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CAMBRIAN AND ORDOVICIAN DEPOSITIONAL SYSTEMS IN ARIZONA

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ABSTRACT

Cambrian and Ordovician rocks in Arizona were deposited in a variety of continental and marine environments and record numerous transgressions and regressions of the seas. Throughout the Cambrian and possibly the Ordovician a north-south strandline migrated from west to east resulting in deposition of coarse clastics on the craton to the east and fine clastics and carbonates in more offshore but not always deeper waters to the west. The overall eastward transgression was interrupted by a number of regressive phases.

The basal Cambrian deposits, the Tapeats Sandstone and Bolsa Quartzite, were deposited in fluvial, beach, and inter- and subtidal settings. The Bright Angel Shale accumulated on a gently sloping marine shelf through migration of low-relief sand sheets and sand waves and emplacement of storm deposits. The carbonate-dominated Muav Limestone and Abrigo Formation formed in subtidal and intertidal areas offshore where carbonate production resulted in upward-shoaling sequences. Ordovician deposits are confined to southern Arizona and are poorly studied. Preliminary analyses of Ordovician strata suggest subtidal and intertidal deposition influenced by both storm and fair-weather processes.

Controls on strandline migration are uncertain. Comparisons with models developed for upper Precambrian and lower Cambrian strata to the west and also the role of varying rates of sea-floor spreading associated with the development of the western Cordilleran margin suggest that extrabasinal factors were probably of considerable importance.

INTRODUCTION

During the past 20 years, only a few studies have been directed at understanding the depositional systems and paleogeography of Cambrian and Ordovician strata in Arizona. Stratigraphic relationships in southeastern Arizona have been discussed by Hayes (1972, 1978) and Hayes and Cone (1975). McKee and Resser (1945) provided important data concerning Cambrian stratigraphy of the Grand Canyon region in northern Arizona. Sedimentology and regional paleogeography were addressed by Hayes (1972, 1978) and by Wanless (1973, 1975), Hereford (1977), Middleton and Hereford (1981), and Martin (1985). These studies have increased our understanding of these strata, nevertheless, more detailed sedimentologic and petrologic studies are needed to fully reconstruct the depositional systems and to understand the nature of the early Paleozoic margin of the North American craton.

Cambrian and Ordovician strata crop out discontinuously throughout Arizona. The thickest and most continuous exposures of Cambrian rocks are in northern Arizona along the Colorado River, with scattered and incomplete sections occurring in northwestern and central Arizona. Discontinuous and thinner sections of Cambrian and Ordovician rocks occur throughout southeastern Arizona. In the Grand Canyon the formations are flat lying and exhibit a persistent tabular geometry, forming a series of ledges and slopes. In southern Arizona, Cambrian and Ordovician rocks are deformed where they are associated with Tertiary metamorphic core complexes and other Laramide and Tertiary structures.

This paper has three objectives: (1) to review the stratigraphic relationships of Cambrian and Ordovician rocks in Arizona; (2) to present new data about the depositional and paleogeographic settings; and (3) to discuss the nature of and controls on transgressive and regressive sequences and relate these to the development of the Cordilleran miogeocline and craton during the early Paleozoic.

CAMBRIAN STRATIGRAPHY

Stratigraphic nomenclature for Cambrian deposits in Arizona is summarized in the seminal works of McKee and Resser (1945), Lochman-Balk (1971), and Hayes and Cone (1975). The framework for the following discussion is derived from these studies as well as from other mapping and stratigraphic works (Stoyanow, 1936; Sabins, 1957; Epis and Gilbert, 1957; Epis, 1958; Krieger, 1961, 1968; Hayes and Landis, 1965; Hayes, 1972; Chaffee, 1974). Stratigraphic terminology differs between the Basin and Range sections in southern Arizona and the Colorado Plateau-Transition Zone sequences to the north, and each will be discussed separately.

Southern Arizona

Bolsa Quartzite. The Bolsa Quartzite was named by Ransome (1904) for exposures in the Mule Mountains of Cochise County, Arizona (fig. 1). The formation unconformably overlies a variety of Precambrian units such as the Pinal Schist, Dripping Spring Quartzite, and Troy Quartzite and is considered to be Middle Cambrian (fig. 2) based on its gradational contact with the overlying late Middle Cambrian Abrigo Formation (Hayes and Cone, 1975). The Bolsa is between 3 and 250 m thick; this variation is a function of topographic relief on the underlying Precambrian surface (Cooper and Silver, 1964; Chaffee, 1974; Hayes and Cone, 1975).

The Bolsa is a white to brown or reddish-brown ledge-forming quartz arenite. Feldspar rarely constitutes more than 10 percent of the framework fraction and becomes less abundant up section. Sandstone at the base is typically medium- to coarse-grained, although thick beds of conglomerate occur at several localities (Cooper and Silver, 1964). The upper part is medium- to fine-grained sandstone with interbeds of siltstone and shale. A decrease in bedding thickness accompanies the upward decrease in grain size. Sedimentary structures include planar-tabular cross-stratification and horizontal stratification with the latter more abundant near the top of the formation. Hayes and Cone (1975) reported mud cracks from several localities as well as *Skolithos* tubes and other trace fossils in the upper part of the formation.

Abrigo Formation. The Abrigo Formation, named by Ransome (1904), is a carbonate-dominated sequence. Hayes and Landis (1965) remeasured the type section of the Abrigo in the Mule Mountains and subdivided it into four members. These are, in ascending order, a lower siliciclastic mudstone, a middle siliciclastic-carbonate sequence, an upper dolomitic sandstone and sandy dolostone, and the Copper Queen Limestone (fig. 2).

The lower member is late Middle Cambrian and is from 70 to 200 m thick (Hayes, 1978). It consists of interbedded bioturbated silty shale and limestone in the south grading into siltstone and fine-grained sandstone to the north (Hayes and Cone, 1975). Cross-stratification and trace fossils are common features at several localities. Carbonate facies include intraformational conglomerate, algal grainstone, peloidal packstone, and rare mudcracked wackestone (Hayes and Cone, 1975).

The middle member is from 42 to 94 m thick, although its absence in many areas is due to post-Cambrian erosion (Hayes, 1978). This member is probably early Late Cambrian (Dresbachian), although in the Mule Mountains and the northern Swisshelm Mountains a late Middle Cambrian (*Bolaspidella* zone) age is indicated (Hayes and Landis, 1965). This member consists of thinly bedded limestone and dolomitic limestone in the south, which grade northward into sandstone and dolostone. At the northern limit of its exposure the member is a fine-grained quartz arenite (Hayes and Cone, 1975). Intraformational limestone conglomerate and mudcracked carbonate mudstone are common, as are trace fossils and some cross-stratification (Hayes, 1978).

The upper sandy member is Late Cambrian (Dresbachian) and varies from 30 to 55 m thick. The sandy member grades eastward into the Coronado Sandstone. This member consists of dolomitic sandstone and sandy dolostone, with dolomite increasing toward the south and west (Hayes and Cone, 1975). Limestones are rare and typically are grainstones and packstones. Planar-tabular cross-stratification and intraformational dolomitic conglomerates are common.

The Copper Queen Limestone Member is present only in the easternmost areas and is up to 45 m thick. At the type section in the Mule Mountains it consists of thinly bedded, sandy to silty limestone (Hayes, 1972; Hayes and Landis, 1965; Hayes and Cone, 1975). Elsewhere, it is a sandy dolostone. The Late Cambrian (Franconian) Copper Queen gradationally overlies the upper sandy member and is conformably overlain by the El Paso Limestone in southeastern Arizona.

Coronado Sandstone. The Coronado Sandstone crops out in southeastern Arizona and varies from 110 to 190 m thick. Stratigraphic relationships indicate a late Middle to early Late Cambrian age (Hayes and Cone, 1975). The Coronado gradationally overlies the Abrigo Limestone in some areas and is conformably overlain by the El Paso Limestone. At the type locality near Morenci (Lindgren, 1905) it consists of feldspathic and quartz arenites. The formation becomes finer grained toward the south. Planar-tabular and trough cross-stratification and horizontal and vertical trace fossils are locally abundant.

Northern Arizona

Tapeats Sandstone. The Tapeats Sandstone is the basal formation of the Tonto Group (fig. 3) (Gilbert, 1874; Noble, 1914) and crops out almost continuously through the Grand Canyon and in scattered areas from Lake Mead southeast across central Arizona to Roosevelt Lake; it usually forms a series of ledges and recesses. The formation is Early Cambrian in the western Grand Canyon and Middle Cambrian in the eastern areas (McKee and Resser, 1945: Lochman-Balk, 1971). Thickness of the Tapeats is from 0 to 110 m; this variability is a function of relief of the underlying Precambrian crystalline and sedimentary rocks in northern Arizona or related to pre-Devonian erosion in central Arizona. Contact with the overlying Bright Angel Shale is gradational (fig. 4).

Lithologically the Tapeats is a medium- to coarse-grained feldspathic and quartz arenite. Feldspathic sandstone and small-pebble conglomerate are common at the base and decrease in abundance upwards in the formation. Planar-tabular and trough cross-stratification, horizontal and low-angle stratification, and vertical burrows are common features. Vertical burrows are common especially near the top of the formation.

Bright Angel Shale. The Bright Angel Shale is over 135 m thick in the western Grand Canyon, thinning markedly toward the east (McKee and Resser, 1945). Although the formation typically overlies the Tapeats Sandstone, the Bright Angel unconformably rests on Precambrian rocks where basement relief is extreme. Thickness variations are also the result of complex intertonguing of the Bright Angel with the overlying Muav Limestone (fig. 3). Like the Tapeats Sandstone, the Bright Angel becomes younger toward the east (fig. 3). In the western Grand Canyon, the base of the formation lies below the *Olenellus-Antagmus* assemblage zone and hence is late Early Cambrian, whereas in the eastern canyon the upper two-thirds of the Bright Angel lies above the *Alokistocare-Glossopleura* zone and is Middle Cambrian.

Lithologically, the formation consists of sandstone, siltstone, and shale. Although sandstones are more common near the base, they occur at several horizons within the Bright Angel (McKee and Resser, 1945; Martin, 1985). In addition to sand- and silt-size grains of quartz, feldspar, and sedimentary rock fragments, glauconitic grains and hematitic ooids are also common. Sedimentary structures are diverse and include trough and planar-tabular cross-stratification, horizontal stratification, and abundant ripple cross-stratification. Trace fossils are abundant and include a diverse array of horizontal and vertical burrows as well as tracks and trails (Martin, 1985).

Muav Limestone. The Muav Limestone forms resistant cliffs in the Grand Canyon. Contact with the underlying Bright Angel Shale is gradational and characterized by complex intertonguing of the two formations (fig. 3). Because of this intertonguing the Muav varies greatly in thickness throughout the region, although it generally thins towards the east (fig. 3). McKee and Resser (1945) reported over 250 m in the Grand Wash Cliffs near Lake Mead, 134 m at Toroweap in the central canyon, and only 41 m at the confluence of the Colorado and Little Colorado Rivers in Marble Canyon. The Muav also becomes younger to the east. In the western canyon the Muav lies above the *Alokistocare-Glossopleura* zone and is Middle Cambrian, whereas to the east the upper part of the Muav contains the *Bathyuriscus-Elrathina* (late Middle Cambrian) zone.

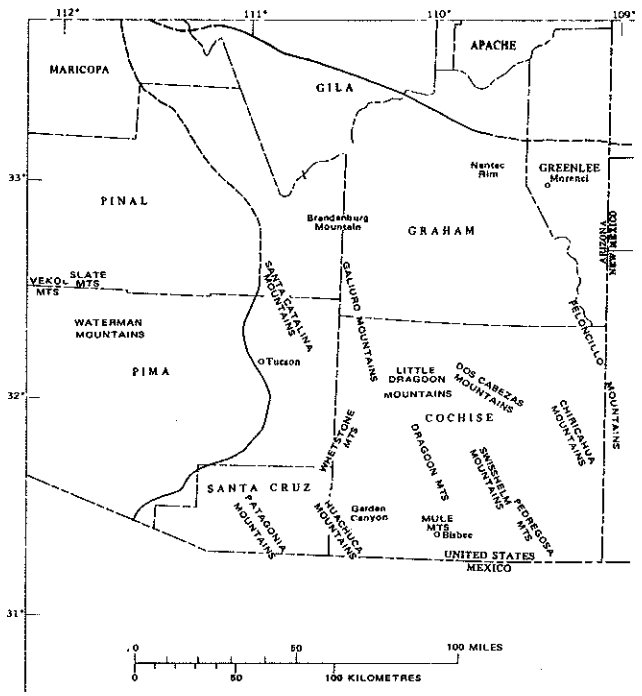


Figure 1. Location map of southeastern Arizona where Cambrian and Ordovician strata are exposed.

The Muav consists of thin to thick beds of dolomitic packstone and mudstone and nodular limestone. Persistent beds of intraformational limestone conglomerate were used by McKee and Resser (1945) to subdivide the Muav into seven members. Billingsley and others (1987) have documented an erosional unconformity between the basal Rampart Cave Member of the Muav Limestone and the upper Flour Sack Member of the Bright Angel Shale in the western Grand Canyon. In other areas the contact appears gradational. Thin beds of micaceous and dolomitic mudstone and minor sandstone occur at several horizons where they form minor recesses in the Muav cliffs.

Undifferentiated Cambrian. In the western part of the Grand Canyon, a thick sequence (up to 130 m at the Grand Wash Cliffs) of thin- to thick-bedded dolostone termed the "undifferentiated dolomites" overlies the Muav Limestone (McKee and Resser, 1945). McKee and Resser (1945) considered this unit to be Upper Cambrian, although paleontologic evidence is lacking. Wood (1956) proposed the term "supra-Muav" for this unit.

McKee and Resser (1945) recognized three lithofacies in the undifferentiated dolomite: white to buff, massive dolostone; white to yellow, very fine grained, thin-bedded dolostone; gray, fine-grained, thick-bedded dolostone. Sedimentary structures include wavy and asymmetric

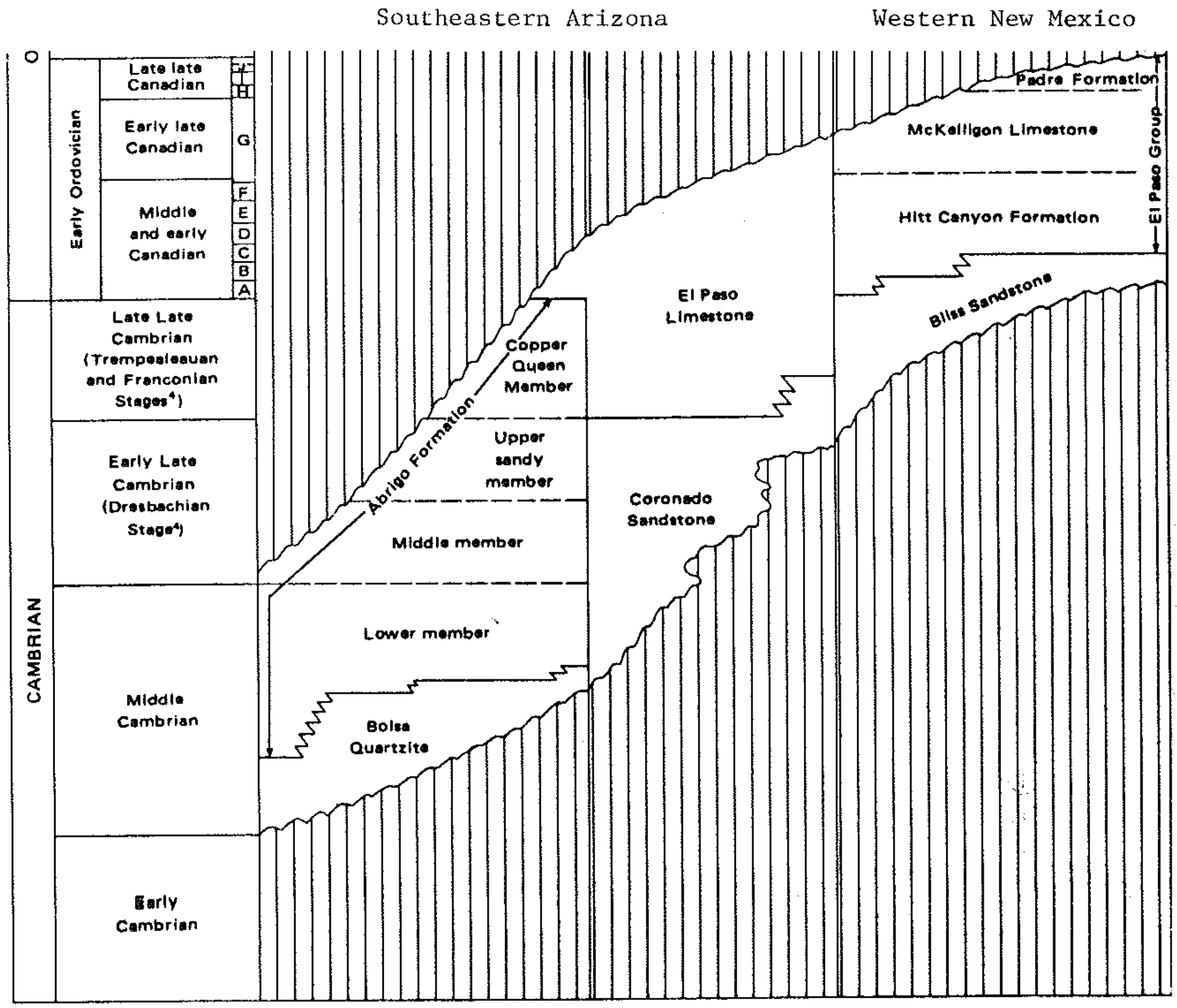


Figure 2. Age and stratigraphic correlation of Cambrian and Ordovician units in southern Arizona (after Hayes and Cone, 1975).

ripple laminations. Biogenic structures include a variety of horizontal traces and stromatolitic beds. Dolomitized oolitic grainstones are abundant locally.

CAMBRIAN DEPOSITIONAL SYSTEMS

In northern Arizona several sections of Cambrian strata have been studied in detail (Wanless, 1973, 1975; Hereford, 1977; Middleton and Hereford, 1981; Martin, 1985), although much work remains to be done. In southern Arizona, regional stratigraphic relationships have been established (Hayes and Cone, 1975); however, comparatively little sedimentologic work has been published. Accordingly, the sedimentology of the Cambrian of northern Arizona will be discussed and, where possible, comparisons made with equivalent strata to the south.

Northern Arizona

Tapeats Sandstone. Although McKee and Resser (1945) interpreted the Tapeats Sandstone in the Grand Canyon as having been deposited in nearshore areas, details of the lateral distribution of marine subenvironments were not documented. These workers interpreted the Tapeats as representing the deposits of a shallow-marine sand complex influenced by tides as well as fair-weather processes.

Islands of Precambrian rocks were common along the coastline and were important in controlling sedimentation. Large, angular blocks of Precambrian crystalline and sedimentary rocks occur in the basal Tapeats in the central part of the canyon. These blocks were eroded from the islands by waves. The role of storms in generating these breccias was documented for Cambrian strata in the mid-continent by Dott (1974), and a similar origin is likely for the Tapeats. Studies by Graham and Suttner (1974) and Middleton and others (1980) on similar deposits in Montana and Wyoming have demonstrated the importance of these features in controlling facies distributions in Cambrian strata. In particular, the possible effects of tidal amplification between islands and resulting sedimentation in interisland areas should be examined. The complexities of the coastal sedimentation and strandline configuration in the Cambrian are indicated by major embayments in the coastline, such as reported by McKee and Resser (1945) for the Phantom Ranch area of the central Grand Canyon.

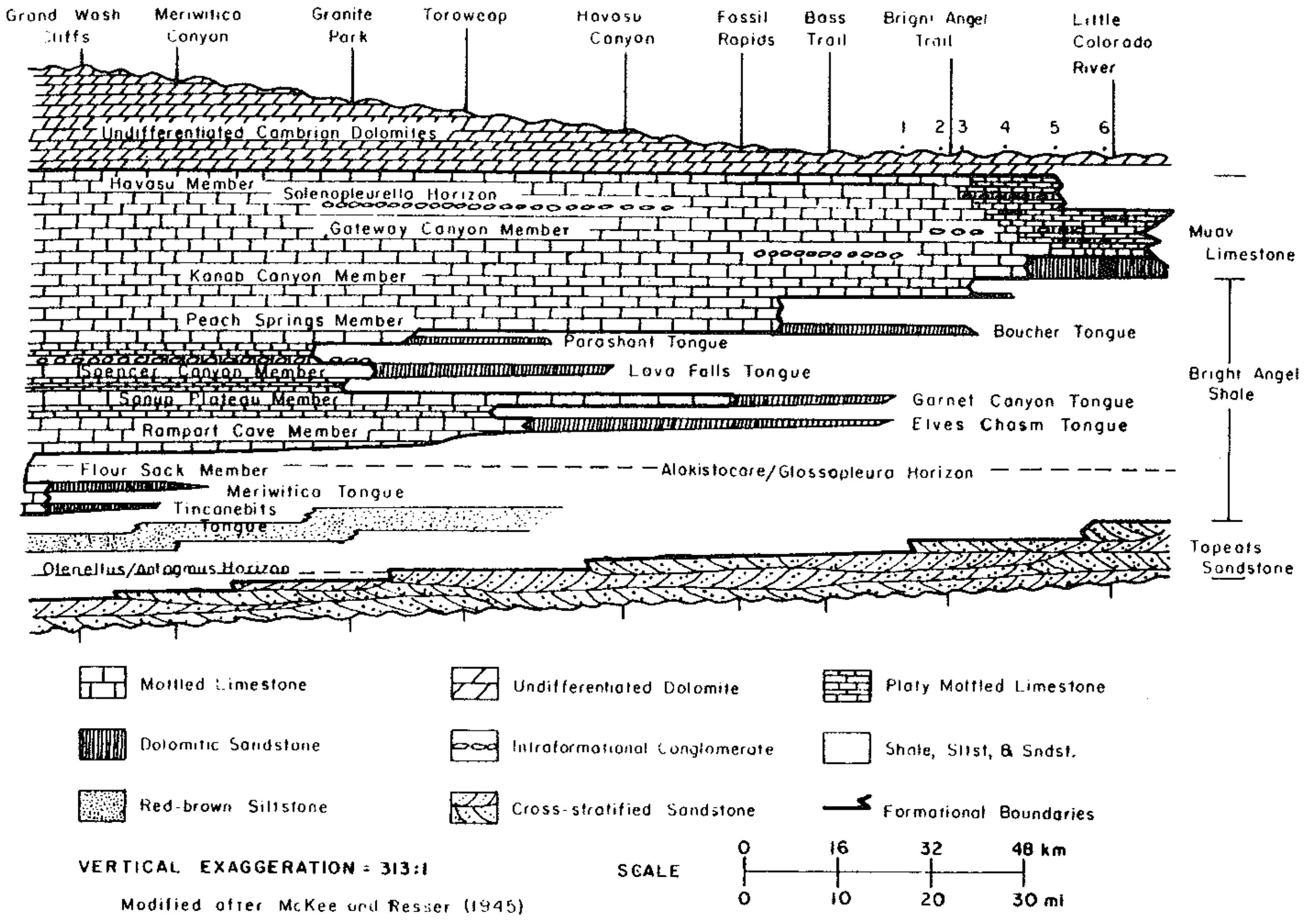


Figure 3. Stratigraphy of Cambrian rocks in the Grand Canyon (from McKee and Resser, 1945).

The most detailed sedimentologic study of the Tapeats Sandstone is that of Hereford (1977). This study concentrated on exposures in central Arizona from the Chino Valley area southeast to Payson. Hereford (1977) recognized and mapped six lithofacies. The facies were defined on the basis of type and scale of primary sedimentary structures, presence or absence of trace fossils, and topographic expression. Most of the Tapeats was deposited in shallow marine waters apparently influenced by significant tidal currents. Hereford (1977) recognized a number of distinct intertidal and shallow subtidal environments, many of which were dissected by tidal channels. Beach deposits were common around the margins of many of the islands of Precambrian rocks. Similar deposits have been reported from other lower Paleozoic sequences (e.g., Swett and others, 1971; Middleton and others, 1980).

Fluvial deposits have also been reported from the basal part of the Tapeats (Hereford, 1977; Middleton and Hereford, 1981). These are less mature texturally and mineralogically than associated marine deposits, and this reflects lack of extensive reworking of sediment along the Cambrian coastline (Odom and others, 1976; Middleton and others, 1980). Broad, shallow troughs are common in the fluvial deposits. Pebble- and granule-size detritus line the bases of many troughs; siltstone and shale lenses occur near the top of the channel fill. These sediments were deposited in very shallow and broad braided channels. In other parts of the sequence, up to 1-m-thick sets of planar-tabular and complex trough cross-stratification overlain by horizontally stratified sandstone and conglomerate occur (fig. 5) and represent the deposits of slightly sinuous transverse bars similar to those reported by Cant and Walker (1976) and Miall (1977) from other ancient braided river deposits. Such sequences are reported from many modern braided river systems (Ore, 1964; Bluck, 1976; Smith, 1970, 1971).

Although no comparable studies have been made of the Bolsa and Coronado Sandstones, cursory examination of these units indicates that these formations were deposited in similar settings. Bryant (1978) has reported tidal deposits from both the Bolsa and the Coronado.

Bright Angel Shale. Except for the early work of McKee and Resser (1945) there have been few sedimentologic studies of the Bright Angel Shale. These workers thought that the Bright Angel Shale represented fairly deep water sedimentation seaward of the Tapeats. Recent work by Wanless (1973) and Martin (1985) has provided new data that allow more precise environmental reconstructions.

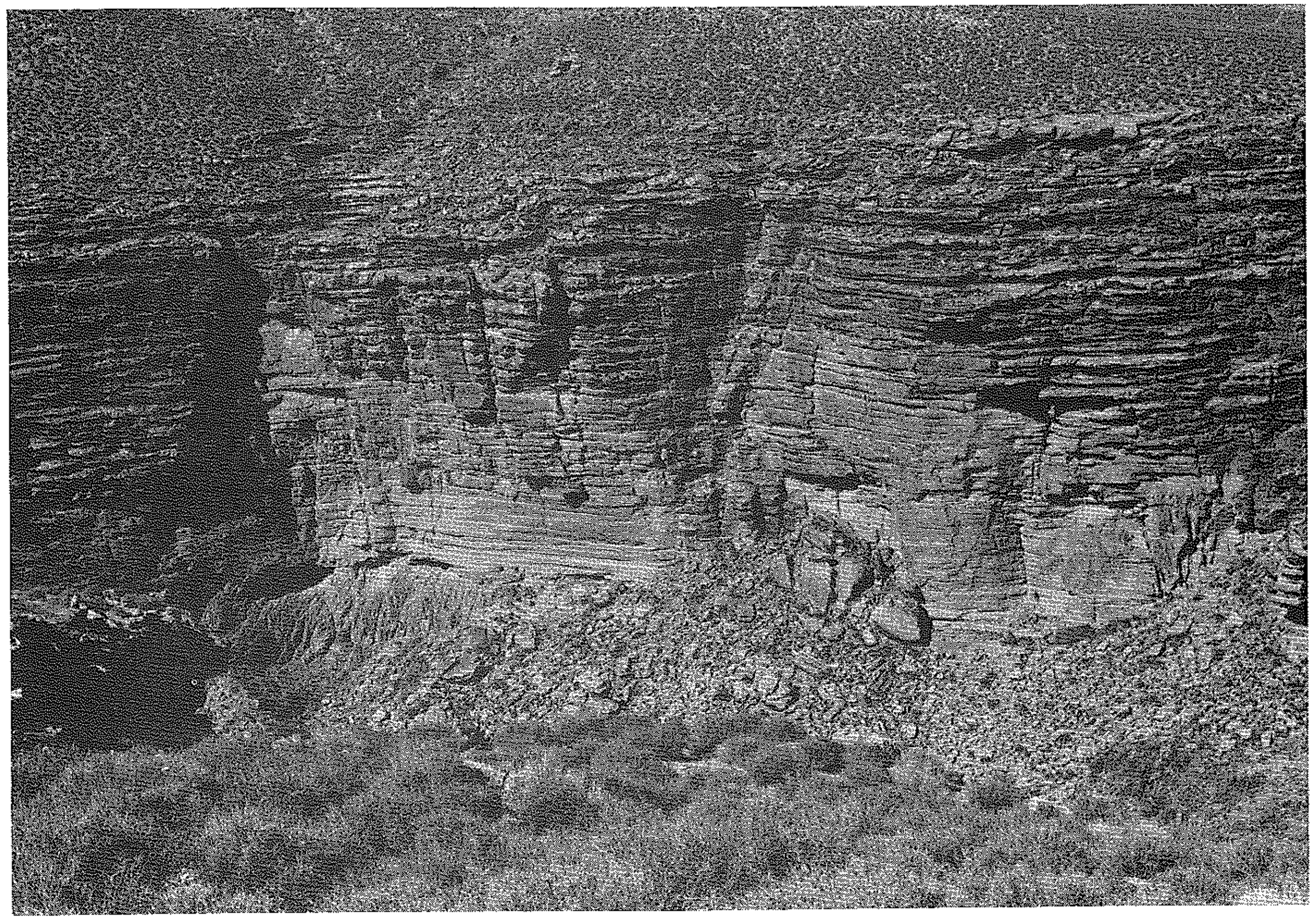


Figure 4. Typical outcrop of Tapeats near Horn Creek in the Grand Canyon, illustrating sheetlike geometry of the bedding. Tapeats unconformably overlies Precambrian crystalline rocks and is gradational with overlying Bright Angel Shale.

Although generally supporting the basic interpretation of McKee and Resser (1945), these studies have documented shallow-water deposits in the Bright Angel Shale as well as a complex history of strandline migration.

Martin (1985) recognized eight facies in the Bright Angel Shale and from these identified three genetically significant facies sequences (fig. 6). These include both upward-coarsening and upward-fining sequences and a sequence consisting of medium- to large-scale, planar-tabular, cross-stratified sandstone overlain by either trough cross-stratified or ripple cross-stratified sandstone.

Upward-coarsening sequences are up to 8 m thick and can typically be traced for several tens of kilometers. Martin (1985) interpreted these as the products of sand-sheet migration. The lower portions of these were deposited by fair-weather suspension settling of silt and clay and as well as by storm-enhanced currents (fig. 6a). The coarser grained upper parts were deposited in shallower water by migration of ripples and small sand waves on tops of the sand sheets. The tops of the sheets were subjected to periodic storm events, as indicated by hummocky cross-stratification.

Fining-upward sequences are common. Sedimentary and biogenic structures indicate that they were deposited by high-energy bed load transport followed by suspension settling. This sequence (fig. 6b) is similar to modern continental-shelf storm deposits (Swift and others, 1983). On modern shelves, however, these deposits form broad sheets unlike the lenticular geometry in the Bright Angel Shale. Lenticular fining-upward sequences are reported from the Upper Jurassic of Wyoming (Brenner and Davies, 1973), where they are interpreted to represent deposition due to rapidly waning flow following major storms. A similar origin is proposed for these deposits in the Bright Angel Shale.

The third sequence probably resulted from migration of sand waves (fig. 6c). The tops of the sand waves were apparently covered with small dunes and ripples. These sequences are most common in the lower half of the Bright Angel Shale and are typically encompassed by thick packages of interbedded siltstone and claystone. Although such features do occur on tidal flats, the absence of any subaerial features and the association with other subtidal facies support a subtidal interpretation (Boersma, 1969; Johnson, 1977).

In the central part of the canyon, large-scale, planar-tabular sets occur. Some of these are up to 4 m thick (fig. 7). The sandstone is medium to coarse grained and well sorted. The foresets dip up to 20 degrees, with foreset dip decreasing in the direction of flow. A marine sand-wave origin is indicated from the presence of trilobite crawling and resting traces on a few foresets (fig. 8) and from the occurrence of the associated bioturbated siltstone and claystone.

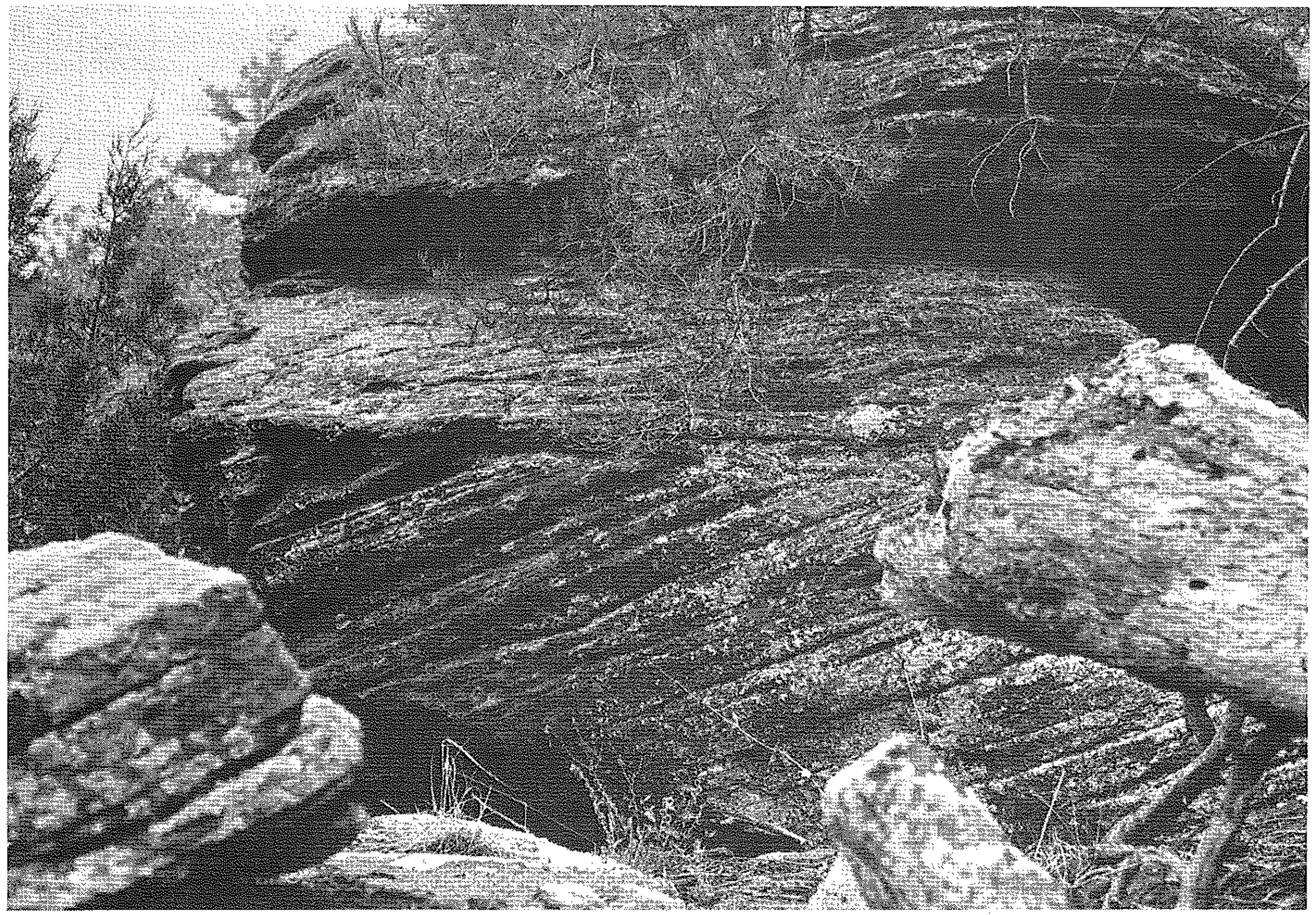


Figure 5. Large-scale (1.1-m-thick) planar-tabular cross-stratification overlain by horizontal to low-angle conglomeratic sandstone along East Verde River near Payson. These deposits formed on transverse bars in a braided river system.

Most of the Bright Angel Shale was deposited in a subtidal environment where both tidal and storm processes were active. However, intertidal deposits in the western Grand Canyon were reported by Wanless (1975, 1981).

Muav Limestone. McKee and Resser (1945) considered the Muav Limestone to be an outer-shelf, subtidal carbonate sheet deposited seaward of the Bright Angel detrital belt. Although many of the limestone and dolostone beds in the Muav are subtidal, as suggested by faunal and petrologic characteristics, Wanless (1973, 1975) reported intertidal and supratidal facies from the Muav. Evidence for very shallow water deposition includes algal-ball limestones, cryptalgal laminations, and nonburrowed laminated dolomitic mudstone and wackestone, which typically contain fecal pellets (Wanless, 1975). Thin-bedded laminated mudstones are common features on modern supratidal flats. Wanless (1975) suggested that the mud and pellets were deposited when storms flooded the tidal flats. Between storms, algae colonized the surface of the flat and this resulted in the formation of the cryptalgal structures. These beds are abundant near the top of the Muav in the western canyon.

Undifferentiated Dolomites. The "undifferentiated dolomites" (McKee and Resser, 1945) that overlie the Muav are not well understood in terms of their temporal and environmental significance. Brathovde (1986) has documented thick beds of oolitic grainstones and stromatolites interbedded with fine-grained carbonates. This association suggests shallow-water, possibly peritidal deposition. Brathovde (1986) has proposed the name "Grand Wash Dolomite" for these strata.

Southern Arizona

Bolsa Quartzite. Although the Bolsa Quartzite is well exposed in several areas, there have been no detailed sedimentologic analyses of the formation. The presence of various trace fossils, including *Skolithos*, and the sequences of cross-bedded facies (Hayes and Cone, 1975) suggest a depositional setting similar to that reported by Hereford

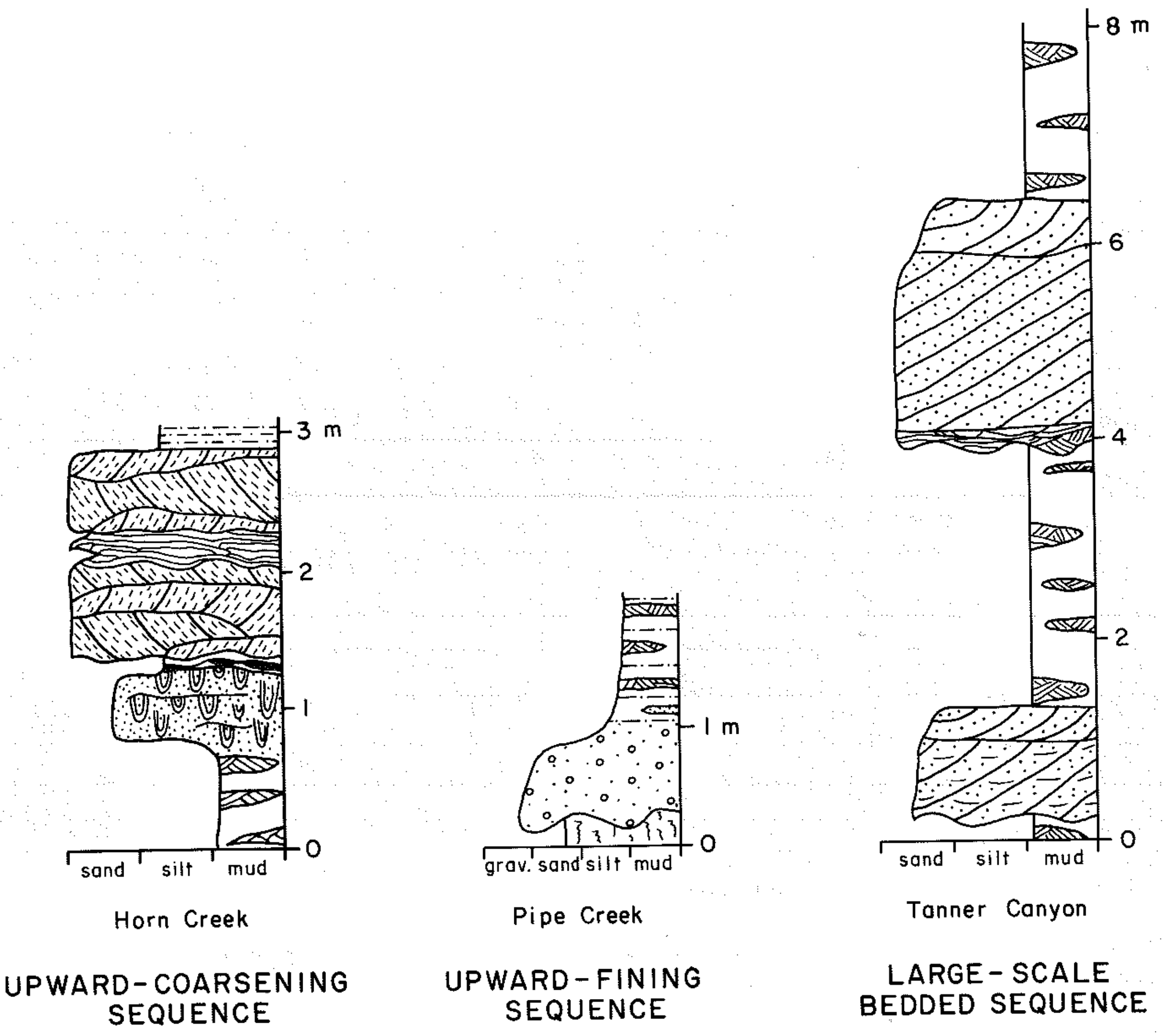


Figure 6. Three typical facies sequences in the Bright Angel Shale (after Martin, 1985).

(1977) for intertidal deposits in the Tapeats Sandstone. It is likley that the Bolsa Quartzite also includes shallow subtidal deposits.

Abrigo Formation. Work by Hayes and Cone (1975), McClure (1977), and Hayes (1978) provided the first environmental interpretation of the carbonates of the Abrigo Formation. Intraformational conglomerates, mudcracks, and small-scale cross-stratification in the lower member suggest an intertidal depositional environment. The middle member, which consists of carbonate in the south and sandstone in the north, records shallow subtidal and intertidal deposition. The upper sandy member reflects storm-influenced, inter- to supratidal depositional settings. The Copper Queen Member possibly represents shallow, subtidal deposition. Until detailed facies and petrologic studies are undertaken, regional environmental interpretations must be considered tenuous.

ORDOVICIAN STRATIGRAPHY

Ordovician rocks are present only in southeastern Arizona. They comprise two formations, which, in ascending order, are the El Paso Limestone and the Second Value Dolomite of the Montoya Group (Hayes and Cone, 1975). Although these formations were originally assigned to the Longfellow Limestone (Lindgren, 1905), Hayes (1972) argued that the lower part of the Longfellow is equivalent to the El Paso Limestone and that the upper 4.5 m near Morenci is correlative to the Second Value Dolomite in southwestern New Mexico.

El Paso Limestone. The term "El Paso Limestone" was first used in Arizona by Sabins (1957) for the thin-bedded limestone and dolostone overlying the Coronado Sandstone. The El Paso is exposed east of the Mule Mountains and northward to the erosional edge of Cambrian and Ordovician strata approximately 40 km north of Morenci (fig. 1). In the Pedregosa and Swisshelm Mountains, the formation conformably overlies the Copper Queen Member of the Abrigo Formation where it consists largely of fine- to coarse-grained packstone and wackestone with local intraformational conglomerate and chert nodules. Small-scale cross-stratification is present locally, and cryptalgal laminations and trace fossils are common. Epis and Gilbert (1957) reported over 130 m of the El Paso Limestone in the Swisshelm Mountains.

North of the Pedregosa and Swisshelm Mountains, the El Paso conformably overlies the Coronado Sandstone and consists of two units: a basal unit, up to 55 m thick, composed of sandy to silty dolomite and dolomitic quartz arenite and an upper unit composed of carbonate mudstone, grainstone, and dolomitic boundstone (Hayes and Cone, 1975). The lower unit is cross-stratified and in places contains intraformational conglomerate and trace fossils. The upper member has cryptalgal laminations, birdseye structures, and desiccation cracks (Hayes and Cone, 1975).

In the Swisshelm and Pedregosa Mountains, the Cambrian-Ordovician boundary is at the base of the El Paso Limestone. East and north of these ranges, the lower part of the El Paso contains a Late Cambrian (Franconian) fauna, whereas the upper part contains an Early Ordovician fauna (Sabins, 1957; Hayes and Cone, 1975).

Second Value Dolomite. Near Morenci, a coral-crinoidal limestone of Middle to Late Ordovician age disconformably overlies the El Paso Limestone (Hayes and Cone, 1975; Carroll, 1977). The unit is 4.6 m thick (Hayes, 1978) and has been correlated with the Second Value Dolomite (Montoya Group) to the east. As previously mentioned, both the El Paso Limestone and Second Value Dolomite originally had been included in the Longfellow Limestone (Lindgren, 1905).

ORDOVICIAN DEPOSITIONAL SYSTEMS

Detailed petrologic and sedimentologic studies of Ordovician strata remain to be done. Environmental reconstructions, therefore, are tenuous and quite generalized. Both the El Paso Limestone and the Second Value Dolomite were deposited in shallow marine waters, as suggested by their lithology, sedimentary and biogenic structures, and faunal assemblages.

The abundant fine- to coarse-grained packstones and interbedded wackestones and local intraformational conglomerates indicate alternating high-energy (probably stormrelated) and quiet-water conditions. North of the Swisshelm and Pedregosa Mountains, the siliciclastics in the lower part of the El Paso indicate deposition in fairly high energy environments. The upper member contains structures suggestive of tidal-flat deposition (Aitken, 1967; Shinn, 1968).

The upward change from subtidal to intertidal environments probably reflects either lowering of sea level or shoaling conditions with rapid carbonate accumulation. A major shift in strandline position, however, is suggested by the change from siliciclastics to carbonates.

CAMBRIAN-ORDOVICIAN PALEOGEOGRAPHY

Cambrian and Ordovician strata were deposited in shelf, nearshore, and continental settings (McKee and Resser, 1945; Stewart, 1970; Lochman-Balk, 1970, 1971; Hereford, 1977; Hayes, 1978; Middleton and Hereford, 1981). These deposits are generally regarded as a transgressive sequence composed of sandstone, shale, and limestone that was deposited on the slowly subsiding Cordilleran miogeocline and on the adjacent craton of the western margin of the North American plate (McKee and Resser, 1945; Lochman-Balk, 1971; Stewart, 1972; Stewart and Suczek, 1977). During the early Paleozoic, a north-south shoreline migrated progressively eastward across Arizona, and this resulted in deposition of coarse clastics on the craton to the east and fine clastics and carbonates in offshore environments to the west. Cambrian rocks thicken to the west and thus reflect greater subsidence in the miogeocline (fig. 9). Continued subsidence and (or) sea-level rise interrupted by a number of minor retreats of the sea (Lochman-Balk, 1971) resulted in a series of complex facies changes in both the siliciclastics and carbonates.

The geographic position of North America during the lower Paleozoic is just now beginning to be understood from paleomagnetic studies in the Rocky Mountains and the Appalachians. Paleomagnetic data from rocks of this age are questionable because of a lack of marine magnetic anomalies with which to compare reversal stratigraphy and chronology and the likelihood of diagenetic overprinting or complete removal of the natural remanent magnetization. However, a few studies have utilized enough samples and proper demagnetization techniques to determine the approximate lower Paleozoic paleolatitude of the North American craton (Van der Voo and others, 1976; French and others, 1977). These studies indicate that the Cambrian equator is presently oriented north-south and extends through the central portion of North America.

The implications of a low-latitude position are twofold. First, Arizona was within the trade-wind belt (Dott and Batten, 1981, p. 249), with open ocean to the north; thus, the coast was probably subjected to major storms. Studies by Dott (1974) of Cambrian deposits in the mid-continent and by Hayes (1967) of Holocene sediments off the Gulf Coast have documented the importance of such events on coastal erosion and sedimentation. Second, an equatorial climate should have influenced weathering and erosional and transportational processes operating on exposed cratonic crystalline rocks.

Although few data exist concerning the topography of the craton, it likely was characterized by rolling hills of Precambrian crystalline and metasedimentary rocks. A regolith developed on the crystalline basement and is

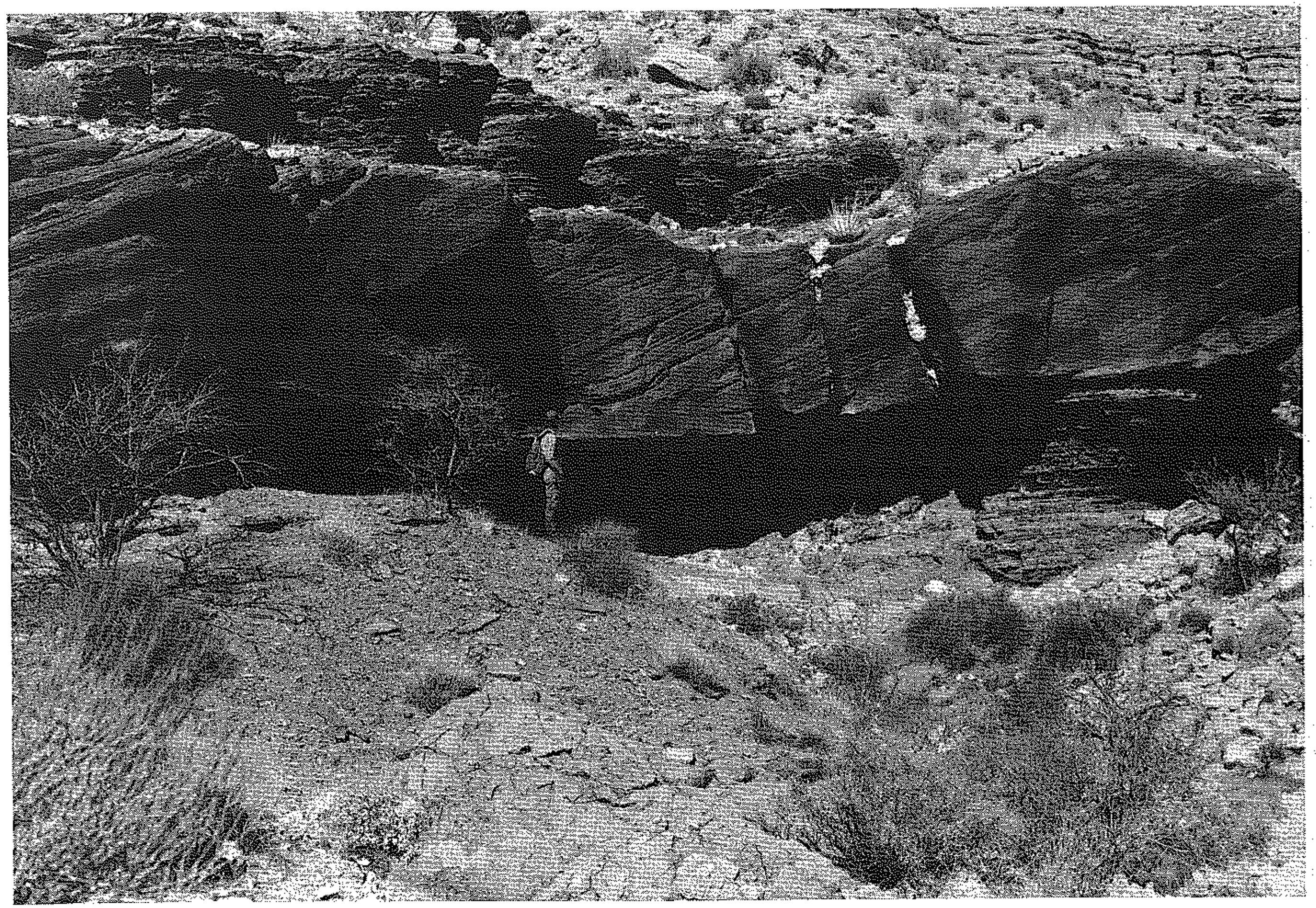


Figure 7. Large-scale cross-bedding in Bright Angel Shale at Horn Creek in the Grand Canyon. Over- and underlying beds are bioturbated siltstone and shale.

preserved locally beneath Tapeats Sandstone and Bolsa Quartzite (Sharp, 1940). Studies in Arizona (Sharp, 1940; McKee and Resser, 1945), Montana (Graham and Suttner, 1974), Utah (Hintze, 1973), and Wyoming (Middleton and others, 1980) demonstrated that considerable relief existed on the Precambrian surface before deposition of the basal sand. These positive features provided large amounts of detritus to Cambrian fluvial and nearshore systems.

Gross lithologic contrasts in the Cambrian strata of Arizona suggest an onlapping sequence of offshore lithosomes over nearshore ones. Thus, nearshore coarse clastics (Tapeats and Bolsa) are overlain by offshore muds and mixed fine siliciclastics and carbonates (Bright Angel) that are in turn overlain by more offshore carbonates (Muav and Abrigo). While fundamentally correct, this view has led to the assumption that the basal siliciclastics were exclusively marine and that vertical facies changes reflect deposition in progressively deepening waters. Fluvial deposits that surely were accumulating on the subaerially exposed craton east of the shoreline were presumably reworked during transgression.

Recognition of prevegetation (pre-Silurian) fluvial deposits is difficult because of a lack of modern analogues and well-established facies models (Miall, 1977; Cotter, 1978). Absence of terrestrial vegetation would have affected both erodibility and weathering and also degree of stream channelization (Schumm, 1968). Downslope movement of weathered material would have been rapid because slopes were not stabilized by vegetation. Schumm (1968) hypothesized that the landscape was similar to that in arid regions, even in relatively wet climates. The absence of land vegetation precluded significant bank stabilization (Schumm, 1968; Smith, 1976) thereby prohibiting development of meandering stream patterns. Braided streams with wide bedload channels should have been common, with vast sheetlike deposits of coarse-grained, poorly sorted alluvium spread over the land surface during flooding events (Schumm, 1968). A review of many pre-Silurian fluvial deposits has shown that all were apparently braided (Cotter, 1978).

Many characteristics of the basal Tapeats in Arizona fit Schumm's model for prevegetation fluvial deposition. The horizontally bedded sandstones and conglomerates probably represent sheets of crystalline detritus spread across broad alluvial plains following major flooding episodes (Middleton and Hereford, 1981). The Transcontinental Arch was the source of much of this detritus. Lack of deep cut-and-fill structures and well-defined channels in this Tapeats would be expected in a fluvial system where the stabilizing effects of vegetation were absent.

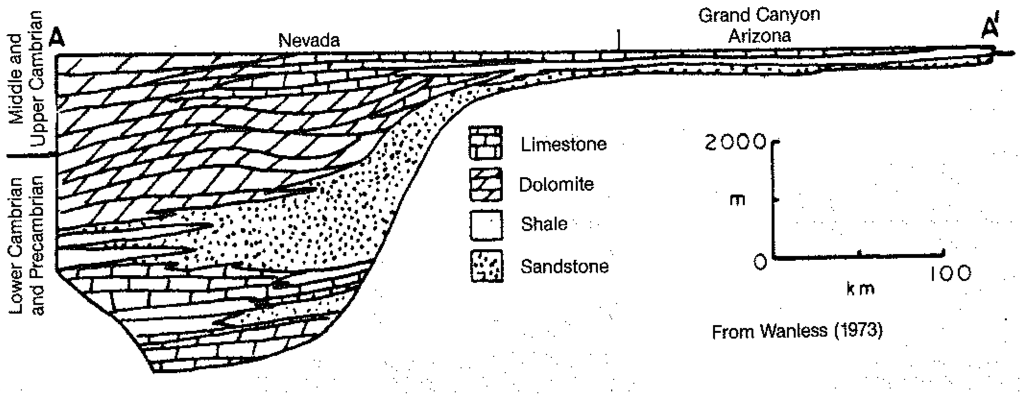
Figure 8. Trilobite crawling trace on foreset of large-scale cross-bedded set at Horn Creek.

Fluvial sediments in the Tapeats and Bolsa are overlain typically by mineralogically more mature sandstones that were deposited in a variety of nearshore environments. The mineralogic changes are best explained by preferential destruction of mechanically less stable grains during prolonged transport along the coast (Odom and others, 1976). Facies changes can be abrupt, with shallow subtidal deposits of thick-bedded quartz arenite resting directly on fluvial sandstones or on Precambrian crystalline rocks.

The rates of sea-level rise and (or) basin subsidence, together with the rate of sediment influx, are controlling factors in the stratigraphic relationships of marine sequences during transgression (Curray, 1964; Swift, 1968; and Ryer, 1977). Slow basin subsidence or sea-level rise and moderate to high rates of deposition produce onlapping of offshore over nearshore environments, resulting in a depositional transgression. Slight changes in rates of sea-level rise and sedimentation, however, can cause a transition from a depositional transgression to an offlapping regressive sequence. A discontinuous depositional transgression results in deposition of a thin veneer of littoral deposits overlain by more offshore sediments. This type of strandline movement results in a vertical stacking of facies that during deposition were not laterally contiguous.

If sea level remains stable and sediment supply is minimal, an erosional transgression can occur. Shoreface erosion of barrier beaches, lagoons, and marshes by waves results in offshore movement of sediment into sublittoral settings. Many modern coastal sequences are partly or completely reworked by shoreface erosion (Swift, 1968). The resulting vertical sequence consists of nearshore sands disconformably overlying an erosional surface or ravinement that develops with the steplike retreat of the shoreline (Swift, 1968). The thin erosional surface is recognized by its sharp contact, conglomeratic texture, and vertical superposition of lower shoreface sediments over fluvial lithosomes. This type of transition is common in the Tapeats at many localities.

The transgression that resulted in deposition of the basal portion of the Tapeats Sandstone was erosional, as evidenced by the overstepping of fluvial and Precambrian crystalline rocks by sublittoral deposits. The transition



From Wanless (1973)

Figure 9. Stratigraphic cross-section through Arizona illustrating westward thickening into miogeocline (modified from Wanless, 1973).

between these environments is typically abrupt and slightly conglomeratic, which suggests transgression associated with shoreface erosion (Swift, 1968). In many areas, however, there is no obvious erosional surface, and this suggests that sea level was no longer static but was rising. The rate of this rise is uncertain.

Further evidence of the oscillatory nature of strandline movement is the vertical superposition of intertidal over subtidal deposits in the Tapeats Sandstone and Bright Angel Shale. Transgression was followed by basinward migration of the shoreline, although it is unclear whether this reflects increase in sediment supply (shoreline progradation) or falling sea level. McKee and Resser (1945) recognized a number of transgressive-regressive cycles in the Bright Angel Shale that indicate successive movements of the strandline. Strata deposited in settings farther offshore also provide evidence of these periodic advances and retreats of the strandline, although in many areas the presence of peritidal deposits considerable distances offshore was controlled by processes intrinsic to sedimentation in offshore carbonate settings (Palmer and Halley, 1979).

The causes of the transgressions and regressions are unclear. The sediments accumulated along the trailing margin of the North American plate (Stewart and Poole, 1974). Stewart and Suczek (1977) postulated that the initial transgression in late Precambrian time and Early Cambrian time along the western margin of the North American craton was related to erosion of a thermally uplifted area as it migrated away from the spreading center. These authors suggested that as the newly formed crust moved away from the rift zone, it cooled and eventually subsided. Marine erosion coupled with subsidence resulted in breaching of the thermally uplifted western margin of the craton and inundation of the continental margin. Studies by Hays and Pitman (1973) and others have suggested a correlation between rates of sea-floor spreading and transgressions (and regressions) in the Cretaceous of North America. The numerous transgressive-regressive cycles in the lower Paleozoic of the southwestern United States might be best explained by this hypothesis, although it is difficult to verify.

Detailed sedimentologic analyses are needed to resolve not only paleogeographic problems but also to resolve the complex transgressive and regressive stratigraphies. Virtually the entire Cambrian and Ordovician sections need reanalysis.

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