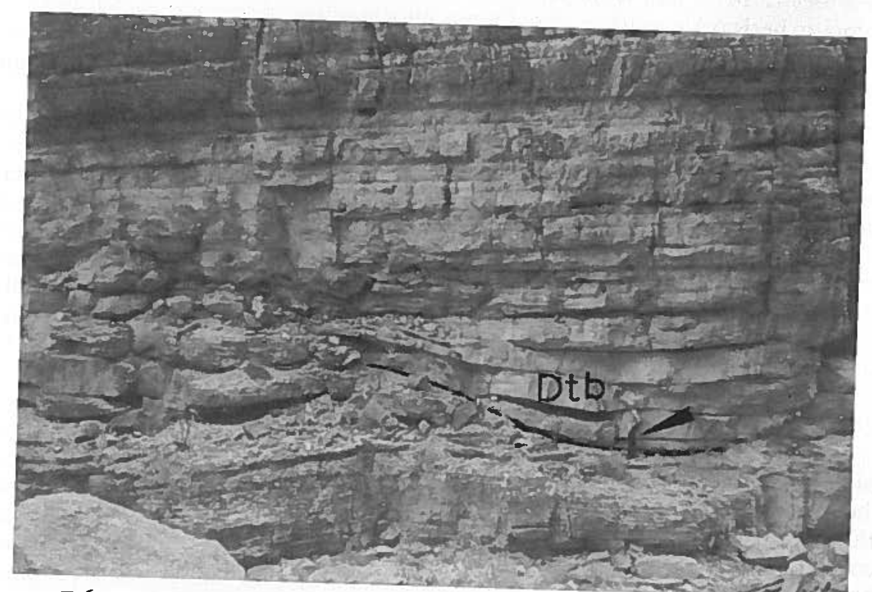

Figure 7.6. Basal part of Temple Butte Formation at Tuckup Canyon near river mile 164.5 showing channel cut into underlying Cambrian undifferentiated dolomite beds. Arrow indicates a 6-ft figure.

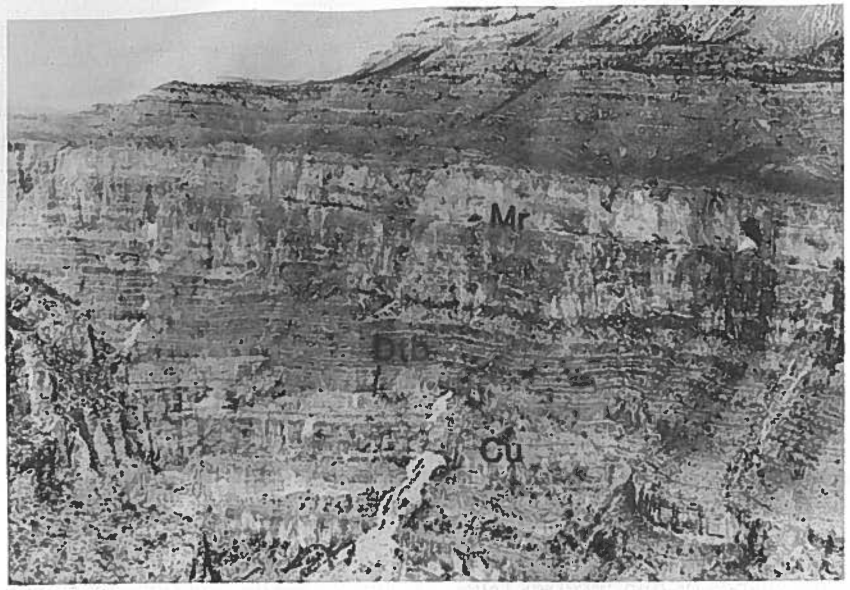

Figure 7.7. Paleozoic strata in western Grand Canyon near the mouth of Quartermaster Canyon, about river mile 265. Mr, Redwall Limestone; Dtb, Temple Butte Formation; Cu, Cambrian undifferentiated dolomite.

aphanitic dolomite beds. Locally they form marker beds of quartz sandstone up to 7 feet (2 m) thick, as at Havasu and Matkatamiba Canyon (Fig. 7.3). The dolomite beds commonly crop out as a uniform series of steep, receding ledges, but in central Grand Canyon they may form part of a vertical 1550-foot (460-m) carbonate wall that appears unbroken from the Muav Limestone up through the Redwall Limestone.

At Iceberg Ridge, five miles west of the Grand Canyon mouth, the upper 85 feet (26 m) of the Devonian carbonate section, above the Temple Butte Formation, consists of light gray lime mudstone to oolitic wackestone with a basal conglomerate or breccia. These beds previously were treated as part of the basal Whitmore Wash Member of the Redwall Limestone. Ritter (1983, p. 6) has recovered diagnostic latest Devonian (Famennian) conodonts from these strata and assigned them to the Late Devonian Crystal Pass Limestone. This unit apparently pinches out eastward and thus has not been recognized in the Grand Canyon.

PALEONTOLOGY AND AGE

Although marine fossils are abundant in the Upper Devonian Jerome Member of the Martin Formation in central Arizona (Teichert 1965; Beus 1978), the Temple Butte formation has yielded surprisingly few identifiable organic remains. Walcott (1883, p. 221) reported indeterminate brachiopods, gastropods, corals and "placoganoid" fish from the walls of lower Kanab Canyon. Noble (1922) reported fish plates identified as *Bothreolepis*, a fresh or brackish water form, from Sapphire Canyon. Denison (1951) confirmed the identification and the Late Devonian age assignment for the fish plates. Rare silicified corals, gastropods, crinoid

plates, and massive stromatoporoids occur in the Temple Butte section at Iceberg Ridge, just west of the Grand Canyon mouth, but none are identifiable to the generic level.

Peculiar cylindrical trace fossil forms, somewhat resembling *Paleophycus*, occur in dolomite beds near the base of the Temple Butte at the type section and at Tuckup Canyon. These forms are subhorizontal, are straight to gently arcuate, and have a micritized central core within a cylindrical sleeve about 3 mm in diameter (Fig. 7.8). They may be trace fossils or possibly some sort of algal or stromatoporoid structure, but to date are indeterminate.

Conodont microfossils reported from the Temple Butte Formation at Matkatamiba Canyon (about 148 miles on the Colorado River in Grand Canyon) by Elston and Bressler (1977) are the most diagnostic and significant fossils yielded by the Temple Butte. These conodonts, identified by D. Schumacher (1978, written communication), indicate possibly a late Givetian lowermost *Polygnathus as-*

Figure 7.8. Subhorizontal cylindrical structures near the base of the Temple Butte type section on Temple Butte. Scale is in inches.

symertricus Zone assemblage occur at the base of the section, an early Frasnian assemblage between 20 and 40 feet (6 to 12 m) above the base of the Temple Butte, and probable early late Frasnian forms 20 feet (6 m) below the top of the formation. The uppermost 20 feet (6 m) of the Temple Butte at the Matkatamiba Canyon section are devoid of fossils. The conodonts suggest a latest Givetian to late Frasnian age (latest Middle Devonian through early Late Devonian) for most of the formation in central Grand Canyon. The Temple Butte is thus the approximate age equivalent of the Jerome Member of the Martin Formation in central Arizona, the Muddy Peak Formation of southern Nevada, and the Elbert Formation (Knight and Cooper 1995) of the subsurface in northeastern Arizona.

DEPOSITIONAL ENVIRONMENT

Devonian strata in the Grand Canyon are perhaps the least understood of the Paleozoic units studied there. The dolomitization of the original limestone and rarity of recognizable fossils have made refined environmental interpretations difficult. The carbonate facies and the few known fossils—crinoids, corals, stromatoporoids, and conodonts typical of nearshore biofacies (D. Schumacher 1984, written communication)—indicate accumulation in shallow, subtidal, open marine conditions for most of the Temple Butte in central and western Grand Canyon. The aphanitic dolomite beds in the Temple Butte may record local supratidal conditions. Similar modern, supratidal dolomites have been formed through evaporative pumping of magnesium-enriched sea water moving through porous supratidal sediments in the Persian Gulf as described by Illing et al. (1965) and Shinn et al. (1965).

The thinner and discontinuous channel-fill deposits of the Temple Butte in eastern Grand Canyon may have been deposited in tidal channels in an intertidal environment.

The regional paleogeography of the Frasnian (early Late Devonian) appears to have been shallow (but generally open-circulation) marine conditions in northwestern and central Arizona; intertidal-to-very-shallow subtidal conditions in central and eastern Grand Canyon; and shallow marine, restricted circulation conditions in northeastern Arizona (Beus 1980).

SUMMARY

The Temple Butte Formation records minor deposition of a thin carbonate sequence that gradually thickens westward across the Grand Canyon region. The easternmost outcrops and some of the basal strata to the west were probably deposited in narrow tidal channels. The more laterally extensive dolomite beds in the central and western portions of Grand Canyon accumulated in more subtidal but probably very shallow marine conditions across a gently submerged continental shelf.

• 8 •

Redwall Limestone and Surprise Canyon Formation

Stanley S. Beus

INTRODUCTION

Two formations of Mississippian age are recognized in the Grand Canyon. The oldest of these, the Redwall Limestone, of Early and early Late Mississippian age, is one of most prominent topographically displayed units in the canyon wall. Recently, a second unit, The Surprise Canyon Formation of latest Mississippian age, has been recognized as the result of detailed mapping in the more remote part of western Grand Canyon. It occurs as isolated patches and lenses and occupies stream valleys, caves, and collapse structures developed in the top of the Redwall Limestone.

REDWALL LIMESTONE

Nomenclature

The Redwall Limestone consistently forms massive, vertical cliffs 500 to 800 feet (150 to 250 m) high about midway up the canyon wall (Figs. 8.1 and 8.2). The cliff face generally is stained red by iron oxide material washed down from the redbeds in the overlying Supai Group. The name Red Wall Limestone was applied first by Gilbert (1875, p. 177) in a report for one of the early surveys west of the 100th Meridian directed by John Wesley Powell. A type locality was later established by Darton (1910) when he introduced the name Rewall Canyon to a Grand Canyon tributary in the Shinumo quadrangle and thus provided a geographic place name for the Redwall. Four distinct stratigraphic units within the Redwall were recognized by Darton (1910). Later, they were described in more detail by Gutschick (1943) and given formal names by McKee (1963). These units, in ascending order, are: the Whitmore Wash, Thunder Springs, Mooney Falls, and Horseshoe Mesa members. All four have their type locality in the Grand Canyon or its tributaries, and all four can be traced throughout the Grand Canyon and beyond (McKee and Gutschick 1969, p. 3).

Distribution

The Redwall Limestone originally was deposited across virtually all of northern Arizona except the Defiance positive area of east central Arizona It is recognized

Figure 8.1. View of Redwall Limestone cliffs along the South Kaibab Trail in eastern Grand Canyon. Lowermost sharp "V" in trail switchbacks marks approximate Redwall Limestone-undifferentiated Cambrian(?) contact.

as far south as the Gold Gulch area just north of globe in south central Arizona (Racey 1974). Within the Grand Canyon, it is exposed in almost continuous outcrop on both canyon walls from about river mile 22 in Marble Canyon, where it first appears at river level, to the mouth of the canyon at the Grand Wash Cliffs (mile 277). Thickness of the formation gradually increases northwestward. It is just over 400 feet (120 m) thick at the Tanner Trail section in easternmost Grand Canyon and about 800 feet (245 m) at Iceberg Ridge, 5 miles (8 km) west of the Grand Canyon's mouth (Fig. 8.3) (McKee and Gutschick 1969, p. 3).

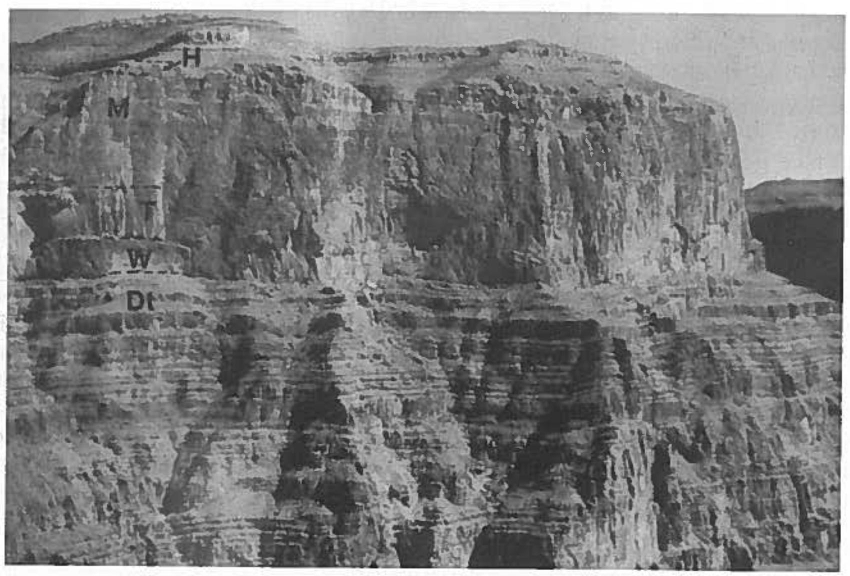

Figure 8.2. View of Redwall Limestone cliffs in western Grand Canyon near Separation Canyon, mile 240. H, Horseshoe Mesa Member; M, Mooney Falls Member; T, Thunder Springs Member; W, Whitmore Wash Member; Dt, Temple Butte Formation.

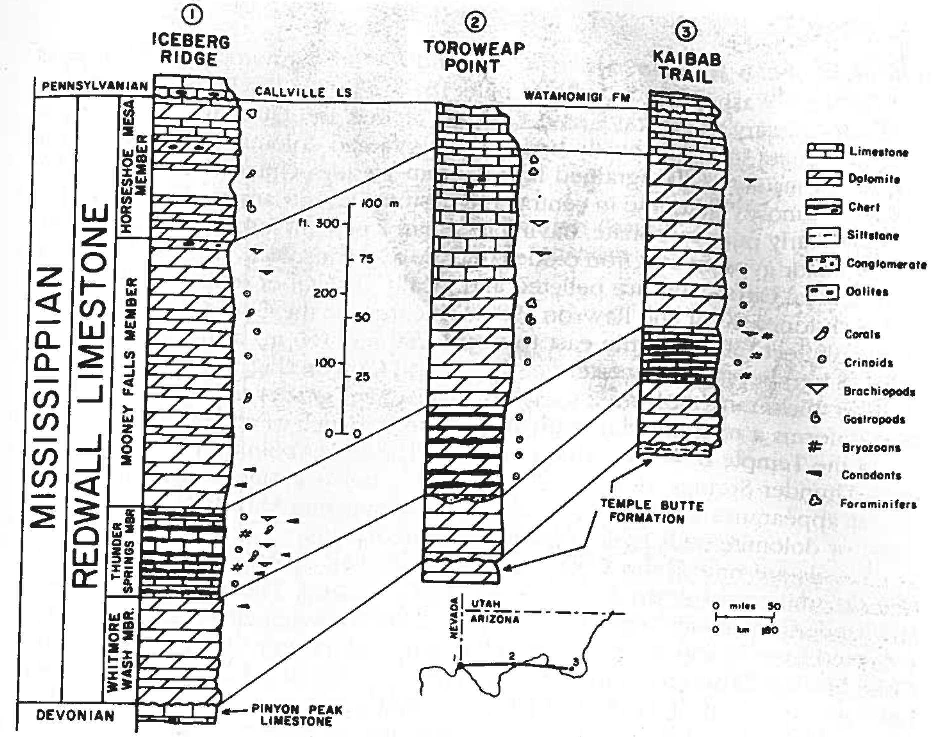

Figure 8.3. Selected stratigraphic sections of the Redwall Limestone in Grand Canyon.

PRE-REDWALL UNCONFORMITY

Throughout most the Grand Canyon, the Redwall Limestone rests without angular discordance upon Devonian strata or, where the Temple Butte Formation is missing, upon rocks of Cambrian age. The unconformity at the base of the Redwall spans all of latest Devonian (Famennian) time and the earliest part of the Missisippian Period (early Kinderhookian) through central and western Grand Canyon. The magnitude of the unconformity increases eastward because of the transgressive nature of the basal Redwall (Fig. 8.2). Basal beds are Early Mississippian (Kinderhookian) age in westernmost Grand Canyon, as indicated by diagnostic foraminifers, and are of late Early Mississippian (Osagian) in eastern Grand Canyon. In western Grand Canyon, the Mississippian–Devonian contact generally is well-marked by an irregular surface of erosion having up to 10 ft (3 m) of relief in a lateral distance of 100 to 200 feet (30 to 60 m).

Locally, a basal conglomerate composed of angular dolomite or limestone blocks of the Devonian Temple Butte Formation occurs at the unconformity. In eastern Grand Canyon, the Redwall–Devonian contact most commonly is a nearly horizontal surface with little or no relief. It is more difficult to recognize the unconformity where the Cambrian or Devonian strata beneath the unconformity and the Mississippian strata above are both dolomite. The nature of the unconformity suggests only gentle uplift and mild, though perhaps locally prolonged, erosion of pre-Mississippian strata before the beginning of Redwall deposition.

Stratigraphy

Whitmore Wash Member The type section for the Whitmore Wash Member is at Whitmore Wash (Colorado River mile 187.5 below Lees Ferry). Along this northern tributary to central Grand Canyon, McKee and Gutschick (1969) measured 101 feet (30 m) of thickly bedded, fine-grained dolomite. This member is composed mainly of fine-grained limestone in western Grand Canyon, but this changes to mostly dolomite in central and eastern Grand Canyon. The Whitmore Wash is nearly pure carbonate, having less than 2 percent insoluble residue content of minor gypsum and iron oxides (McKee and Gutschick 1969, p. 27). Common textural carbonates are pelleted and locally skeletal or oolitic wackestones and packstones (Kent and Rawson 1980). Thickness in the Grand Canyon is from about 100 feet (30 m) in the east to nearly 200 feet (60 m) at Iceberg Ridge, 5 miles (8 km) beyond the western end of Grand Canyon (Fig. 8.3). Bedding generally is thick, ranging from 2 to 4 feet (0.6 to 1 m) and even thicker locally. It usually forms a resistant cliff overlying a narrow bench or series of ledges typical of the Temple Butte Formation beneath. The upper boundary with the overlying Thunder Springs Member is conformable but is easily recognized by the lowest appearance of thin, dark, chert beds alternating with lighter gray limestone or dolomite beds typical of the Thunder Springs.

Fossils are rare in the Whitmore Wash Member, probably owing to extensive dolomitization of the original lime mud and sand. The Whitmore Wash is late Kinderhookian and early Osagian (late Early Mississippian) to the east as interpreted from brachiopod and foraminiferid fossils. Racey (1974) reported late Kinderhookian conodonts from the lower part of the member in the Salt River Canyon area south of the Grand Canyon. Ritter (1983, p. 17) noted conodonts of possible earliest Osagian age in the upper part at Iceberg Ridge.

Thunder Springs Member The Thunder Springs type section is at the head of Thunder River about 2 miles (3 km) north of Colorado River mile 136 in central Grand Canyon. The Thunder Springs Member is the most distinctive member of the Redwall Limestone because of the light and dark banded appearance imparted by alternating chert and carbonate beds. It consists of thin beds of light gray limestone or dolomite alternating with thin beds of dark reddish brown or dark gray weathering beds or lenses of chert. Thickness of the member increases gradually from 100 feet (30 m) in eastern Grand Canyon to about 150 feet (46 m) in the west (McKee and Gutschick 1969, p. 41).

Most of the carbonate rock in the Thunder Springs is thin-bedded, crinoidal grainstone or packstone. The rock tends to be limestone in the west and dolomite in the east. Thin-section analyses of the chert beds in the member as exposed in central Arizona reveal them to be silicified former bryozoan wackestones and mudstones (Bremner 1986, p. 55).

Invertebrate marine fossils are especially abundant in the chert beds of the Thunder Springs Member. These include corals (particularly colonial *Syringopora*), bryozoans, brachiopods, crinoids, and a few gastropods, blastoids, and cephalopods. Similar forms occur in the carbonate beds, though they are less abundant and not so well-preserved, probably owing to dolomitization. Diagnostic conodonts of Osagian age are reported by Racey (1974) and Ritter (1983). The Thunder Springs is everywhere conformable with the underlying Whitmore Wash Member. It is disconformable with the overlying Mooney Falls Member except in the extreme western end of Grand Canyon (Fig. 8.4). Locally the contact with the Mooney Falls is a low-angle unconformity, as at Kanab Canyon and Marble Canyon. This indicates minor structural activity, as well as erosion between Thunder Springs and Mooney Falls deposition.

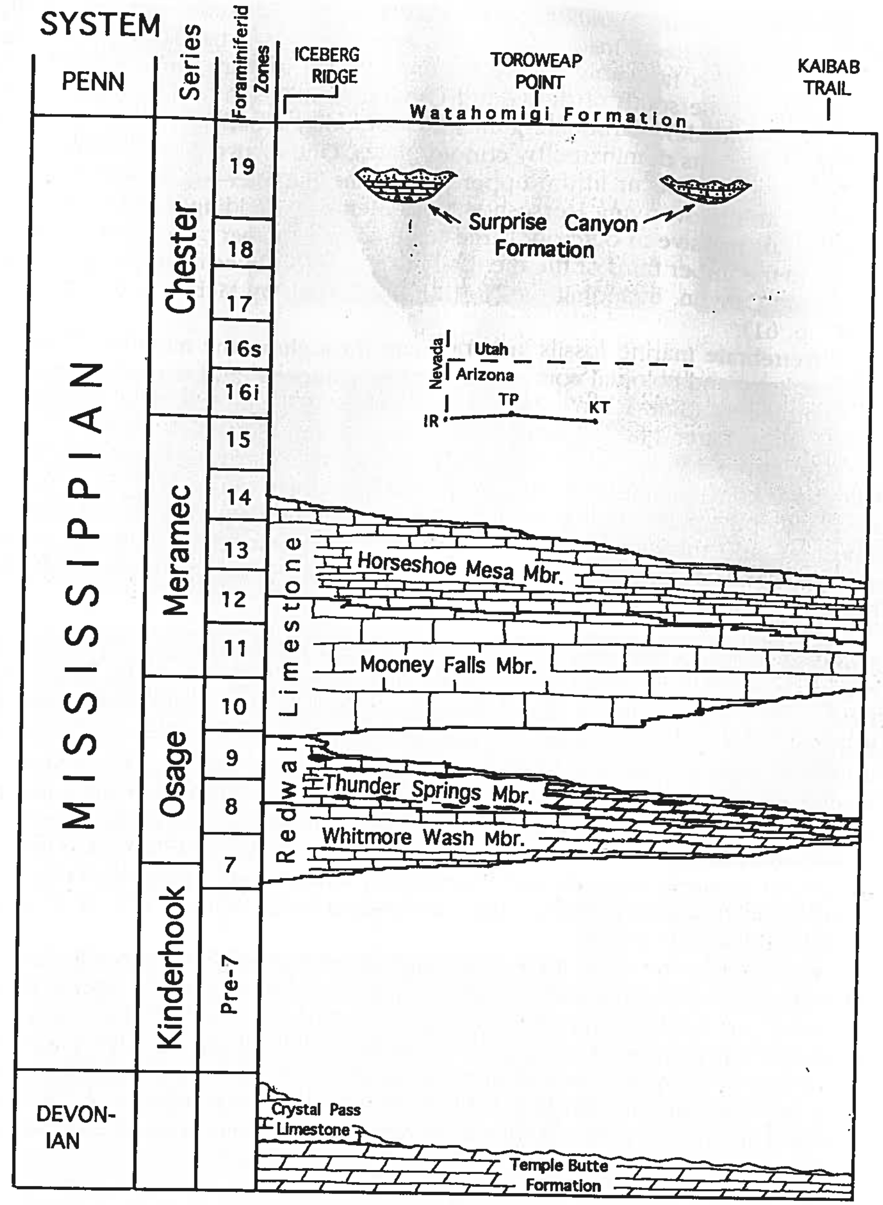

Figure 8.4. Diagram showing age and distribution of Mississippian rock units across northwestern Arizona. Foraminiferal zones after Mamet (Mamet and Skipp 1970). (Modified from Skipp 1979, Fig. 76.)

Mooney Falls Member The type section of the Mooney Falls Member is at Mooney Falls in Havasu Canyon (mile 153), about 4 miles (6.5 km) south of the Colorado River (McKee 1963). It is the thickest member of the Redwall, ranging from about 200 feet (60 m) in eastern Grand Canyon to nearly 400 feet (122 m) at the western end (McKee and Gutschick 1969, p. 56). It forms the major part of the sheer wall to which the Redwall name refers.

The Mooney Falls Member is predominantly pure limestone—except locally where it is dolomitized. Insoluble residue generally is less than 0.5 percent. The rock constitutes a favorable source for lime, which is being processed at two major plants (one south of the Grand Canyon near Peach Springs, and a second near Clarkdale). Carbonate grains include oolites, pellets, and a variety of skleletal fragments dominated by crinoid plates. One or two zones of thin beds or lenses of chert occur in the upper part of the member—generally near the contact with the overlying Horseshoe Mesa Member. Bedding is normally thick and appears massive in outcrop. Large-scale, tabular-planar cross-bedding is reported in the upper third of the member at several localities in central and eastern Grand Canyon, including the North Kaibab Trail, by McKee and Gutschick (1969, p. 61).

Invertebrate marine fossils are abundant throughout the member. They include solitary and colonial corals, spiriferid brachiopods, and crinoids. Diagnostic foraminifers (Skipp 1969; Mamet and Skipp 1970, p. 338) and conodonts (Racey 1974; Ritter 1983) indicate a late Osagian and Meramecian age.

The upper contact of the Mooney falls with the overlying Horseshoe Mesa Member is conformable and difficult to locate precisely in the field. The boundary is generally placed at the vertical change from medium- and coarse-grained limestone and thick or massive bedding of the Mooney Falls Member to the aphanitic and relatively thin-bedded, receding-ledge-forming limestone of the Horseshoe Mesa.

Horseshoe Mesa Member The type section for this member is on the south rim of Grand Canyon, along the Grandview Trail north of Horseshoe Mesa. The Horseshoe Mesa Member is the thinnest and least extensive member of the Redwall Limestone. Its thickness in the Grand Canyon varies from 45 to 125 feet (14 to 38 m), with the thinnest section in the east. Erosion causes the member to wedge out 30 to 40 miles (50 to 65 km) south of the Grand Canyon. It is also missing from the top of the Mooney Falls Member in most of central Arizona. The Horseshoe Mesa is composed of thin-bedded, light gray limestone with a mudstone to wackestone texture. Some chert lenses occur in the lower part. It typically forms weak receding ledges in contrast to the massive cliff of Mooney Falls below.

Well-preserved, invertebrate fossils are rare but present throughout the Horseshoe Mesa. Spiriferid brachiopods, bivalves, and corals are among the most abundant forms. At least 16 species of foraminifers are recognized in the member and indicate a Meramecian (early Late Mississippian) age (Skipp 1979, p. 298). The upper boundary of the Horseshoe Mesa Member is a major unconformity overlain by Early Pennsylvanian redbeds of the Watahomigi Formation, Supai Group. Locally it is overlain by lenses of the Surprise Canyon Formation of Late Mississippian age.

Paleontology and Age

Invertebrate fossils are common in certain lithofacies of the Redwall throughout the Grand Canyon (Fig. 8.5). Data from some 500 localities collected by McKee and Gutschick (1969) indicate that the most abundant megafossils are brachiopods and corals—followed by bryozoans, crinoids, bivalves, and cephalopods. Additional minor elements include blastoids, trilobites, ostracods, fish teeth, and algal remains. Foraminifers are abundant and were found in half of all samples selected for thin-sectioning.

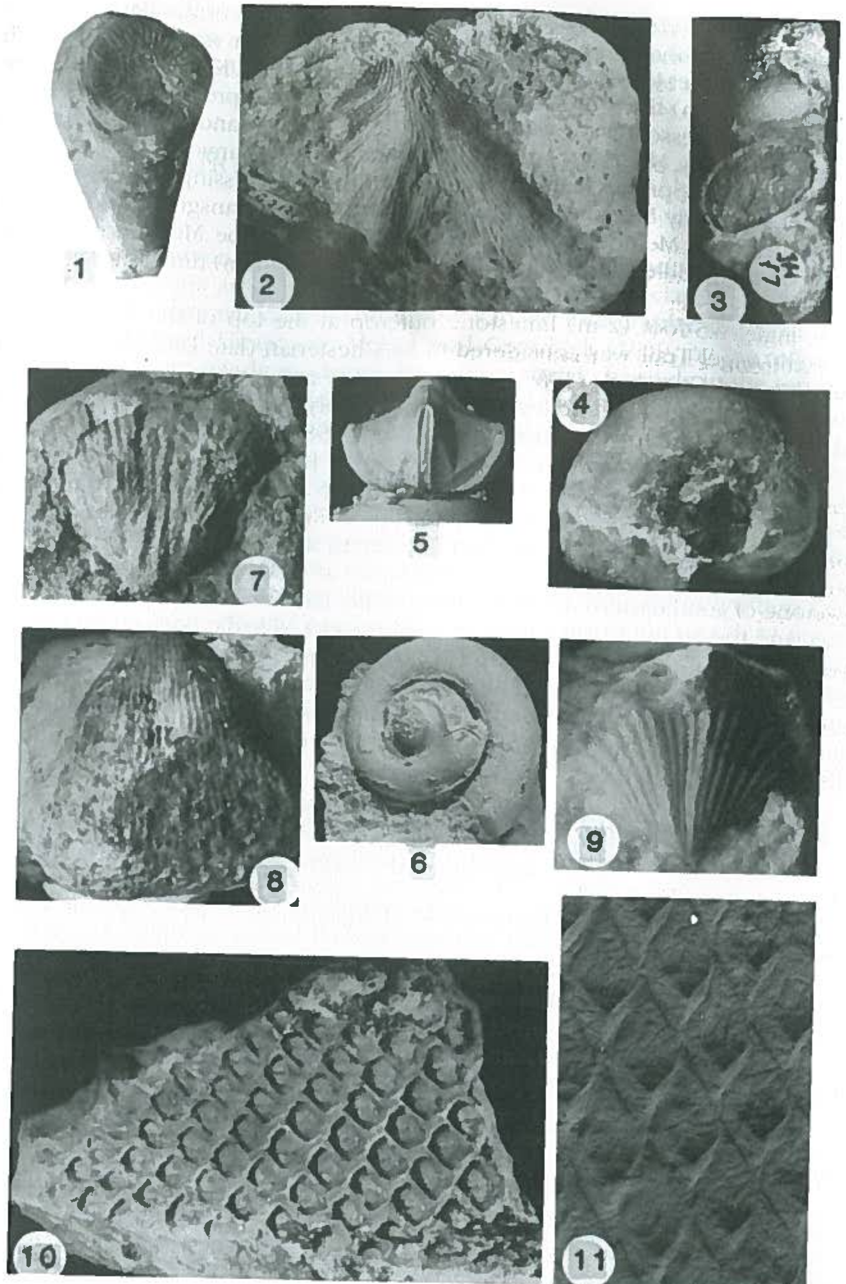

Figure 8.5. Invertebrate fossils from the Redwall Limestone (1–9) and plant fossils from the Surprise Canyon Formation (10,11). (1) *Zaphrentites,* a solitary coral; (2) *Fenestella,* a bryozoan; (3) *Loxonema;* (4) *Bellerophon;* (5) *Orophocrinus saltensis* Macurda, a blastoid; (6) *Straporollus,* gastropods; (7,8) *Buxtonia viminolis* (White); (9) *Anthracospirifer,* brachiopods; (10,11) two forms of *Lepidodendron.* All figures 3 1.

Index fossils permit relatively accurate dating of most of the Redwall Limestone. As Fig. 8.4 shows, initial Redwall deposition began with the basal Whitmore Wash Member in western Grand Canyon during latest Kinderhookian (early Early Mississippian) time. Basal Redwall deposits become progressively younger as the sea transgressed eastward across northern Arizona and are no older than Osagian (late Early Mississippian) age in eastern Grand Canyon.

The Thunder Springs Member was formed by a regressing sea during middle Osagian to early Meramecian time. A second marine transgression deposited the Mooney Falls Member of the Redwall. The Horseshoe Mesa Member was formed during middle Meramecian (early Late Mississippian) time as a regressive deposit.

A single, 6.5-foot (2-m) limestone outcrop at the top of the Redwall near the Bright Angel Trail was considered to be Chesterian (late Late Mississippian) by McKee and Gutschick (1969, p. 74). The age assignment, based upon rare brachiopod and foraminiferid fossils, is considerably younger than any other Redwall Limestone outcrop. Billingsley and Beus (1985, p. 27) treated this occurrence as part of the overlying Surprise Canyon Formation of Chesterian age. However, recent examination of the fossils McKee and Gutschick collected from this outcrop confirms that they are from typical Redwall, not Surprise Canyon, lithology and are probably older than Chesterian age.

The age assignment of the Redwall places it as a correlative of the Escabrosa Limestone of southeastern Arizona, the Leadville Limestone of southwestern Colorado, and the Monte Cristo Group of southeastern Nevada. Four of the five formations in the Monte Cristo Group (excluding only the Arrowhead Limestone) are nearly identical in lithology and stratigraphic position with the four members of the Redwall in Grand Canyon (McKee and Gutschick 1969, p. 14). It is likely that the Redwall Limestone deposits were laterally continuous with all the above units at the end of the Mississippian Period.

Depositional Setting

Deposition of the Redwall Limestone sediments occurred in a shallow, epeiric sea that produced a submerged continental shelf across northern Arizona. Deposits formed during two major transgressive–regressive pulses, as demonstrated by McKee and Gutschick (1969). Detailed facies analysis by Kent and Rawson (1980) and Bremner (1986) have confirmed and refined this interpretation.

The basal part of the Whitmore Wash Member records initial deposition during the first transgression under nearshore, shallow subtidal conditions where high-energy currents produced oolitic shoals. As the transgression proceeded, more offshore deposits of skeletal grainstone and packstone accumulated under quieter water and more open marine conditions.

The Thunder Springs Member accumulated in increasingly shallower conditions as the sea regressed westward. The abundant chert layers (which exhibit a lack of sorting, an original lime mud texture, and a high proportion of delicate bryozoan fossil fragments) are considered by Bremner (1986, p. 62) to be preferentially silicified blue-green algal mats, which may have locally baffled marine currents. Alternating with the chert beds are abraded, sorted, crinoidal grainstone and packstone deposits that record a more vigorous current-washing of skeletal sand.

The Mooney Falls Member of the Redwall formed during a second marine transgression as crinoidal packstone and grainstone sediments developed widely

across northern Arizona under generally open marine, offshore conditions. The Horseshoe Mesa Member formed under conditions of increasingly shallow and more restricted circulation during a final, slow regression of the sea.

SURPRISE CANYON FORMATION

Nomenclature

The Surprise Canyon Formation is a newly recognized rock unit in the Grand Canyon. It appears as isolated, lens-shaped exposures of clastic and carbonate rocks that fill erosional valleys and locally karsted topography and caves in the top of the Redwall Limestone. McKee and Gutschick (1969, p. 76) recognized conglomerate and gnarly mudstone beds filling local channels in the top of the Redwall Limestone but considered them part of the basal Supai Group. Subsequently these strata were recognized by Billingsley (1979) as a separate unit belonging to neither the Supai or the Redwall, and they were referred to as pre-Supai buried valley deposits by Billingsley and McKee (1982).

The name Surprise Canyon Formation was applied formally by Billingsley and Beus (1985, p. 27) and is taken from a large, northern tributary canyon in western Grand Canyon at mile 248. The type section (Fig. 8.6) is on the east-facing slope of a narrow ridge near the Bat Tower viewpoint in western Grand Canyon (mile 265)—about 12 miles (20 km) northwest of the mouth of Surprise Canyon. The Surprise Canyon Formation is probably the least visible rock unit in the Grand Canyon because of its discontinuous nature and the remoteness of the larger outcrops (Billingsley and Beus 1999).

FIGURE 8.6. Type section of the Surprise Canyon Formation, Bat Tower Section 2, is located about 1 mile (1.6 km) west of mile 263 on the Colorado River and 1.6 miles (2.6 km) southwest of the mouth of Tincanebits Canyon.

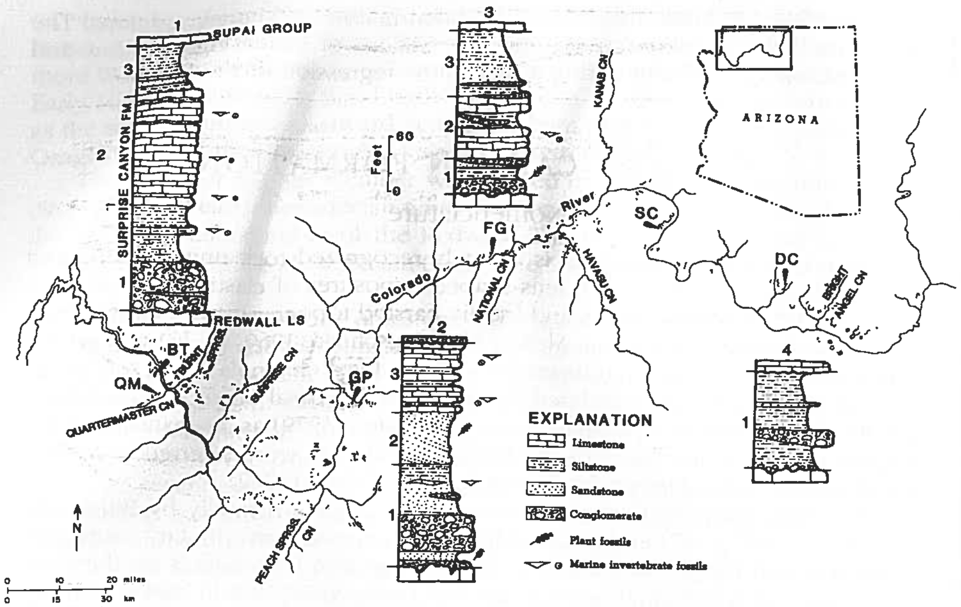

Figure 8.7. Index map showing major outcrops of the Surprise Canyon Formation (black patches) and selected stratigraphic sections. Section 1—type section near the Bat Tower (BT); section 2—Granite Park (GP); section 3—Fern Glen (FG); section 4—Dragon Creek (DC); the outcrop at Quartermaster Canyon (QM) is illustrated in Fig. 8.9.

Distribution

The Surprise Canyon Formation is nowhere a continuous stratum. Instead, it crops out as isolated, lens-shaped patches throughout much of the Grand Canyon and in parts of Marble Canyon to the east (Fig. 8.7). The valleys in which the formation occurs commonly are 150 to 200 feet (45 to 60 m) deep and up to 0.5 mile (1 km) wide in western Grand Canyon. They become shallower and relatively wider in central and eastern Grand Canyon (Fig. 8.8). Thickness of the formation corresponds to the depth of the valleys in which it occurs. The thickest section observed is at Quartermaster Canyon (Fig. 8.9), a southern tributary canyon in western Grand Canyon, where the formation is about 400 feet (122 m) thick (Billingsley and Beus, 1985, p. 27). Outcrops in central Grand Canyon are up to 50 feet (45 m) thick, whereas in eastern Grand Canyon and Marble Canyon they rarely are more than 60 or 70 feet (18 or 20 m) thick.

Redwall-Surprise Canyon Unconformity

The erosion surface that developed on the Redwall Limestone during the Late Mississippian Period, and on which the Surprise Canyon was deposited, must have been a relatively flat, resistant, limestone platform with considerable local relief. Most of the Surprise Canyon outcrops occupy gentle U-shaped or V-shaped notches cut into the top of the Redwall. By their nature and distribution these notches appear to have been part of a major dendritic drainage system that

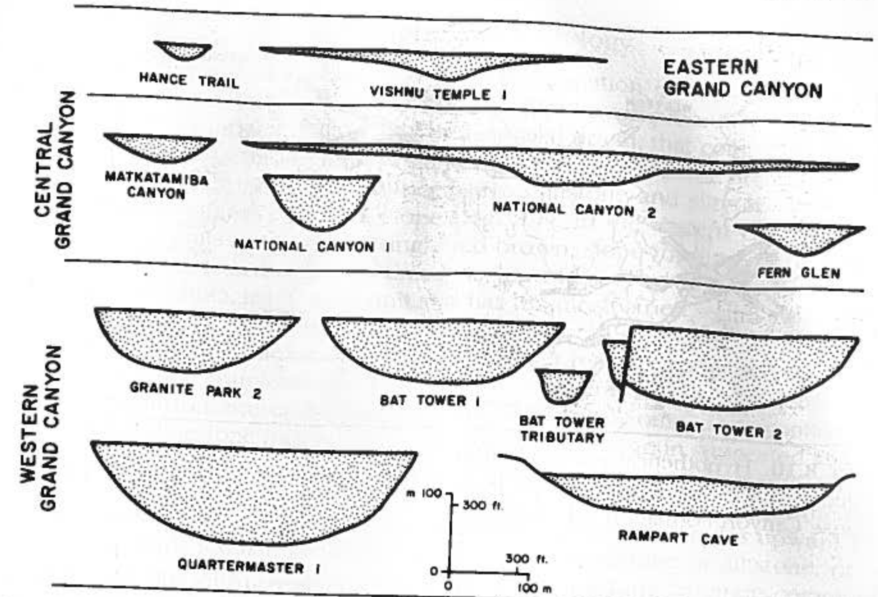

Figure 8.8. Cross sections of selected channel fill outcrops of the Surprise Canyon Formation, illustrating a general increase in thickness from eastern to western Grand Canyon.

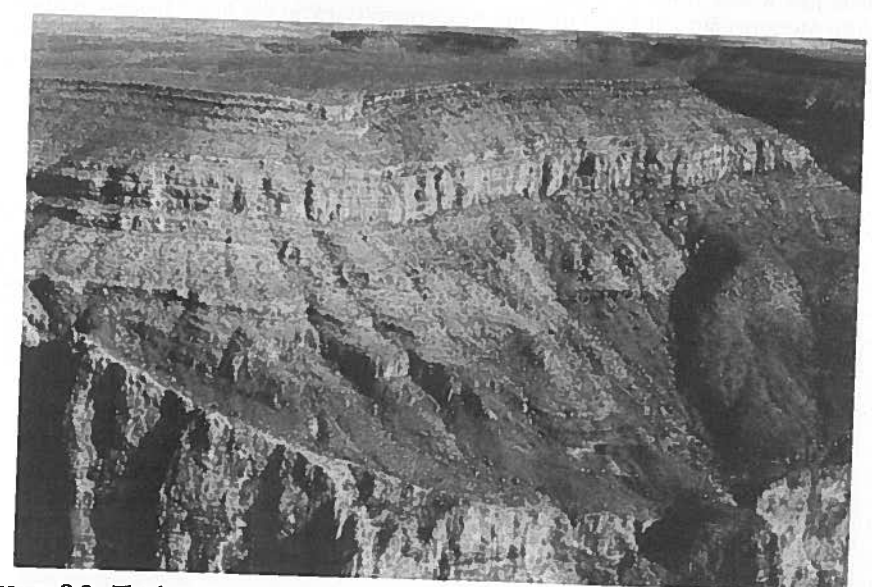

Figure 8.9. Thickest known exposure of the Surprise Canyon Formation on the west wall of Quartermaster Canyon (QM in Fig. 8.7). Note distinct curved surface of the pre-Surprise Canyon valley wall cut into the top of the Redwall Limestone.

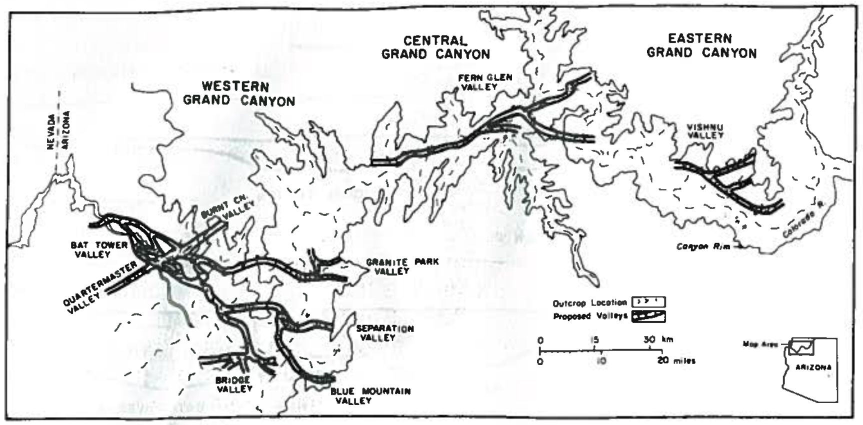

FIGURE 8.10. Hypothetical reconstruction of segments of the ancient valley system eroded into the Redwall Limestone in Late Mississippian time and onto which the Surprise Canyon Formation was deposited.

flowed generally from east to west. They were incised up to 400 feet (122 m) deep into the top of the Redwall Limestone in the western end of Grand Canyon. A preliminary reconstruction of the drainage pattern (Grover 1987) illustrates several major valleys that merge westward (Fig. 8.10). In addition, solution depressions and caves in the upper Redwall Limestone are filled locally with red mudstone of the Surprise Canyon Formation, indicating the development of this eroded and karsted topography prior to Surprise Canyon desposition (Fig. 8.11). The time available for the development of this topography on the top of the Redwall is just a few million years—the interval between the youngest Redwall (of middle Meramecian age) and the oldest Surprise Canyon (of late Chesterian age). The depth of the stream valleys cut into the top of the Redwall indicates either an uplift of the land surface or a drop in sea level of several hundred feet (120 m).

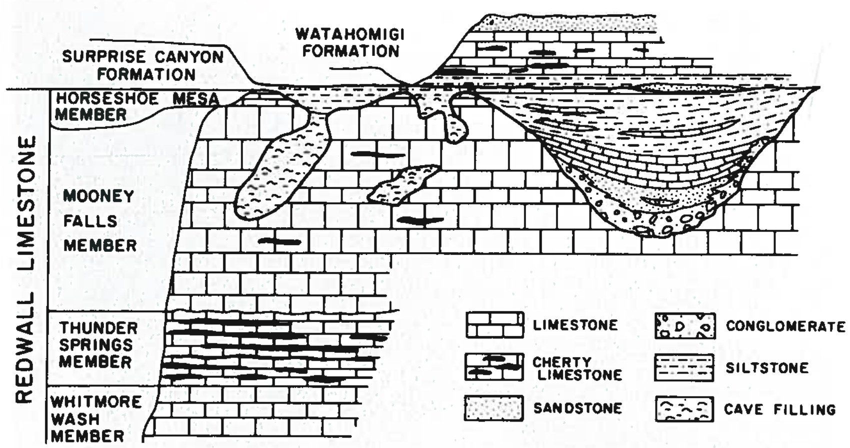

FIGURE 8.11. Cross section showing the stratigraphic relationship of the Surprise Canyon Formation to the underlying Redwall Limestone and overlying Watahomigi Formation of the Supai Group. (Modified from Billingsley and Beus 1985, Fig. 3.)

Stratigraphy and Lithology

The thicker sections of the Surprise Canyon Formation in western and central Grand Canyon can be divided into three major stratigraphic units: (1) a lower conglomerate and sandstone, mainly of terrestrial origin, that commonly forms a cliff and slope; (2) a middle unit of skeletal limestone of marine origin that commonly forms a cliff; and (3) an upper marine siltstone and silty or sandy limestone unit that typically forms a slope (Fig. 8.7). In the eastern Grand Canyon area the three units merge into a single red-brown, slope-forming conglomeratic sandstone and siltstone unit containing no limestone. The Surprise Canyon exhibits several lithofacies in each unit and has the most varied sedimentary lithology of any Paleozoic formation in Grand Canyon.

The basal part of unit 1 in most sections is a ferruginous pebble-to-cobble and local boulder conglomerate. Clasts are predominantly chert with minor limestone derived from the underlying Redwall Limestone. Some clasts contain typical Redwall Limestone fossils. The clasts commonly are grain supported and enclosed in a matrix of nearly pure quartz sand grains and some hematite. Locally the cobbles are sufficiently imbricated to indicate current directions at the time of deposition (Fig. 8.12). In most sections the conglomerate grades upward into a yellow to dark reddish brown or purple quartz sandstone or siltstone, or, in some sections, a dark, carbonaceous shale. The sandstone beds are commonly flat bedded, but some exhibit trough cross-strata or ripple laminations. *Lepidodendron* log impressions occur in the sandstone at numerous localities between Burnt Springs Canyon (mile 259.5) in western Grand Canyon and Cove Canyon (mile 169) in central Grand Canyon. Numerous plant fossils also occur in carbonaceous shale at Granite Park near mile 209. In eastern Grand Canyon, red-brown to purple mudstone beds, together with subordinate chert pebble conglomerate lenses typical of this lower unit, make up the entire formation. Trace fossils in sandstone beds of this unit are simple, vertical burrows and rare *Conostichus*, suggestive of the *Skolithus* ichnofacies of Crimes (1975).

Unit 2 is a coarse-grained, skeletal limestone that typically has a grainstone texture and is composed of whole or fragmented shells. Quartz sandstone beds

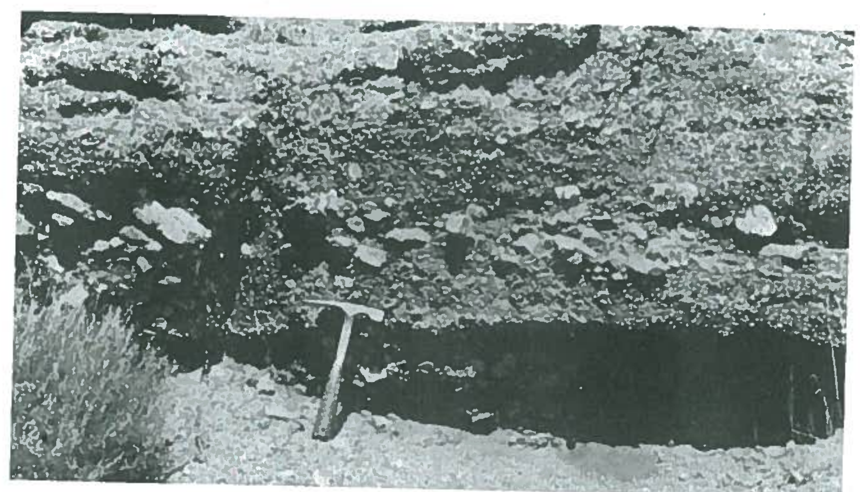

FIGURE 8.12. Imbricated pebbles and cobbles in conglomerate of the Surprise Canyon Formation, Dragon Creek section.

Figure 8.13. Algal stromatolites (oncolites) from near the top of the Surprise Canyon Formation at Quartermaster Canyon section 4. Centimeter scale.

up to 1.5 inches (3 or 4 cm) thick, alternating with skeletal limestone beds up to 4 inches (10 cm) thick, are common. The base of the limestone commonly truncates the unit 1 sandstone or siltstone beds occurring below on an erosion surface. The limestone unit typically forms resistant cliffs or ledges and weathers yellowish brown, rusty, or purple gray. Small-scale trough cross-strata exhibiting bimodal current directions are common. Marine invertebrate fossils are abundant and include crinoids and other echinoderms, brachiopods, bryozoans, corals, mollusks, and trilobites. Trace fossils typical of the shallow marine *Cruziana* ichnofacies of Crimes (1975) occur in sandy limestone beds. This unit is the thickest and topographically most prominent feature in many sections of central and western Grand Canyon. It is absent, however, in eastern Grand Canyon, east of Fossil Bay (about mile 130).

In western and central Grand Canyon, unit 3, the upper unit, is typically a dark red-brown to purple, ripple-laminated to flat-bedded calcareous siltstone of sandstone that forms weak slopes or receding ledges. Linguoid ripples are common. Resistant ledges of algal or ostracodal limestone are also common within the unit in western Grand Canyon. In at least three localities—Burnt Spring Canyon, Quartermaster Canyon, and National Canyon—nearly spherical algal stromatolites (oncolites) occur near the top of the unit (Fig. 8.13).

The boundary between the Surprise Canyon Formation and the overlying Watahomigi Formation of the Supai Group commonly is obscured by limestone rubble from above or by a covered slope developed on weak mudstone. Where well-exposed, the basal Watahomigi consists of: (1) a thin widespread, but locally discontinuous, limestone pebble conglomerate that contains minor chert clasts; or (2) where the conglomerate is absent, a purplish red calcareous siltstone and mudstone overlain generally by resistant gray limestone beds containing pale red-to-orange chert nodules (Billingsley and Beus 1985, p. 29). In a few localities a low-angle unconformity is recognizable at the contact (Fig. 8.14).

Figure 8.14. Low-angle unconformity at the Surprise Canyon–Watahomigi Contact (between the two arrows in the center of the photograph) about 0.3 mile (0.5 km) due west and on the opposite side of the ridge from the type section southeast of the Bat Tower.

Paleontology and Age

The fossil record of the Surprise Canyon Formation is one of the richest and most diverse of any Paleozoic unit in the Grand Canyon. It has yielded more than 60 species of marine invertebrates (Fig. 8.15) including foraminifers, conodonts, corals, bryozoans, brachiopods, echinoderms, mollusks, trilobites, and ostracodes (Beus 1995). Brachiopods are the most abundant forms preserved and are commonly associated with bryozoans, corals, and echinoderms. A reconstruction of a fossil community dominated by *Composita* and spiriferid brachiopods with associated echinoderms, bryozoans, corals, and mollusks typical of the unit 2 limestone beds in the Bat Tower area is shown in Figure 8.16.

Microfossil invertebrates are moderately abundant in the middle limestone unit, and some are present in the upper limestone beds of unit 3. Seven species of foraminifers were identified by Betty Skipp (Billingsley and McKee 1982, p. 144) Ten forms of conodonts have been identified from limestone beds in the middle and upper units of the Surprise Canyon (Martin 1992). Of these, *Adetognathus unicornis, Cavusgnathus unicornis, C. naviculus,* and *Gnathodus* spp. are the more abundant and significant forms.

The Surprise Canyon has also yielded a modest number of plant fossils (Fig. 8.5) mainly from sandstone and siltstone or shale beds of the lower unit in western Grand Canyon. Palynomorphs (spores) representing 22 species from the Granite Park area were identified by R. M. Kosanke (Billingsley and McKee 1982, p. 144). Several types of algal structures, and 12 species of plant megafossils, including three species of *Lepidodendron, Calamites,* and seed ferns (Tidwell et

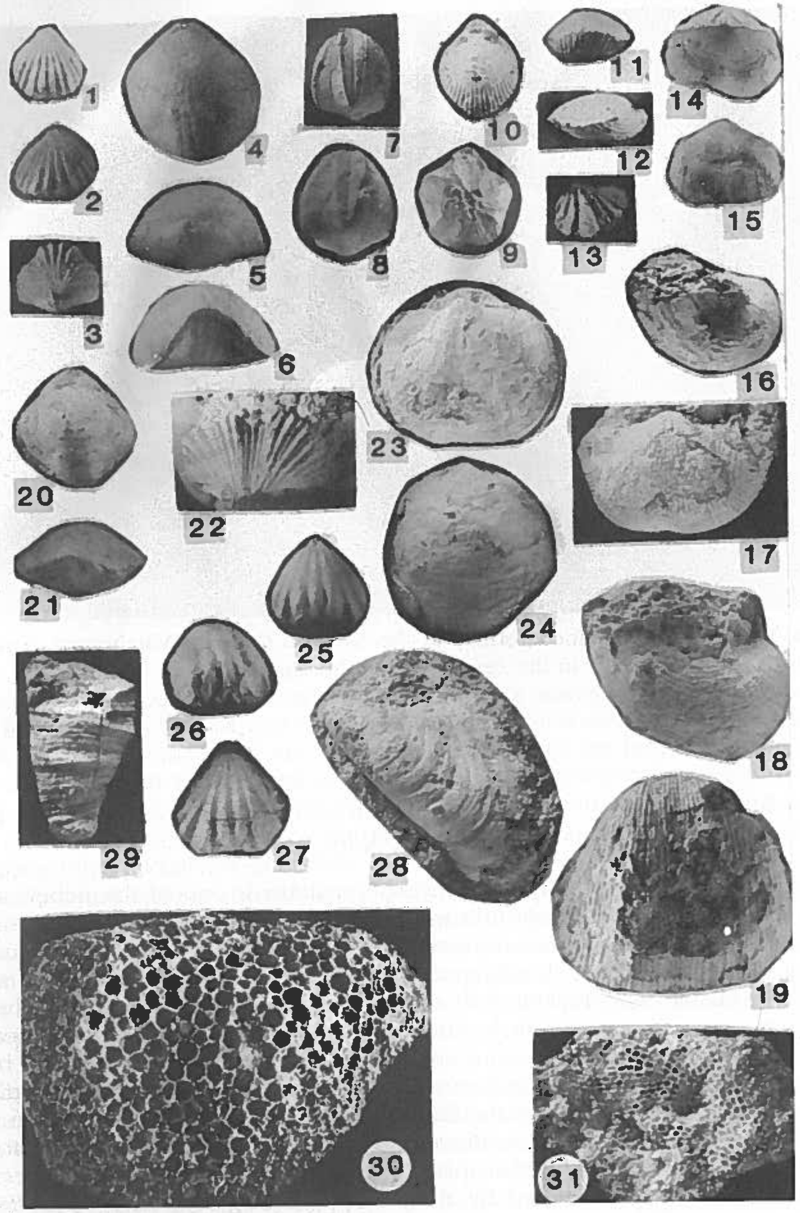

Figure 8.15. Invertebrate fossils from the Surprise Canyon Formation. (1–3) ?*Macropotamorhynchus* cf. *M. purduei* (Girty); (4–6) *Leiorhynchoidea carbonifera* (Girty), rhynchonellid brachiopods; (7–9) *Pentremites*, a blastoid; (10–12) *Eumetria*, (13) punctate spiriferid, brachiopods; (14,15) *Diaphragmus*, (16) *Inflatia*, (17–19) *Flexaria*, productid brachiopods; (20–21) *Composita*; (22) *Anthracospirifer*, (23) *Schizophoria*; (24) *Rhipidomella nevadensis* (Meek); (25–27) *Rotaia*, brachiopods; (28) *Septimyalina*, a bivalve; (29) *Barytchisma*, a solitary coral; (30) *Michelinia*, a colonial coral; (31) fenestellid bryozoan. All figures × 3/4.

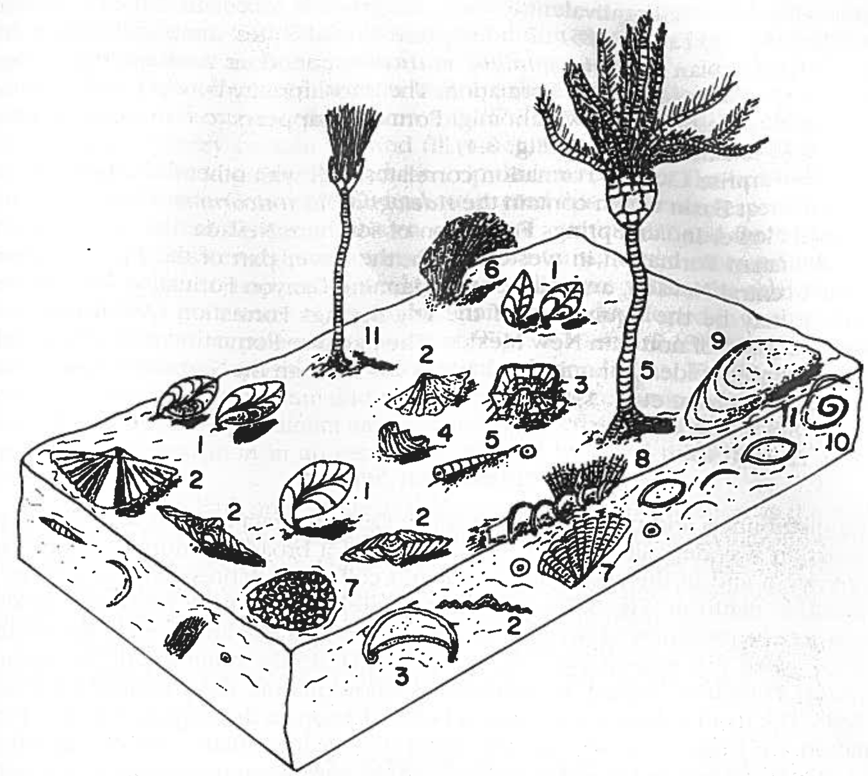

Figure 8.16. Reconstruction of a bottom-dwelling marine community typical of the middle limestone unit as it may have appeared in Late Mississippian time. Brachiopods are: (1) *Composita*; (2) punctate spiriferid; (3) *Inflatia*; and (4) *Ovatia*; (5) a crinoid; (6) fenestrate bryozoan; (7) the colonial coral *Michelinia*; (8) the bryozoan *Archimedes*; (9) the bivalve *Septimyalina*; (10) rare bellerophontid gastropod; (11) the blastoid *Pentremites*.

al. 1992), are known. Also bone fragments and some identifiable shark teeth are present in the middle unit.

Fossil evidence from spores, foraminifers, conodonts, brachiopods, and corals documents a Late Mississippian (Chesterian) age for the entire Surprise Canyon Formation. The brachiopod *Rhipidomella nevadensis* occurs through some 35 feet (10 m) of the middle limestone unit in the Blue Mountain Canyon area and marks the *Rhipidomella nevadensis* Assemblage Zone that Gordon (1984) considered coincident with foraminiferid Zone 19 in the western United States. Foraminiferids identified from limestone beds in the upper unit by Betty Skipp (Billingsley and McKee 1982, p. 144) include *Eosigmoilina explicata* that is restricted to formaniniferid Zone 19.

Conodonts provide the most precise data for age determination of the Surprise Canyon Formation. The upper Chesterian *Adetognathus unicornis* Zone is recognized in virtually all Surprise Canyon outcrops yielding conodonts. Although some zones barren of conodonts exist, taken collectively the occurrence of *Adetognathus unicornis* extends from the lowest limestone in the middle unit to the highest limestone in the upper unit of the formation. Thus the entire formation is considered equivalent to the *Adetognathus unicornis* Zone as defined by Baseman and Lane (1985) in the western United States. Strata containing the latest Mississippian *Rhachistognathus muricatus* conodont zone are not recognized in the Surprise Canyon Formation. The unconformity between the Surprise Canyon and the overlying Watahomigi Formation appears to represent the time interval of this missing zone (Fig. 8.4).

The Surprise Canyon Formation correlates well with other formations in the eastern Great Basin which contain the *Adetognathus unicornmis* Zone. These include the lower Indian Springs Formation of southern Nevada, the upper part of the Chainman Formation in western Utah, the lower part of the Ely Formation in west-central Nevada, and part of the Manning Canyon Formation in northern Utah. It may be the equivalent of the Log Springs Formation (Armstrong and Repetski 1980) of northern New Mexico. The Paradise Formation of southern Arizona is slightly older (foraminiferid zones 15–18) than the Surprise Canyon Formation (Armstrong et al. 1984).

Depositional Setting

The distribution and nature of the Surprise Canyon strata suggest that the entire formation was deposited within the confines of a broadly dendritic stream valley system and in the associated caves and collapsed depressions formed on a limestone platform (Fig. 8.17). The major shoreline must have been somewhere between the present western edge of the Colorado Plateau and Frenchman Mountain, located just east of Las Vegas, Nevada. Here, the outcrops of the Indian Springs Formation record an extensive shallow marine environment (Webster 1969). The marked lateral and vertical facies changes in the Surprise Canyon Formation are believed to record deposition in a major estuary system extending for at least 80 miles (130 km) east–west across northwestern Arizona to the eastern limit of marine fossils (Fig. 8.16). For the western and central Grand Canyon outcrops the depositional environment appears to have been fluvial for the basal unit. The strata become progressively more marine-dominated up section into the middle limestone and sandstone unit and end with the record of a restricted marine environment in the upper unit.

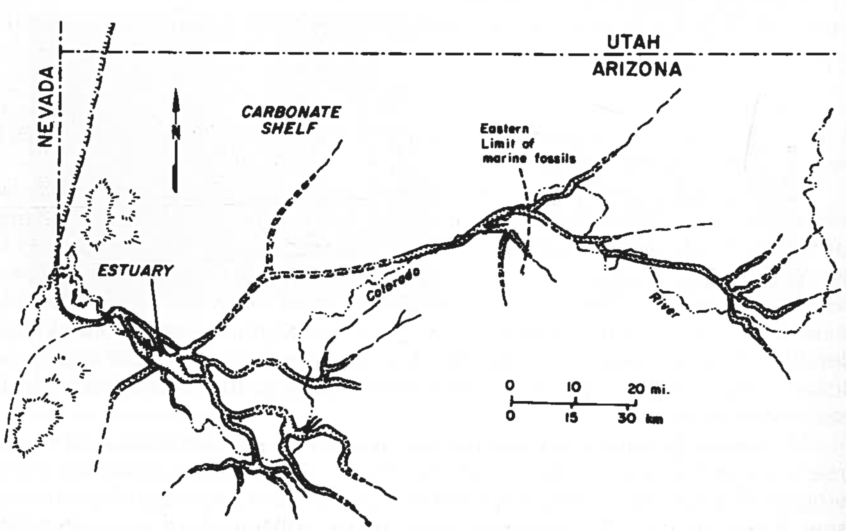

Figure 8.17. Hypothetical paleogeography of northern Arizona during Surprise Canyon Formation deposition in Late Mississippian time.

The basal conglomerate beds in unit 1 exhibit local imbrication that indicates strong, unidirectional, fluvial currents. The sandstone beds above the conglomerate commonly contain cut and fill structures and locally abundant plant remains. These beds grade upward into ripple-laminated or flat-bedded sandstones and siltstones. Grover (1987) has interpreted this sequence (especially well displayed in the Bat Tower area) as a record of continental and fluvial conditions changing to intertidal conditions as the sea transgressed eastward into the estuary and began to trap and rework clastic sediments within the tidal range.

The skeletal grainstone of the middle unit is well-developed in both the Bat Tower and Fern Glen paleovalley systems (Figs. 8.10 and 8.18). Abundant marine fossils attest to deposition in shallow marine conditions. The bimodal current directions (both upstream and downstream, but predominantly upstream or east), as recorded by prominent small-scale trough cross-strata in the Bat Tower area, suggest deposition in an estuary dominated by flood tides (Grover 1987) (Fig. 8.16).

In the Granite Park area, strata at the same stratigraphic position as the middle limestone unit are predominantly cross-stratified sandstone containing abundant plant fossils (section 2, Fig. 8.7; Fig. 8.19). Grover (1987) has interpreted this as deposition in a more fluvial- and ebb-tide-dominated valley where sand supply and deposition eclipsed minor marine limestone deposits.

For the most part, the upper unit (unit 3) of the Surprise Canyon Formation is ripple-laminated sandstone or siltstone, which alternates with algal and/or os-

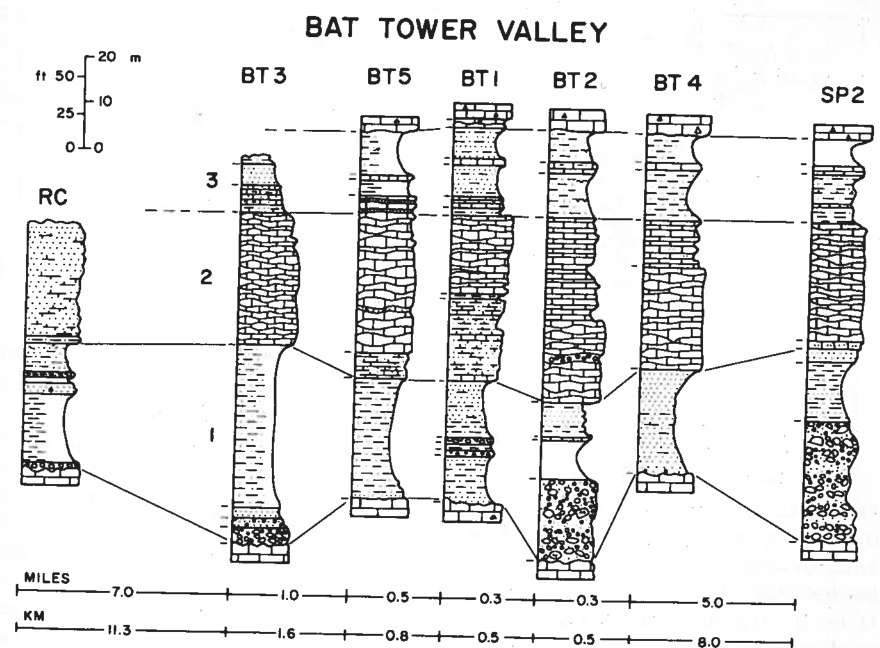

Figure 8.18. Stratigraphic sections of the Surprise Canyon formation along the Bat tower paleovalley portion of the estuary showing the maximum development of the unit 2 limestone. (From Grover 1987.)

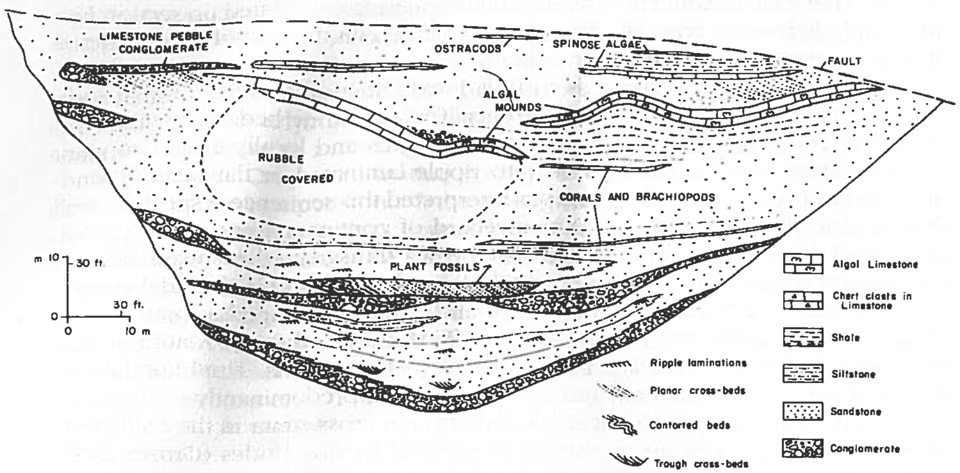

FIGURE 8.19. Sketch of a major part of the exposed lens of Surprise Canyon Formation at the Granite Park 2 section about 3.5 miles (5.6 km) southeast of the mouth of Granite Park Wash and mile 209.

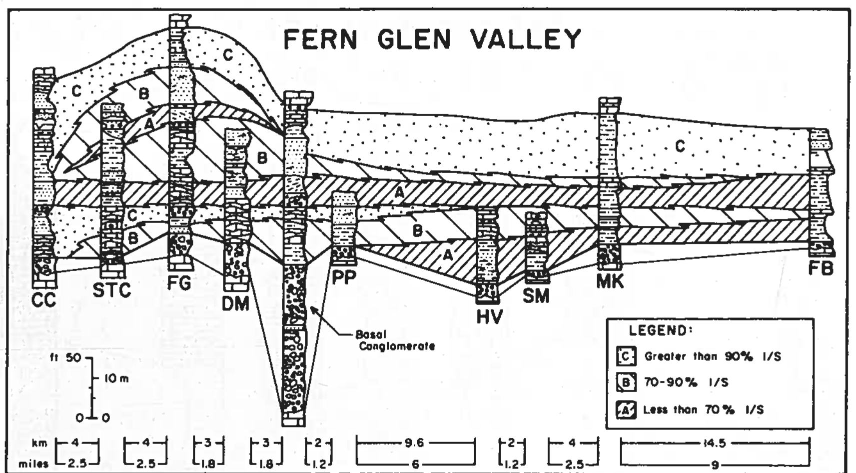

FIGURE 8.20. Selected stratigraphic sections of the Fern Glen paleovalley portion of the estuary, showing clay mineralogy facies distribution and environmental interpretations. Correlation is attempted only on fine-grained lithologic units. Facies A, having less than 70% illite/smectite clays, is considered to record nonmarine conditions; facies B, containing 70–90% illitic clays, is considered to record transitional or intermediate conditions; facies C, greater than 90% illitic clays, is considered to mark marine conditions. Two cycles of marine transgression and regression are indicated. (From Shirley 1987, Fig. 29.)

tracodal limestone beds that lack a normal, open marine fauna. Grover (1987) has interpreted these strata as a record of restricted-marine to upper tidal-flat environments developed during final infilling of the estuary. Detailed sedimentary and paleontological analysis of sections in the Fern Glen paleovalley indicate several fluctuations of marine and terrestrial or, at least, marginal marine conditions during deposition.

It is also possible that the original Surprise Canyon strata were part of a more widespread sheet formed in a shallow sea that covered much of northwestern Arizona. Subsequent uniform erosion then may have removed all but the lower valley-fill portion of the deposits. This would more easily explain the unusually great lateral extent of marine fossils [80 miles (130 km) east–west] within a narrow channel. However, in the absence of convincing evidence for marine deposition outside the confines of the narrow paleovalleys that presently contain the formation, deposition confined to a narrow estuary system seems the most reasonable interpretation.

Independent environmental interpretation based upon mineralogical and geochemical analysis of clays in the generally structureless and unfossilferous mudstones of the Surprise Canyon permits confirmation and refinement of the above interpretation. Differential flocculation of clays in the zone of sea water/fresh water mixing within the Fern Glen paleovalley section of the estuary appears to have produced lateral distribution of kaolinite predominantly to the east (the landward direction) and illite to the west (seaward direction) (Shirley 1987, Fig. 23a–j) as in modern estuaries (Edzwald and O'Melia 1975). Using ratios of kaolinite/illite clays, Shirley (1987) has demonstrated a convincing record of two marine transgression–regression cycles in the Fern Glen paleovalley (Fig. 8.20).

The depositional environment of the fine-grained, red mudstones and local conglomerates of the Surprise Canyon Formation in easternmost Grand Canyon and Marble Canyon must have been mainly fluvial—in fresh, or perhaps brackish, water conditions. Almost totally absent are limestone beds and marine fossils, and only a few plant fossils have been recovered. Imbricated cobbles in the basal conglomerate of the Dragon Creek section (Fig. 8.12) indicate a vigorous westward-flowing current at the time of deposition.

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