The Pennsylvanian System in the Sacramento Mountains, New Mexico, USA

Stratigraphy, Petrography, Depositional Systems,
Paleontology, Biostratigraphy, and Geologic History

*Spencer G. Lucas, William A. DiMichele, Karl Krainer,
James E. Barrick, Daniel Vachard, Michael P. Donovan,
Cindy Looy, Hans Kerp, and Dan S. Chaney*

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Smithsonian Institution
Scholarly Press

WASHINGTON, D.C.
2021

ABSTRACT

Lucas, Spencer G., William A. DiMichele, Karl Krainer, James E. Barrick, Daniel Vachard, Michael P. Donovan, Cindy Looy, Hans Kerp, and Dan S. Chaney. The Pennsylvanian System in the Sacramento Mountains, New Mexico, USA: Stratigraphy, Petrography, Depositional Systems, Paleontology, Biostratigraphy, and Geologic History. *Smithsonian Contributions to Paleobiology*, number 104, x + 215 pages, 123 figures, 7 tables, 2021. — Pennsylvanian sedimentary rocks in the Sacramento Mountains, New Mexico, comprise an ~1 km thick stratigraphic section. The Morrowan-Desmoinesian Gobbler Formation was deposited by shallow marine processes in and near the Alamo clastic trough. In this trough, the Desmoinesian-Missourian Gray Mesa Formation (Bug Scuffle Member, Gobbler Formation) is a relatively thin unit (Space History Member) representing the glacioeustatic Amado event. The Missourian-Virgilian Beeman Formation includes the lower, siliciclastic Indian Wells Canyon Member and overlying, carbonate-rich Horse Ridge Member. The Virgilian Holder Formation consists of algal bioherms (Little Dry Canyon Member) overlain by the mixed carbonate-siliciclastic Mill Ridge Member. The Virgilian-Wolfcampian Bursum Formation is mixed siliciclastic-carbonate strata that represent shallow marine and nonmarine paleoenvironments. Animal and plant remains occur throughout the section. Unit age determinations are primarily based on conodont faunas recovered from the Gobbler, Gray Mesa, and Beeman Formations. Many conodont faunas correlate with Midcontinent cyclothems. Extensive algal and foraminiferal fossils also were identified in limestones from the section and contributed to age determinations. The Beeman Formation in particular contains an extensive Missourian macroflora. The macroflora is of “mixed” composition, containing typical wetland elements intimately intermixed with taxa indicative of seasonally dry habitats. A seasonally wet-dry background climate is indicated. It is unlikely that drought-tolerant plants were transported exclusively from “uplands.” Some plant remains have arthropod-feeding evidence. Previous analyses identified late Paleozoic ice-age glacioeustasy as the primary depositional driver of Pennsylvanian sedimentation in the Sacramento Mountains. We question this because of problems with those analyses and because of ample evidence of local tectonics and microclimate changes as important drivers of sedimentation in this area. Three Pennsylvanian Ancestral Rocky Mountain orogeny tectonic pulses can be identified in the Sacramento Mountains: Morrowan-Atokan, Missourian, and late Virgilian-Wolfcampian.

Cover image: Detail from Figure 10, view looking north at Horse Ridge near Alamogordo, New Mexico, which is a nearly complete stratigraphic section of the Pennsylvanian System in the Sacramento Mountains.

Published by SMITHSONIAN INSTITUTION SCHOLARLY PRESS
P.O. Box 37012, MRC 957, Washington, D.C. 20013-7012
<https://scholarlypress.si.edu>

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Library of Congress Control Number: 2020944263

ISSN: 1943-6688 (online); 0081-0266 (print)

Publication date (online): 22 February 2021

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Stratigraphy, Petrography, Depositional Systems, Paleontology, Biostratigraphy, and Geologic History

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Manuscript received 20 December 2019; accepted 28 May 2020.

INTRODUCTION

The Sacramento Mountains are a late Cenozoic fault-block uplift of the Basin and Range physiographic province in southern New Mexico (Figure 1). Extending about 100 km north–south and 20–30 km east–west, the Sacramento Mountains expose a stratigraphic section of Paleozoic sedimentary rocks about 2,400 m thick (Pray, 1961). A substantial portion of this section is about 900 m of Pennsylvanian strata. These strata have a broad outcrop belt along the western escarpment of the Sacramento Mountains and in the faulted northern part of the range near Alamogordo (Figure 2).

Pray (1961) represents the most recent synthesis of the Pennsylvanian System in the Sacramento Mountains, though many subsequent publications have focused on portions of the Pennsylvanian section. Here, we present new and extensive data on the Pennsylvanian strata in the Sacramento Mountains from that portion of the range immediately east and northeast of the city of Alamogordo (Figures 1, 2). This is an area where the Pennsylvanian strata were deposited on a narrow shelf (Sacramento shelf) between the late Paleozoic Pedernal uplift to the east and the late Paleozoic Orogrande Basin to the west (Figure 3). Our goals here are to present new data on and to review and revise the Pennsylvanian lithostratigraphy in this area, to present new petrographic and paleontological data, and to interpret the biostratigraphy, general nature of the depositional systems, and sedimentary environments and geological history of the Pennsylvanian strata.

MATERIALS AND METHODS

The fieldwork this study is based on took place between November 2013 and April 2017. Fourteen stratigraphic sections were measured (Figure 2, Table 1). While

FIGURE 1. Location map of the study area in the Sacramento Mountains of southern New Mexico, USA.

measuring these sections (see Figures 2, 8, 9, 11, 12, 13, 20, 24, 25, 26, 27, 33, 34, 40, 41, 43; note, figures appear in later sections where they are discussed more), samples were collected for micropaleontological analysis, both of calcareous microfossils (see Figures 103–110) and conodonts (see Figures 111–118). Fossil plants (see Figures 60–102) from localities located in green or black shale and fine-grained siltstone beds of the Gobbler, Beeman, and Holder Formations were quarried by splitting the sediments with hand tools. Some macroscopic animal fossils were recovered from the plant fossil localities (see Figures 119–120). Numerous petrographic and carbonate thin sections were

prepared from collected rock samples (see Figures 17, 21, 22, 23, 29, 30, 31, 36, 37, 38, 45, 46).

Macroscopic photography was by digital camera, and microscopic photography was by a standard transmitting light microscope or via scanning electron microscopy. Line figures were created with standard graphics software. Macrofossils and microfossils reported here are in the collections of the New Mexico Museum of Natural History, Albuquerque, New Mexico (NMMNH), or the National Museum of Natural History, Smithsonian Institution, Washington, D.C. (USNM).

FIGURE 2. (*Opposite*) Generalized geological map of the study area showing the location of the measured stratigraphic sections and important geographic features mentioned in the text. The base map is based on Pray (1959b). Measured stratigraphic sections: 1 = La Luz Canyon B; 2 = La Luz Canyon A; 3 = La Luz Canyon C; 4 = High Rolls; 5 = Beeman Canyon B; 6 = Beeman Canyon A; 7 = Holder bioherm; 8 = Space History D; 9 = Space History C; 10 = Space History B; 11 = Space History A; 12 = Indian Wells Canyon; 13 = Long Ridge; 14 = Mule Canyon.

FIGURE 3. Late Paleozoic paleotectonic map of southern New Mexico.

TABLE 1. Locations of measured stratigraphic sections in Universal Transverse Mercator (UTM) and section (Sec), township (T), and range (R) coordinates.

Measured section	Base	Top	Sec, T, R
Beeman Canyon A	413656, 3645501	413133, 3645524	4, 16S, 10E
Beeman Canyon B	413116, 3645580	413133, 3645524	4, 16S, 10E
High Rolls	419777, 3654846	419818, 3646069	6, 16S, 11E
Holder bioherm	414201, 3645892	414642, 3645764	3, 16S, 10E
Indian Wells Canyon	416669, 3642754	416634, 3642424	1, 16S, 10E
La Luz Canyon A	414125, 3649972	414502, 3649814	30, 15S, 11E
La Luz Canyon B	414047, 3649684	416732, 3632825	30, 15S, 11E
La Luz Canyon C	413823, 3649758	414125, 3649972	30, 15S, 11E
Long Ridge	415880, 3632875	416732, 3632825	Not surveyed
Mule Canyon	415225, 3632905	415726, 3633090	Not surveyed
Space History A	415968, 3642553	416016, 3642750	11, 16S, 10E
Space History B	414080, 3642691	414625, 3643983	10, 16S, 10E
Space History C	414126, 3643725	414097, 3644404	10, 16S, 10E
Space History D	414423, 3644513	414567, 3644309	3, 16S, 10E

PREVIOUS STUDIES

Darton (1917, 1922, 1928) provided the earliest published observations on the Pennsylvanian strata exposed in the Sacramento Mountains. He assigned these strata to the Magdalena group (Figure 4) and included them in several structural cross sections (Darton, 1922: figs. 23, 24; 1928: figs. 80–84; Figure 5). Darton (1928:202–205) provided more detailed information than Darton (1922) and indicated an overall thickness of the

Pennsylvanian strata in the Sacramento Mountains of at least 459 m. He based this thickness on a stratigraphic section of the Pennsylvanian strata measured by George H. Girty on a ridge “N63° E. of Alamogordo,” which is very likely the equivalent of our Space History A–D sections (see sections on the Gobbler through Bursum formations below).
Geologic mapping in 1913–1916 (Darton, 1928: pl. 44) identified a broad outcrop belt of the Magdalena group (above Mississippian Lake Valley limestone and below Permian Abo sandstone)

Darton (1917, 1922, 1928)	Thompson (1942), Lloyd (1949)	Pray (1961)	this paper	
Magdalena group	Fresnal Group	Bursum Formation (part)	Bursum Formation (part)	
	Keller Group	Holder Formation	Holder Formation	Mill Ridge Member
				Little Dry Canyon Member
	Missouri series	Beeman Formation	Beeman Formation	Horse Ridge Member
	Des Moines series	Bug Scuffle Limestone Member	Space History Member	Indian Wells Canyon Member
			Gray Mesa Formation	
Atoka or Morrow series		Gobbler Formation	Gobbler Formation	

FIGURE 4. Evolution of Pennsylvanian lithostratigraphic nomenclature in the Sacramento Mountains.

FIGURE 5. Darton's (1928: fig. 80) cross section of the western front of the Sacramento Mountains. Note that the view is to the south, so west is to the right side of the diagram. Abbreviations are as follows: b = Bliss? Formation (Cambro-Ordovician); e = El Paso Group (Ordovician); f = Fusselman Formation (Silurian); gr = granite (Proterozoic); m = Montoya Formation (Ordovician); mf = shaly limestone with Mississippian fossils; w = limestone, white when weathered (Fusselman?).

along the western escarpment of the Sacramento Mountains from La Luz on the north to Sand Canyon on the south. Darton (1928) also published lists of invertebrate fossil taxa (mostly brachiopods, identifications by G. Girty) collected from various Pennsylvanian localities in the Sacramento Mountains.

Laudon and Bowsher (1941), in a study of the Mississippian strata in the Sacramento Mountains, referred to the overlying Pennsylvanian strata as the Magdalena Formation. In a later publication, Laudon and Bowsher (1949:22–31) referred to the base of the Pennsylvanian section in the Sacramento Mountains as “Derry.”

Thompson (1942:73) introduced the term Fresnal Group for strata exposed along the highway in Fresnal Canyon

(NW 1/4 section 30, T15S, R11E) northeast of Alamogordo (Figure 2). The type section of Thompson's Fresnal Group (Figure 6) was shown by him to be 162 m thick at the type locality but said by him to be much thicker elsewhere. Thompson's type Fresnal Group strata are a mixture of clayey limestone, limestone, arkosic sandstone, conglomerate, and gray to red shale, and some of these strata yield Virgilian fusulinids. These strata are now assigned to the Holder Formation and lower part of the Bursum Formation (Figure 4). Thompson (1942) also coined the name “Pedernal land mass” for a late Paleozoic uplift in southern and central New Mexico that supplied sediment to the Pennsylvanian–lowermost Permian strata in the Sacramento Mountains.

FIGURE 6. Thompson's (1942) type section of the Fresnal Group.

Lloyd (1949: pl. 4, section K) presented a complete Pennsylvanian section in the northern Sacramento Mountains measured by M. L. Thompson (Figure 4). No detailed location of this section was given, but the upper part of the section is the type section of the Fresnal Group of Thompson (1942). Lloyd (1949) reported the following thicknesses: Atokan = 133 m (though Lloyd noted the lower half might be of Morrowan age),

Desmoinesian = at least 230 m, Missourian = 274 m, Virgilian Keller Group = 107 m, and Virgilian Fresnal Group = 162 m.

Plumley and Graves (1953) presented the first description of Virgilian "reefs" in the Sacramento Mountains. These are the algal bioherms of the Holder Formation studied by numerous workers since the original 1953 report (see below). Pray and Graves (1954) briefly discussed Desmoinesian facies in

the Sacramento Mountains. They distinguished a so-called deltaic facies in the western part of the Sacramentos that grades “abruptly” (in their terms) to the northeast and south into what they described as a “shelf limestone” facies.

During the 1950s, a variety of student master’s and doctoral theses were completed on different aspects of the Pennsylvanian strata in the Sacramento Mountains (e. g., Pray, 1952; Lokke, 1953; Otté, 1954; Bergeron, 1957; Oppel, 1957; Reyer, 1958). Some of these theses, particularly those of Pray and Otté, were published to provide the first detailed understanding of the Pennsylvanian strata in the Sacramento Mountains.

Otté (1959a) presented a detailed study of the strata across the Pennsylvanian–Permian (Virgilian–Wolfcampian) boundary in the northern Sacramento Mountains. This work focused on the unit Otté (1959a) named the Laborcita Formation (Bursum Formation of Pray, 1961, and of this work).

In a guidebook to the geology of the Sacramento Mountains, jointly sponsored by the Permian Basin section of the Society for Sedimentary Geology and the Roswell Geological Society (Pray, 1959a), several articles were devoted to aspects of the Pennsylvanian section. Thus, Pray (1959b) presented an overview of the Pennsylvanian section that first named its constituent units the Gobbler, Beeman, and Holder Formations (also see Pray, 1961). Cline (1959) identified cyclic sedimentation in the upper Virgilian strata exposed near La Luz. Oppel (1959) described in detail the stratigraphy across the Pennsylvanian–Permian boundary in Fresno Canyon. Otté (1959b) presented a brief summary of his more extensive publication (Otté, 1959a) on the Bursum (=Laborcita) Formation in the northern Sacramento Mountains. Similarly, Bachman and Hayes (1959) briefly reprised their earlier article (Bachman and Hayes, 1958) on the Pennsylvanian–Permian section in the southern Sacramento Mountains.

Kottlowski (1960:92–96) briefly reviewed the Pennsylvanian stratigraphy in the Sacramento Mountains based largely on Thompson (1942), Lloyd (1949), Pray (1952, 1954), and Otté (1959b). In a stratigraphic cross section (Kottlowski, 1960: pl. 12), he presented the Pennsylvanian section in the Sacramento Mountains, much as did Lloyd (1949), as a succession of series and stages, though in the text Kottlowski used the lithostratigraphic nomenclature of Pray and Otté.

Pray (1961) named the Pennsylvanian formations in the Sacramento Mountains (ascending) the Gobbler, Beeman, and Holder Formations (Figure 4). He first named them in his unpublished doctoral dissertation (Pray, 1952) and later published the names in Pray (1959b), but because this article was published in a guidebook, he did not regard that as the official publication of his new lithostratigraphic names. Pray (1961) designated the type sections of his new units in one stratigraphic section on Long Ridge (Figures 2, 7). He coined the name Gobbler Formation for the lower 366 to 488 m of the Pennsylvanian section, a unit of diverse lithotypes, mostly clastic—quartz sandstone, chert-cobble conglomerate, shale, and siltstone. He included a thick, cherty limestone-dominated interval up to 305 m thick in the middle upper Gobbler Formation stratigraphic interval as the Bug Scuffle Limestone Member. He noted

that in the western Sacramento Mountains, this member is equivalent to a more clastic interval of the upper Gobbler Formation that he termed the “detrital facies” (Figure 4). Pray’s (1961) Beeman Formation is 107–152 m thick and consists mostly of thin-bedded, clayey limestone and shale. The Holder Formation is up to 274 m thick and crops out as ledges of limestone with shale interbeds and some prominent algal bioherms. Pray’s (1961) Pennsylvanian lithostratigraphic nomenclature has been used in the Sacramento Mountains for more than 50 years without modification (Figure 4).

Meyer (1966, especially pl. 1) reviewed the Pennsylvanian section in the Sacramento Mountains and, as had Lloyd (1949), assigned series and stage names to the strata. His isopach and isolith maps document the abrupt thinning of the Pennsylvanian strata of the Sacramento Mountains eastward onto the thus well-established position of the late Paleozoic Pedernal uplift. His data (see especially Meyer, 1966: figs. 31–33) also identified what he called the “Alamo clastic shelf” northeast of Alamogordo during Desmoinesian time, which he attributed to deltaic deposition, as had Thompson (1942) and Pray (1961). Thus, Meyer (1966:46) stated that “the detrital sediments [of the Gobbler Formation] were carried into the area by a major distributary whose headwaters lay in the Pedernal uplift to the east.”

Wilson (1967; Wilson, 1972a, is a reprint of this article) identified cycles of deposition in shelf deposits of the Holder Formation and compared and correlated them to basinal cycles of the Panther Seep Formation in the San Andres Mountains to the west. He attributed these cycles to sea-level fluctuations. He identified at least 20 such cycles during Holder deposition and estimated periodic drops of sea level of at least 30–46 m. Wilson (1972b:100) argued that late Paleozoic tectonic movement associated with the La Luz anticline in the northern Sacramento Mountains “strongly controlled facies and thickness in Late Pennsylvanian sediments.” He thus noted rapid facies changes from west to east in a small (12.95 km²) area from Indian Wells to La Luz Canyon, much of the same area we studied in detail (Figure 2). Steiner and Williams (1968) reported early Wolfcampian fusulinids from an interval in the Bursum (=Laborcita) Formation that begins about 27 m above the formation base in the vicinity of Laborcita Canyon.

In a West Texas Geological Society field trip guidebook to the geology of the Sacramento Mountains, several articles focused on the Pennsylvanian section, particularly the algal bioherms in the Holder Formation. Thus, Wilson (1977) discussed the overall distribution of such bioherms in the Upper Pennsylvanian–Wolfcampian of New Mexico. Toomey et al. (1977) reviewed the growth of what they referred to as the Yucca mound complex in Dry Canyon (Figure 2). Parks (1962) had earlier reported on the mound-building organisms, and more than one student thesis and diverse articles have been published about these mounds (e.g., Wilson, 1967, 1975; Cys and Mazullo, 1977; Parks, 1977a, 1977b; Toomey et al., 1977; Mazullo and Cys, 1979; Bowsher, 1986; Mazullo, 1988; Toomey, 1991; George, 1992; Kirkland et al., 1993; Rivera, 1999; Cho, 2004; Wasson, 2005; Goldstein et al., 2007; Wasson and Lohmann, 2015).

FIGURE 7. Type sections of the Gobbler, Beeman, and Holder Formations at Long Ridge (modified from Pray, 1961).

Van Wagoner (1977a, 1977b) analyzed deposition of the Gobbler Formation, identifying carbonate and siliciclastic-carbonate cycles. Cys and Mazullo (1977) and Delgado (1977) discussed some of the depositional facies of the Bursum Formation, as did Fly (1985, 1986).

Wilson (1989) constructed a northeast–southwest cross section of Morrowan–Atokan strata anchored in the northeast in the Sacramento Mountains. In his section in the Sacramento Mountains, Wilson identified Derryan (i.e., Atokan) and Desmoinesian strata. According to Wilson, these strata were deposited during a time of relative tectonic stability. He noted that “Middle Pennsylvanian . . . strata are widely represented by limestones with upward-shoaling cyclothems (generally upward-coarsening sequences) that number about a dozen at each location . . . [and] are not correlative from section to section” (p. 13). Wilson described these as thin cyclothems, about 15–40 m thick.

Algeo et al. (1991) identified cyclic sedimentation in the Gobbler Formation as 20–25 “carbonate shoaling cycles” and claimed basinwide correlateability of the cycles. They posited cycle periodicities of 100–200 kyr (=glacioeustasy) and identified sets of five to seven cycles that alternate between shale rich and carbonate rich. They viewed these as ~1.0 Ma cycles that reflect longer-term eustatic or tectonic cycles.

Raatz et al. (1994) studied Missourian deposition in the northern Sacramento Mountains. They identified three types of “lithologic associations” (LAs), which were interpreted as transgressive-regressive cycles in the upper Gobbler and Beeman Formations: (1) sandstone-dominated Horse Ridge LA, (2) open-marine carbonate-dominated Indian Wells LA, and (3) restricted carbonate-dominated Dry Canyon LA.

Raatz (1996) undertook a dissertation studying deposition of the Beeman Formation in the area of our study. He identified a hierarchy of parasequences, sequences, and sequence sets in the Beeman and attributed their deposition to glacioeustatic cycles. Raatz and Simo (1998) published a stratigraphic and sedimentological study of the Beeman Formation in the Dry Canyon area northeast of Alamogordo based on Raatz (1996). They presented a high-resolution sequence stratigraphy that identified high-frequency cycles in the Beeman Formation posited to have been forced by glacioeustasy. Raatz and Simo (1998) stated that tectonics and climate controlled longer-term trends in sedimentation and basin geometry.

Krainer et al. (2003) and Krainer and Lucas (2013) presented some stratigraphic, petrographic, and microfossil data on the Bursum Formation in the Sacramento Mountains. Kues (2004a) documented invertebrate fossils from two localities in the Holder Formation in Dry Canyon. Koning et al. (2007) published a preliminary geologic map of the Alamogordo North 7.5-minute quadrangle, which encompasses most of our study area. Hunt et al. (2012) and Hunt and Lucas (2017) described coprolites from the Beeman Formation in Dry Canyon.

Very few of the results of the research reported here have been previously published. Lucas et al. (2014b) documented body fossils of the horseshoe crab (limulid) *Euproops danae* Meek and Worthen from the Beeman Formation. Looy et al.

(2017) briefly discussed a plant fossil with a significant combination of cordaitalean and conifer morphologies from the Beeman Formation. Frederick et al. (2018) documented some conodonts from the Bursum Formation near Tularosa. Wahlman and Barrick (2018) reported some elements of the conodont biostratigraphic data presented in much more detail here.

GOBBLER FORMATION

LITHOSTRATIGRAPHY

The stratigraphically lowest (and oldest) Pennsylvanian unit in the Sacramento Mountains is the Gobbler Formation (Figure 4, Table 2). Pray (1961:75) named the Gobbler Formation for Gobbler dome and triangulation station in section 19, T18S, R11E (south of the area of Figure 2). Its type section is the lower part of his Long Ridge section (Figure 7; Pray, 1961: fig. 20), which is our Mule Canyon section (Figures 8, 9). We also measured three other stratigraphic sections of most or all of the Gobbler Formation at Space History A and B and at Indian Wells Canyon (Figures 10–13, Table 1).

Pray (1961) stated that the Gobbler Formation is 366–488 m thick. He divided the formation into three intervals: (1) a lower, up to 180 m thick interval of shale, sandstone, and minor limestone; (2) the Bug Scuffle Member, up to 305 m thick, consisting mostly of limestone, much of it cherty; and (3) laterally equivalent to and locally overlying the Bug Scuffle Member, the “detrital facies” of shale, sandstone, and minor limestone (Figure 4). The base of the Gobbler Formation is an unconformity on underlying Mississippian strata, with local channels at the base of the Gobbler having as much as 30 m of stratigraphic relief (Pray, 1961). Pray (1961) also noted that the most characteristic lithologies of the Gobbler Formation are quartzose sandstone with angular grains that are coarse to granular in size and limestone with chert.

The thickness of the Gobbler Formation in our measured sections ranges from 238 m at the Indian Wells Canyon section (Figure 13) to 180 m at the Mule Canyon section (Figure 9) and 218 m at the Space History B section (Figure 12A,B), where the base is not exposed. At the Space History A section (Figure 11), only the lower 114 m are exposed. At the Mule Canyon section, the Gobbler Formation disconformably overlies crinoidal limestone of the Mississippian Rancheria Formation, and at the Space History A and Indian Wells sections it overlies crinoidal limestone of the Mississippian Lake Valley Formation. At Mule Canyon, Space History B, and Indian Wells Canyon, the Gobbler Formation is overlain by the Gray Mesa Formation (=Bug Scuffle Limestone Member of earlier usage; see below). At Space History A, the upper part of the Gobbler Formation is not exposed.

The Gobbler Formation in general (Table 2, Figure 14) is composed of thick covered/shale intervals, coarse siliciclastic intervals (conglomerate, pebbly sandstone, sandstone), and limestone intervals. A considerable lateral facies change is observed from Indian Wells Canyon to Space History A and B. At Indian

TABLE 2. Summary of Pennsylvanian stratigraphic units in the Sacramento Mountains.

Formation	Maximum thickness (m)	Lower contact	Subdivisions	Principal lithologies	Age
Bursum	268	Disconformable	Undivided Laborcita Member in Sacramento Mountains	Interbedded red-bed siliciclastics and limestones; local algal mounds	Late Virgilian–early Wolfcampian
Holder	274	Conformable	Little Dry Canyon Member (below), Mill Ridge Member (above)	Algal mounds (Little Dry Canyon Member) overlain by interbedded carbonates and siliciclastic strata (shale, sandstone, and conglomerate; Mill Ridge Member); generally cyclical bedding	Early–late Virgilian
Beeman	247	Conformable	Indian Wells Canyon Member (below), Horse Ridge Member (above)	Shale, siltstone, sandstone, and conglomerate with a few limestone beds (Indian Wells Canyon Member), overlain by interbedded limestone and shale with a few beds of sandstone and conglomerate (Horse Ridge Member)	Early Missourian–early Virgilian
Gray Mesa	165	Conformable	Undivided, except for Space History Member (thin, upper member or tongue in Horse Ridge area)	Limestone, thick or thin bedded, many beds cherty, with some shale interbeds	Desmoinesian–early Missourian
Gobbler	342	Disconformable	Upper part laterally equivalent to Gray Mesa Formation	Shale, quartz-rich sandstones, pebbly sandstones and conglomerates, some interbedded limestone beds	Morrowan–Desmoinesian

Wells Canyon, coarse siliciclastic intercalations are thickest (up to 32.6 m) and coarsest; at Mule Canyon coarse siliciclastic intervals are still up to 14 m thick, whereas at Space History B sandstone intervals are less than 5 m thick and less coarse grained. Intercalated limestone units occur either as single limestone beds (mostly <0.5 m thick, muddy limestone) or, less abundantly, as thicker limestone units commonly containing abundant crinoidal debris that are rarely thicker than 2 m. Cherty limestone is common at Space History A but very rare to absent at the other sections. At Space History B, dolomitic limestone beds are intercalated. A distinct facies is crinoidal sandstone and rudstone that crops out as units up to 8 m thick at Space History B.

Mule Canyon Section

Our Mule Canyon section (Figure 9, Table 1) is Pray's type section of the Gobbler Formation (Figure 7). At Mule Canyon, the Gobbler Formation rests on crinoidal limestone of the Mississippian Rancheria Formation and is approximately 180 m thick. The succession is composed of thick, covered (likely shale) intervals; shale intervals; sandstone intervals up to approximately 14 m thick; and intercalated limestone units, most less than 0.5 m thick; maximum limestone thickness is 1.5 m. The shale is mostly gray and subordinately black. Sandstone is present as thin sandstone units (<2 m) and three thicker sandstone

units (3.5–14 m thick). It is mostly quartzose, coarse grained, subordinately pebbly, and trough cross-bedded. Very rare, fine-grained sandstone beds with ripple lamination are present. Two thin Cenozoic sills are intercalated in the section.

Two lithotypes of limestone are distinguished: (1) thin limestone beds (bioclastic mudstone and wackestone) that are less than 0.5 m thick, intercalated in shale, and rarely thin bedded and (2) thicker limestone units (maximum 1.5 m thick), some of which contain chert nodules and chert bands. Most of these limestone units are evenly bedded, and subordinately, wavy bedding is present. The limestone mostly displays muddy textures (bioclastic mudstone and wackestone) but includes crinoidal limestone (crinoidal wackestone to packstone).

Space History A Section

The Gobbler Formation is exposed at the Space History A and B sections (Figures 10–12). Section Space History A includes the basal 114 m of the Gobbler Formation, and at Space History B, 218 m of Gobbler Formation are exposed. These sections do not appear to overlap (they are not readily correlated to each other lithostratigraphically), and the conodont data presented below also indicate little or no temporal overlap of the sections. This indicates a minimum Gobbler thickness of 332 m in the area of the New Mexico Museum of Space History.

FIGURE 8. Legend for lithologic symbols in graphical drawings of measured stratigraphic sections (Figures 9, 11–13, 20, 24–27, 32–34, 40–41, 43).

FIGURE 9. Mule Canyon section, the type section of the Gobbler Formation of Pray (1961). See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols.

FIGURE 10. View looking north at Horse Ridge, encompassing the strata in the Space History B, C, and D measured stratigraphic sections.

At the Space History A section (Figure 11), the Gobbler Formation overlies bedded Mississippian limestone of the Lake Valley Formation composed of crinoidal rudstone that contains crinoid stem fragments up to several centimeters long. The lower Gobbler Formation is a succession of thick covered intervals (probably shale) and limestone and sandstone intervals. The sandstone intervals are 0.6–2.8 m thick and mostly composed of quartzose, coarse-grained, pebbly sandstone with clasts up to 1 cm in diameter. In the lower part of the section, thin, cross-bedded sandstone beds are intercalated, and in the upper part thicker units with multistorey cross beds are exposed. One fine-grained, massive sandstone bed also is present. A mixed siliciclastic-carbonate sandstone interval (unit 6, 1.1 m thick) composed of abundant quartz grains and crinoid fragments is present.

Limestone units of the Gobbler Formation at Space History A are 0.3–2 m thick and thin to medium bedded, with even bedding. Most of the limestone units contain chert. Some units are dolomitic. Two limestone types can be distinguished:

1. Muddy limestone intervals (bioclastic mudstone, wackestone, and floatstone) contain crinoid fragments, locally abundant brachiopods, and some corals. This lithofacies is partly bioturbated.

2. Crinoidal limestone (packstone to rudstone) contains abundant crinoid fragments that are up to 1 cm in diameter and up to several centimeters long. Locally, corals, brachiopods, and echinoid radioles are present. Fossils are partly silicified. Rarely, cross bedding is observed.

Space History B Section

At the Space History B section (Figure 12A,B), 218 m of Gobbler Formation are overlain by the Gray Mesa Formation. The Gobbler Formation strata in this section can be divided into five subunits (A–E, in ascending order; Figure 12A,B). Subunit A is 25.7 m thick and composed of siltstone and mudstone with intercalated sandstone and rudstone. Two cross-bedded sandstone beds are intercalated in the lower part. One of these beds (unit 3) is mixed siliciclastic-carbonate sandstone containing abundant crinoid fragments and a few brachiopod fragments. In the middle part of subunit A, a few thin sandstone beds (5–10 cm thick) are intercalated. The upper 8.2 m are composed of trough-cross-bedded, partly crinoidal sandstone and rudstone, forming a prominent ledge. The rudstone also contains fragments of brachiopods and corals.

Subunit B (Figure 12A) is approximately 48 m thick and composed mainly of dark gray to black shale. Many dolomitic

FIGURE 11. Space History A section. See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "SHA 12").

Space History B

FIGURE 12B. Space History B section of the Pennsylvanian section north of the Space History Museum. Gobbler Formation, Gray Mesa Formation, and Beeman Formation strata are portrayed here. See Table 1 and Figure 2 for location of section and Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "SH 54").

FIGURE 12C. Space History B section of the Pennsylvanian section north of the Space History Museum. Beeman Formation and Holder Formation strata are portrayed here. See Table 1 and Figure 2 for location of section and Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, “SH 54”).

Indian Wells Canyon

FIGURE 13A. Indian Wells Canyon section of part of the Pennsylvanian section exposed in Indian Wells Canyon. Lake Valley Formation and Gobbler Formation strata are portrayed here. See Table 1 and Figure 2 for location of section and Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, “IWC 2”).

FIGURE 13B. Indian Wells Canyon section of part of the Pennsylvanian section exposed in Indian Wells Canyon. Gobbler Formation and Beeman Formation strata are portrayed here. See Table 1 and Figure 2 for location of section and Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "IWC 2").

FIGURE 14. Overview of middle upper part of Gobbler Formation outcrops at the Indian Wells Canyon section. The thick sandstone-dominated interval in the upper part of the photograph is units 109–123 of the stratigraphic section (Figure 13), and the thin-bedded limestone-dominated interval in the middle of the photograph is approximately units 88–95 of the stratigraphic section.

mudstone units (shown as dolomite in Figure 12A), mostly 0.1–0.4 m (maximum 1.8 m) thick, and thin sandstone beds are intercalated. The dolomitic mudstone beds are commonly bioturbated and contain burrows of *Rhizocorallium* and *Palaeophycus*. Thicker dolomitic mudstone intervals contain thin shale partings between the dolomitic mudstone beds. In

the lower part of subunit B, two thin sandstone beds (0.1 and 0.2 m thick), which are mixed siliciclastic-carbonate in composition and contain abundant crinoid fragments, are intercalated (unit 18).

Subunit C (Figure 12A) is approximately 23 m thick and composed of alternating shale, limestone, dolomitic limestone,

and sandstone. Limestone and dolomitic limestone intervals are abundant in the lower part, and sandstone intervals are abundant in the upper part. The base of this subunit is a 1.3 m thick rudstone bed that is cross-bedded and contains abundant crinoid fragments up to 2 cm in diameter, some brachiopod fragments, and probable glauconite grains. The rudstone is overlain by thin (0.1–0.4 m thick), mixed siliciclastic-carbonate sandstone beds alternating with shale and thin dolomitic limestone and limestone beds that are commonly bioturbated and contain chert nodules and burrows of *Palaeophycus*. In the upper part, cross-bedded crinoidal rudstone and beds of horizontally laminated and trough-cross-bedded sandstone up to 2 m thick are intercalated. The rudstone bed thickens laterally to 3 m. Also, one thin crinoidal limestone bed and a lime mudstone bed are intercalated. All of these beds alternate with shale.

Subunit D (Figure 12A) is approximately 37 m thick and composed of mudstone-shale intervals (up to 8.6 m thick) and intercalated, mostly thin (0.1–0.3 m), rarely thicker (maximum 1.1 m) limestone intervals. The thin intervals are individual limestone beds, and the thicker intervals are even- or wavy-bedded units. Intercalated sandstone beds are also present. The following limestone and sandstone lithotypes are distinguished:

1. Rudstone, partly cherty, composed of abundant crinoid fragments (up to 5 cm long), fragments of bryozoans, brachiopods, corals, sponges, and echinoid radioles
2. Crinoidal limestone (packstone to rudstone)
3. Wavy-bedded, cherty limestone containing crinoidal debris, bryozoans, and brachiopods (0.3 and 0.9 m thick)
4. Phylloid algal limestone beds in the middle of the subunit (0.1–0.6 m thick)
5. Mixed siliciclastic-carbonate sandstone containing abundant crinoid fragments and some stem fragments, displaying cross bedding and thinning laterally
6. Thin sandstone beds (0.1–0.2 m thick)

Subunit E (Figure 12A,B) is 84 m thick and is dominantly mudstone to siltstone with intercalated sandstone and one Cenozoic sill. Sandstone intercalations are both thin sandstone beds (mostly 0.1–0.2 m thick) and thicker sandstone units (up to 4 m thick). The two thicker sandstone units at the base of the subunit are trough cross-bedded and consist of mixed siliciclastic-carbonate sandstone containing abundant crinoid fragments. Thinner sandstone units (<1 m) either are massive or display ripple lamination and rare, small-scale trough cross bedding. The thicker sandstone units higher in the section (units 138 and 139) are trough-cross-bedded pebbly sandstone, are arkosic in composition, and lack carbonate grains. Indeed, as Pray (1961) noted, most Beeman sandstones are arkosic, whereas Gobbler sandstones are quartzose, with a few exceptions such as units 138–139 in our Space History B section (Figure 12B).

Subunit F (Figure 12B) is 32 m thick and composed of shale with intercalated sandstone, as well as thin and thick limestone

intervals. Pray (1961: fig. 21; here Figure 15) assigned these strata to the Bug Scuffle Member of the Gobbler Formation, and we term them the Gray Mesa Formation, so these strata are described below under the Gray Mesa Formation.

Indian Wells Canyon Section

At the Indian Wells Canyon section (Figure 13A,B), a complete stratigraphic section of the Gobbler Formation is 238 m thick and disconformably overlies the Mississippian Lake Valley Formation. The Mississippian strata are cross-bedded, coarse-grained crinoidal limestone (rudstone) with crinoid fragments up to several centimeters long and a few brachiopods. The crinoidal limestone is overlain by shale (1.4 m thick and partly covered) and red, coarse-grained, pebbly, cross-bedded sandstone containing quartz clasts up to 1 cm in diameter. These beds are followed by a thick covered interval (20 m thick) with a thin limestone bed (crinoidal wackestone) in the middle.

The stratigraphic architecture of the Gobbler Formation at the Indian Wells Canyon section differs markedly from that of the sections at Space History A and B (Figures 11, 12A,B). At Indian Wells, three thick, sandstone-dominated intervals are present within the Gobbler Formation, which is also characterized by thick covered intervals (up to 22 m thick) and shale intervals with intercalated limestones. Limestones are either individual limestone beds or thicker limestone units (maximum thickness 5.8 m).

Sandstone is present as three thick, sandstone-dominated intervals and as several thin sandstone beds:

1. The lower sandstone unit is 32.6 m thick (Figure 13A, units 12–26). The succession starts with 4.2 m of sandstone beds (0.2–1.2 m thick) alternating with shale intervals (0.3–1.2 m thick). The sandstone is fine grained, is partly mixed siliciclastic-carbonate in composition (containing brachiopods), and displays horizontal lamination and, rarely, ripple lamination. Above follow 12 m of coarse-grained, pebbly, quartz-rich sandstone with multistory trough cross beds, a covered interval (3.8 m), and 11.9 m of coarse-grained, pebbly sandstone with large-scale trough cross bedding and, subordinately, planar cross bedding. On top, a massive sandy conglomerate containing clasts up to 3 cm in diameter is exposed. The conglomerate is poorly sorted, and the clasts, all subangular, are mostly quartz, with rare black chert.
2. The middle sandstone unit is 5.1 m thick (Figure 13A, units 47–49). The succession starts with poorly sorted conglomeratic sandstone containing clasts up to 5 cm in diameter, displaying horizontal lamination and cross bedding (shallow troughs), grading upward into coarse-grained sandstone with a few pebbles up to 1 cm in size, displaying horizontal lamination and trough cross beds. Brachiopods are present on top of the sandstone unit (unit 49).

FIGURE 15. Cross section of Pennsylvanian strata from Fresno Canyon through the study area (columns 1–5) to sections farther south in the Sacramento Mountains (modified from Pray, 1961: fig. 21).

3. The upper sandstone unit measures 25.2 m thick (Figure 13B, units 109–123) and is composed of three fining upward (transgressive) cycles:
 - a. Cycle 1 (Figure 13B, units 109–112) starts with trough-cross-bedded conglomerate containing clasts up to 7 cm in diameter, overlain by trough-cross-bedded sandstone, fine-grained sandstone displaying horizontal lamination, and ripple lamination and rare, massive sandstone. The top of the cycle is formed by 3 m of dark gray shale intercalated with thin beds (5–20 cm) of dark gray lime mudstone.
 - b. Cycle 2 (Figure 13B, units 113–118) starts with sandstone displaying large-scale trough cross bedding

- (lateral accretion), with soft-sediment deformation structures in the upper part, overlain by fine-grained sandstone with small-scale cross bedding and horizontal lamination and rare massive sandstone. A thin (0.2 m thick), poorly sorted conglomerate bed with clasts up to 5 cm in diameter is intercalated. Fine-grained sandstone below this conglomerate contains crinoid fragments and brachiopods. Green shale (1.2 m) forms the top of the cycle.
- c. Cycle 3 (Figure 13B, units 119–123) starts with trough-cross-bedded sandstone containing root structures, overlain by dark gray shale (0.5 m), cross-bedded sandstone, low-angle cross-bedded

sandstone, and two thin limestone beds. This cycle is overlain by cherty limestone and cross-bedded crinoidal limestone (packstone to rudstone).

4. A thin, coarse-grained, cross-bedded sandstone is exposed near the base of the Gobbler Formation (unit 3). Thin, fine-grained, partly micaceous sandstone beds are present between 105 and 110 m above the base of the section (units 53, 57). The sandstone beds, which are 0.1–0.4 m thick, are separated by a thin covered interval (2.5 m) and a thin shale interval (0.3 m). The lowermost sandstone beds contain large crinoid fragments (up to 3 cm) and brachiopods. The micaceous sandstone contains brachiopods near the top. A thin (0.2 m), fine-grained conglomerate with clasts up to 3 cm in diameter and crinoid fragments is exposed 148 m above the base (unit 73). Two thin, fine-grained, micaceous sandstone beds, partly containing shells, are exposed 1.1 and 2.4 m above the conglomerate. Another cross-bedded, fine-grained sandstone containing crinoidal debris (2.3 m) is intercalated in cross-bedded crinoidal limestone (rudstone) approximately 173 m above the base of the formation (unit 83).

Shale is rarely exposed in the lower 170 m of the Indian Wells Canyon section, where thick covered intervals (up to 21.8 m thick) most likely represent shale intervals. In the upper part of the Gobbler Formation, the exposed shale units are up to 7.2 m thick. A typical facies is thin, muddy limestone beds (0.1–0.3 m thick) alternating with thin shale layers (0.1–0.3 m thick). The shale is mostly dark gray and subordinately black and greenish gray.

Limestone is present as individual limestone beds and as thicker limestone units:

1. Individual limestone beds are 0.1–0.6 m (mostly 0.1–0.3 m) thick. Most of the limestone beds are composed of dark gray lime mudstone (bioclastic mudstone, wackestone, rare floatstone) locally containing crinoid fragments and brachiopods. Some of the thicker limestone beds are composed of crinoidal rudstone.
2. Thicker limestone units are 0.8–5.8 m thick, thin to medium bedded, and, rarely, indistinctly bedded to massive. Most common is even bedding, and in the upper part, wavy bedding is also present, partly with thin shale partings between the limestone beds. Chert is rare and present in a few limestone units as small nodules in the upper part of the section. Most common is crinoidal limestone (crinoidal wackestone, packstone, and rudstone), which contains other fossils subordinately, such as brachiopods and bryozoans. Less common is lime mudstone (bioclastic wackestone, crinoidal wackestone); the wavy-bedded limestone is of this type.

SEDIMENTARY PETROGRAPHY

We undertook petrographic analysis of selected sandstone and limestone samples of the Gobbler Formation in the Space

History A and B sections (Figures 16, 17). In the lower part of the Gobbler Formation (Space History A section), sandstone is poorly sorted and composed of mono- and polycrystalline quartz grains, and many small, brownish, rounded grains (phosphoritic/chloritic), embedded in a brownish groundmass (phosphoritic/chloritic). Many detrital quartz grains display authigenic overgrowths. Detrital feldspars and rock fragments are absent.

Stratigraphically higher in the Gobbler Formation (Space History B section), sandstone is mostly moderately sorted and composed of subangular to subrounded grains. Most common are quartz grains (mono- and polycrystalline quartz), and detrital feldspars are also common (plagioclase and potassium feldspars, including microcline). Also present are some chert grains, a few altered grains (chlorite?), a few granitic rock fragments, fine-grained carbonate grains (sedimentary rock fragments), and a few opaque grains and rare micas (muscovite). The sandstone contains some fossil fragments, mostly of echinoderms, and some skeletal fragments of brachiopods, bryozoans, smaller foraminifers, and gastropods (Figure 16).

Many quartz grains display authigenic overgrowths (Figure 16E,H), and the sandstone is cemented by coarse blocky calcite that randomly replaces detrital feldspar grains (Figure 16A,C,D,F). Mixed siliciclastic-carbonate sandstone shows similar textural properties but contains a higher number of carbonate grains and fossil fragments, including echinoderms, brachiopod shells and spines, a few ostracods, fusulinids, rare smaller foraminifers, and bryozoans (Figure 16A,B,C,D,F). Locally, there is some micritic matrix. In a QFL (quartz, feldspar, lithics) diagram (Pettijohn et al., 1987; Figure 18), the Gobbler Formation sandstones show a wide range, most of them plotting into the field of lithic arenite and sublitharenite and rarely into the field of subarkose.

Limestone microfacies types in the lower part of the Gobbler Formation (Space History A section) include bioclastic wackestone (Figure 17C,D) and packstone (Figure 17F), which may grade into floatstone and rudstone (Figure 17E,H). All microfacies contain a diverse fossil assemblage of echinoderm fragments, bryozoans, brachiopods, ostracods, foraminifers, rare gastropods, echinoid radioles, trilobites, and probable recrystallized algae, in addition to many completely recrystallized skeletons. In rudstone and floatstone, some of the skeletal grains are encrusted by *Palaeonubecularia*, rarely forming oncoids. Locally, peloids, micritic intraclasts, and small detrital quartz grains are present.

In the Space History B section, the limestone is composed of various microfacies types such as bioclastic mudstone, grading into bioclastic wackestone (including ostracodal mudstone to wackestone, which is rare; Figure 17A,B), bioclastic wackestone to packstone, bioclastic packstone to grainstone, bioclastic packstone to rudstone, grainstone, algal floatstone, and rare bindstone. In the uppermost 32 m of that section, limestones contain abundant phylloid algae (phylloid algal wackestone and floatstone). Some beds have abundant *Archaeolithophyllum* floatstone (Figure 17G) to rudstone containing large, mostly recrystallized algal thalli of *A. missouriense*. Mudstones are characterized by a low- to moderate-diversity fossil assemblage.

Most common are ostracods, and echinoderms, bivalves, gastropods, and spicules are subordinate (Figure 17A,B).

All other microfacies types contain a diverse fossil assemblage that is commonly dominated by echinoderms, bryozoans, and brachiopod fragments. Individual limestone beds contain abundant foraminifers (foraminiferal wackestone), phylloid algae (phylloid algal floatstone), *Archaeolithophyllum missouriense*, or ostracods (ostracodal wackestone). Other fossils present are sponge spicules (Figure 17C), which locally may be abundant, gastropods, bivalves, trilobites, and many recrystallized skeletons.

Nonkeletal grains are present in some of the wackestones, packstones, and grainstones, including peloids (partly forming peloidal grainstone and packstone), micritic intraclasts, and rare ooids. Rarely, skeletal grains in wackestone are encrusted by foraminifers. Bioturbation is observed in mudstone and wackestone. Angular to subangular small quartz grains are present in small amounts in most of the microfacies types. Rarely, micas, detrital feldspars, and glauconitic grains are present.

PALEONTOLOGY AND AGE

The Gobbler Formation yields numerous and diverse fossils that merit further study. Marine limestone and shale of the Gobbler Formation contain fossils of the typical Pennsylvanian shelly benthos, mostly crinoids and brachiopods and less common bryozoans, mollusks, and trilobites. Trace fossils are common in some beds, notably *Palaeophycus* and *Rhizocorallium*, which dominate low-diversity ichnoassemblages. Fossil wood (mostly oxidized fragments) is present in some sandstone beds of the Gobbler Formation, and fossil plant foliage is found as fragmentary debris in some shale beds (see section on Paleobotany of the Gobbler, Beeman, and Holder formations).

On the basis of fusulinids, primarily identifications by M. L. Thompson (though none of these were specified or documented), Pray (1961) assigned the Gobbler Formation a Morrowan? to middle Missourian age. Boardman et al. (1994; see also Raatz et al., 1994:63, fig. 2) listed the ammonites *Neoaganides*,

?*Eoschistoceras*, and *Glaphyrites* from the upper part of the Gobbler Formation in the Indian Wells Canyon area. However, Boardman and Work (2013) later indicated that the lowest occurrence (LO) of *Neoaganides* is Missourian, so that identification may need to be revised (see later discussion). We attempted to locate ammonites in the upper part of the Gobbler Formation in the Indian Wells Canyon area but were unsuccessful.

The only fossils from the Gobbler Formation we sampled extensively and studied are calcareous microfossils and conodonts, which include biostratigraphically important assemblages of Morrowan, Atokan, and Desmoinesian ages (Table 3; see Marine Microfossils section).

DEPOSITIONAL SYSTEMS

It is important to recognize that previous analyses of Gobbler Formation deposition (Pray, 1961; Van Wagoner, 1977a, 1977b; Wilson, 1989; Algeo et al., 1991) included the Bug Scuffle Member in the Gobbler Formation, so parts of these analyses (particularly much of their identification of cyclic deposition) were based on the Bug Scuffle carbonate-dominated facies. Because we remove the Bug Scuffle Member from the Gobbler Formation and assign it to the Gray Mesa Formation (see below), we focus here on deposition of the clastic-dominated Gobbler Formation, as redefined here.

Sediments of the Gobbler Formation were deposited on the eastern margin of the Orogrande Basin. This basin margin was a narrow, high-relief shelf termed the Sacramento shelf, which was bounded to the east (Pedernal uplift) by a major fault (growth fault; Figure 3). Gobbler strata are characterized by rapid lateral and vertical facies changes, lateral discontinuity of units, and, where cycles are developed, irregular cycle thicknesses (see section on Gobbler Formation lithostratigraphy above).

Both Pray (1961) and Van Wagoner (1977a, 1977b) concluded that Gobbler deposition took place in a shallow marine shelf setting. Van Wagoner (1977a, 1977b) identified 10 carbonate facies and 7 siliciclastic facies, the former in the "Bug

FIGURE 16. (Opposite) Thin-section photographs of sandstones from the Gobbler Formation, Space History section. B and D under plane light; A, C, E–H under polarized light. Width of photographs: A, B, E–H = 3.2 mm; C, D = 6.3 mm. (A) Mixed siliciclastic-carbonate sandstone composed of detrital quartz grains, a few feldspars (partly replaced by calcite), sedimentary rock fragments, and many fossil fragments (mainly echinoderms, also fusulinids, ostracods, and other skeletal grains), cemented by calcite. Sample SH 1. (B) Mixed siliciclastic-carbonate sandstone (rudstone) containing quartz grains, a few feldspars, sedimentary rock fragments, and many fossils (echinoderms, bryozoans). Sample SH 7. (C, D) Mixed siliciclastic-carbonate sandstone (rudstone) composed of quartz grains, a few feldspars, sedimentary rock fragments, and many fossils (mostly echinoderms, subordinately bryozoans, ostracods, and recrystallized skeletons), cemented by coarse blocky calcite. Sample SH 1. (E) Sandstone composed of abundant quartz grains, a few feldspar grains that are partly replaced by calcite, and a few echinoderm fragments. Detrital quartz grains display authigenic overgrowths; some calcite cement is present too. Sample SH 43. (F) Mixed siliciclastic-carbonate sandstone containing quartz grains, many detrital feldspars (mostly potassium feldspars), a few granitic rock fragments, sedimentary rock fragments, and some fossil fragments (mostly echinoderms). The sandstone is cemented by coarse blocky calcite. Sample SH 44. (G) Sandstone composed of quartz, feldspars (including microcline), granitic rock fragments, a few carbonate grains (sedimentary rock fragments), and some fossil fragments (echinoderms, brachiopods). Sample SH 82. (H) Sandstone composed of abundant quartz grains, a few feldspars, rare chert, and rock fragments (including granitic rock fragments). Some of the detrital quartz grains display authigenic overgrowths. Sample SHA 28.

Scuffle” interval. He noted that “vertical facies relationships are dominantly cyclic” (p. 57) and that the carbonate cycles shoal upward, whereas clastic-carbonate cycles are symmetrical. Van Wagoner (1977a, 1977b) thus saw the primary drivers of Gobbler deposition as sea-level fluctuation and shelf paleotopography.

Wilson (1989) identified Morrowan-Desmoinesian “cyclothems” (upward-shoaling cycles) in the Gobbler and “Bug Scuffle” strata. However, he noted that “the Derryan and Desmoinesian cyclothems . . . appear internally consistent but are not clearly correlative over tens of km from section to section” (p. 13). Nevertheless, Algeo et al. (1991) presented just such cycle correlations over many tens of kilometers in the San Andres and Sacramento Mountains, though their study is primarily of the Gray Mesa Formation interval, so it is discussed below.

The main lithologies of the Gobbler Formation at the Space History A and B sections (Figures 11, 12A,B) are as follows: (1) shale/mudstone; (2) intercalated limestone and dolomitic limestone; and (3) sandstone, mostly mixed siliciclastic-carbonate, fossiliferous (rudstone), and cross-bedded. Shale most likely represents deposits of a deeper shelf environment below the storm wave base (“offshore shale”). Limestone and dolomitic limestone are composed of different microfacies types, most of them containing a diverse fossil assemblage, indicating accumulation in a shallow, open-marine, low- to moderate-energy shelf setting. Muddy limestone types with a low-diversity fossil assemblage, dominantly of ostracods, are interpreted as deposits of a shallow, restricted (lagoon) environment. Mixed siliciclastic-carbonate sandstone with highly fragmented fossils (rudstone), which are commonly cross-bedded, indicate deposition in a high-energy, nearshore environment.

The facies of the Gobbler sections we studied are very similar to those described by Algeo et al. (1991). However, we do not recognize shoaling upward cycles in the lower 115 m of the Space History A section (Figure 11), probably because of thick covered intervals. At the Space History B section (Figure 12A,B), some shoaling upward cycles are developed, which are mostly 5–12 m thick and start with offshore shale, grading into shale

with intercalated open-marine limestone and mixed siliciclastic-carbonate, fossiliferous, cross-bedded sandstone on top of the cycle (see Gobbler Formation Lithostratigraphy section). A few cycles are entirely clastic, starting with shale, grading into shale with intercalated thin sandstone beds and finally into thicker, cross-bedded or ripple-laminated sandstone. Thus, cycles are not developed throughout the Gobbler Formation at the Space History B section, but there are a few segments where shoaling upward cycles are present (e.g., units 125–130).

The area of deposition of the upper, clastic-dominated interval of the Gobbler Formation, the “detrital facies” of Pray (1961), has long been referred to as the Alamo clastic trough (Meyer, 1966; Van Wagoner, 1977a, 1977b; Algeo et al., 1991). Coarse-grained terrigenous clastic sediments of the upper Gobbler Formation filled the Alamo clastic trough, which was developed locally on the Sacramento shelf, whereas most of the limestones of the Gray Mesa Formation (Bug Scuffle Limestone Member of previous workers) formed outside of the trough, on the shelf and basinal slope (Algeo et al., 1991).

Thus, Van Wagoner (1977b:59) noted the “presence of a linear, northwest trending, topographic depression slightly deeper than the contiguous shelf . . . down which rivers and streams draining the Pedernal uplift could flow debauching sediment basinward through a series of prograding delta lobes.” Van Wagoner (1977a, 1977b) reported that paleocurrent data of the upper part of the Gobbler Formation indicated the northwesterly flow of prograding delta lobes sourced in the Pedernal uplift. He also noted that the Alamo clastic trough was a northwest-trending feature likely controlled structurally by penecontemporaneous faulting. Therefore, it represents a tectonic feature that influenced Gobbler/Gray Mesa deposition.

There likely was a climatic component to the alternating siliciclastic and carbonate stratigraphic sequences in the Gobbler Formation. Paleosol studies in the region indicate that the climate in western Pangea during the Pennsylvanian was monsoonal, thus strongly seasonal (e.g., Tabor et al., 2008). In the terminology of Cecil (2003), the climate was subhumid to

FIGURE 17. (*Opposite*) Thin-section photographs of different microfacies types of limestones from the Gobbler Formation near the New Mexico Museum of Space History. All photos under plane light. Width of photographs: A, C, E–H = 9 mm; B, D = 4.6 mm. (A) Ostracod mudstone to wackestone composed of ostracods and a few larger shell fragments embedded in micrite. Sample SH 102. (B) Bioclastic mudstone-wackestone composed of recrystallized skeletons, spicules, ostracods, and echinoderms embedded in microsparitic matrix. Sample SH 32. (C) Bioturbated wackestone containing abundant spicules, a few ostracods, echinoderms, and small detrital quartz grains. Sample SH 54. (D) Wackestone to poorly washed grainstone, bioturbated, composed of echinoderm fragments, subordinately of ostracods, smaller foraminifers, and small detrital quartz grains. Sample SH 75. (E) Bioclastic wackestone to floatstone containing a diverse fossil assemblage of echinoderms (crinoids), bryozoans, brachiopods, ostracods, smaller foraminifers, trilobites, and a few other skeletons. Sample SH 94. (F) Wackestone to packstone composed of abundant micritic intraclasts, peloids, recrystallized skeletons, echinoderms, rare ostracods, and ooids. Sample SH 19. (G) Algal floatstone containing large thalli of *Archaeolithophyllum missouriense* embedded in peloidal micrite containing a few small fossils. Sample SH 98. (H) Rudstone to floatstone containing abundant larger echinoderm (crinoid) fragments, bryozoans, brachiopods, ostracods, smaller foraminifers, rare trilobites, and small detrital quartz grains. Sample SHA 18.

FIGURE 18. QFL diagram of sandstones of the Gobbler, Beeman, and Holder Formations.

semiarid, even perhaps arid at times. Studies of modern tropical rivers (Cecil et al., 2003a) indicate that a climate favorable to sediment transport is characterized by seasonality of rainfall, tapering off as climates become either humid to perhumid, when plant cover and rooting strongly bind soils, or arid, when there is insufficient moisture to move large volumes of sediment, except as dust. Siliciclastic deposition, under this model, would most likely occur during periods of seasonal rainfall, whereas carbonate formation would prevail at times when and places where climate was trending toward arid.

If these climates were tied to glacial–interglacial cycles, which prevailed during the Middle and Late Pennsylvanian (Heckel et al., 2007; Fielding et al., 2008; Heckel, 2008, 2013; Eros et al., 2012), some sedimentologic expression might be

expected, even if obscured locally to regionally by the overprinting of tectonic activity. Furthermore, even if many glacial–interglacial cycles were insufficient to cause major sea-level fluctuations (e.g., Isbell et al., 2003), they still may be expected to have generated detectable climatic fluctuations in the tropics because of their effects on atmospheric circulation (Cecil et al., 2003b; Peyser and Poulsen, 2008; Horton et al., 2012), even in mountainous regions (e.g., Opluštil et al., 2015).

Even considering these climatic factors, the case for glacioeustasy as the primary driver of Gobbler deposition appears to be a weak one. The formation generally lacks the organized facies stacking and the lateral facies continuity that identify an allocyclic driver of deposition, particularly glacially driven, eustatic sea-level fluctuations. Some cycles are evident in the

TABLE 3. Conodont age determinations and correlations to Midcontinent units and cycles.

Stratigraphic unit	Level(s) in stratigraphic section(s) ^a	Age and Midcontinent correlation
Holder Formation	Lower Holder cycles J–I2 of Cox (1998), Keairns (2002)	No older than earliest Gzhelian/early Virgilian (Oread cyclothem)
Beeman Formation	La Luz B (33t, 47)	No older than earliest Gzhelian/early Virgilian (Oread cyclothem)
	Dry Canyon (Raatz 2C5a)	Late Missourian (South Bend–Iatan cyclothem)
	Space History B (245, 259, 267) [122–135 m above base]; Dry Canyon (Raatz 2C-4a)	Late middle Missourian (about Iola, Wyandotte, Plattsburg cyclothem)
	Space History B (229, 235) [105–108 m above base]	Middle Missourian (Dewey cyclothem)
	Space History C (44) [65 m above base of formation]; Space History B (212) [87 m above base]	Middle Missourian (Cherryvale up to Dewey cyclothem)
	No conodonts from basal Beeman; by stratigraphic position	Early Missourian (Dennis and Hogshooter cyclothem)
Gray Mesa Formation (Space History Member)/ Gobbler Formation	Space History B (155, 160) [22–10 m below top of Gobbler]; Space History C (1) [12–10 m below top of Gobbler]; Beeman Canyon A (2, 4) [12–10 m below top of Gobbler]	Early Missourian (Swope cyclothem)
Gray Mesa Formation (Space History Member)	Indian Wells Canyon (134) [20 m below top of Space History Member]	Latest Desmoinesian (late Marmaton)
Gobbler Formation	Space History B (75, 82, 84, 86, 118) [83–218 m]	Late Desmoinesian (Marmaton)
	Space History B (41, 43, 46) [64 m]; Indian Wells Canyon (93) [197 m]; High Rolls (38)	Early late Desmoinesian (early Marmaton)
Gray Mesa Formation (main body)	High Rolls (2), LaPeter (2007)	Late early Desmoinesian (late Cherokee)
Gobbler Formation	Space History B (18) [39 m]	Early Desmoinesian (early–middle Cherokee)
	Space History B (1, 3, 9) [1–25 m]	Earliest Desmoinesian (earliest Cherokee)
	Space History A (SHA 36) [93 m]; Indian Wells Canyon (IWC 51, 62, 64) [97–139 m]	Near Atokan–Desmoinesian boundary
	Space History A (SHA 26) [48 m]	Late Atokan
	Mule Canyon (MC 35) [52 m]; Space History A (SHA 16, 18, 20) [33–43 m]; Indian Wells Canyon (IWC 7, 9) [22–25 m]	Middle–late Atokan
	Mule Canyon (MC 18, MC 26) [19–35 m]	Late Morrowan–early Atokan
	Space History A (SHA 3) [5 m above base]; Dog Canyon, Alamo Canyon, Grapevine Canyon (Lane, 1974); Deadman Canyon [10 m above base] (Benne, 1975)	Early Morrowan

^a Unit numbers in the stratigraphic section are in parentheses, and meters above base of stratigraphic section are given in brackets.

Gobbler, and no doubt glacioeustasy was a factor in some aspects of Gobbler deposition. Both local tectonics and climate changes, and also allogenic factors, were likely drivers of some or much of Gobbler deposition, and climate fluctuations likely would have been tied to glacial cycles, even if the sea-level expression thereof was muted or overprinted by tectonic events. In addition, because climate in the region where

the Sacramento Mountains are now located largely oscillated between subhumid and semiarid, the potential for significant sediment transport was often present, which introduces the likelihood for strong autogenic overprinting of allogenic effects. A careful reevaluation of the Gobbler Formation is needed to sort out the possible effects of glacioeustasy, local climate, and tectonics on its deposition.

GRAY MESA FORMATION

LITHOSTRATIGRAPHY

Pray (1961:82) named the Bug Scuffle Limestone Member of the Gobbler Formation for a cliff-forming, limestone-dominated stratigraphic interval above the lower 91–152 m of the Gobbler Formation (Figure 4). He stated that the Bug Scuffle Member is up to 305 m thick (though usually 152–213 m thick) and mappable at a scale of 1:24,000. Pray (1961: fig. 21) showed the Bug Scuffle Member as the upper one-half to two-thirds of the Gobbler Formation south of Long Ridge (Figure 15). In much of the Sacramento Mountains, the Bug Scuffle forms steep cliffs (Figure 19) along a north–south outcrop belt that is about 50 km long. To the north of Long Ridge, Pray (1961) showed it fingering out into what he termed the “detrital facies” of the Gobbler Formation (Figure 15). Indeed, our sections at Space History B (Figure 12B) and Indian Wells Canyon (Figure 13B) include what we identify as a relatively thin tongue of the upper part of the Bug Scuffle Member (Gray Mesa Formation). Substantial outcrops of the Bug Scuffle Member are also present near High Rolls in the northeastern corner of the study area (Figures 2, 20).

Pray (1961) considered the Bug Scuffle Member to be the upper member of the Gobbler Formation, laterally equivalent

to and locally overlain by his “detrital facies” of the Gobbler. However, we are struck by the mappability of the Bug Scuffle Member and its resemblance in lithology, stratigraphic position, and age overlap with the unit called Gray Mesa Formation (originally named by Kelley and Wood, 1946) across much of central New Mexico (Kues, 2001; Nelson et al., 2013; Lucas et al., 2016a). Note that Pray (1961: pl. 3) mapped the Bug Scuffle Member at a scale of 1:31,680, so it fulfills the operational definition of mappability of a formation-rank unit.

The Bug Scuffle Member overlies Gobbler strata that are homotaxial with the Red House and Sandia Formations to the west and northwest, and it underlies Beeman strata that are homotaxial with the lower part of the Panther Seep Formation to the west and Atrasado Formation to the northwest—it thus occupies the same stratigraphic position as the Gray Mesa Formation to the west (between Red House and Panther Seep) and northwest (between Sandia and Atrasado; Nelson et al., 2013). The lithology of the Bug Scuffle Member is very much the same as that of the Gray Mesa Formation in being dominantly indurated beds or intervals of beds of limestone, many of them cherty, with thinner interbeds of nodular limestone or shale, so that it weathers to cliffs or a step-and-ledge topography (Figure 19; Lucas et al., 2016a, 2016b; Nelson et al., 2013). The Bug Scuffle Member is primarily of late Desmoinesian age (its uppermost strata are Missourian), which is the same age as some of the

FIGURE 19. Gray Mesa Formation outcrop photos. (A) High Rolls section (part). (B) Section in Mule Canyon.

FIGURE 20. High Rolls section. See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "HRO 19").

Gray Mesa Formation to the west (San Andres Mountains) and northwest (Oscura Mountains; Kottlowski et al., 1956; Bachman and Myers, 1969, 1975; Lucas and Krainer, 2009; Nelson et al., 2013).

The lateral relationship of the Bug Scuffle Member to the detrital facies of the Gobbler Formation is very similar to the lateral relationship the Gray Mesa Formation has to similar clastic-dominated facies of the Sandia Formation in the Los Pinos Mountains of northern Socorro County (Kottlowski, 1960; Krainer et al., 2017a). There, a substantial interval of clastic-dominated strata of the Sandia Formation is of Desmoinesian age and laterally equivalent to strata of the Gray Mesa Formation to the south and the north (Krainer et al., 2017a).

Working in the San Andres Mountains (Figure 1), Bachman and Myers (1969:C22; see also Bachman and Myers, 1975) named the Lead Camp Limestone for the “massive cliff-forming cherty limestone beds that unconformably overlie the Mississippian and underlie the rocks of the Panther Seep Formation.” However, their Lead Camp Limestone is actually a composite unit that includes an underlying ~75 m of Pennsylvanian shale-dominated strata, later assigned to the Red House Formation (Kalesky, 1988; Lucas et al., 2012). As Bachman and Myers (1975) made clear, those clastic strata are equivalent to the Sandia Formation to the north, and the upper limestone interval of the Lead Camp is equivalent to the “Los Moyos Limestone” (=Gray Mesa Formation) to the north (Nelson et al., 2013; Lucas et al., 2016b).

Given its lithologic distinctiveness, relatively great thickness, and mappability, the Bug Scuffle Member should be of formational rank. Furthermore, the Bug Scuffle Member clearly is the same lithosome as the Gray Mesa Formation to the northwest. Therefore, we abandon the name Bug Scuffle Limestone Member and refer to this unit as the Gray Mesa Formation.

Nevertheless, an alternative interpretation by one of us (JEB) of the cliff-forming Desmoinesian limestones (“Bug Scuffle”) assigned here to the Gray Mesa Formation is that the “Bug Scuffle” represents a separate lithosome on the eastern side of the Orogrande Basin that is younger (late Desmoinesian) than most of the cliff-forming units of the Gray Mesa (mostly early Desmoinesian) on the western side of the Orogrande Basin. According to this view, the variable thickness and irregular distribution of the Bug Scuffle is largely the result of depositional and tectonic events along the western edge of the Pederal uplift and unrelated to events on the western side of the Orogrande Basin. The Bug Scuffle would then be retained as a member within the Gobbler Formation, as originally proposed by Pray (1961).

Furthermore, the ~30 m thick interval of carbonates at the top of the Gobbler Formation in the Space History B, Space History C, and Beeman Canyon A sections was assigned to the Bug Scuffle by Pray (1957; Figure 15), probably because it formed a small cliff below the level at which the arkosic sandstones of the Beeman appeared. Conodonts date this interval as

crossing the Desmoinesian-Missourian boundary, which makes it apparently younger than most or all of the thicker Desmoinesian Bug Scuffle in other sections. However, note that we have limited age control of the upper Bug Scuffle at High Rolls, the only thick section of that unit we studied in detail. JEB suggests that this carbonate interval in the Space History and the Beeman Canyon A sections does not comprise any portion of the Bug Scuffle and that it is a separate stratigraphic unit higher in the section.

Here, we name this relatively thin interval the Space History Member of the Gray Mesa Formation (or it could be assigned to the Gobbler Formation, according to JEB) for the New Mexico Museum of Space History just south of Horse Ridge. The type section of the Space History Member is units 147–169 of our Space History B section (Figure 12B) and is described below.

We measured sections of the Gray Mesa Formation at High Rolls (Figure 20), Space History B (Figure 12B), Indian Wells Canyon (Figure 13B), and Beeman Canyon A (see Figure 25, Table 1). In the Sacramento Mountains, the Gray Mesa Formation is a limestone-dominated unit that laterally interfingers with the Gobbler Formation (Table 2). The Gray Mesa Formation is 190 m thick at the High Rolls section, where the contacts with the underlying Gobbler Formation and the overlying Beeman Formation are well exposed. At High Rolls, the Gray Mesa Formation is mainly composed of different types of noncherty and cherty, thin-bedded to thick-bedded and massive, fossiliferous limestone. Shale-covered intervals rarely exceed 5 m in thickness and are less abundant compared to those in the Gobbler Formation. A few shallowing upward cycles are developed (see Gray Mesa Formation Depositional Systems).

High Rolls Section

We measured a complete stratigraphic section of the Gray Mesa Formation at the tunnel on U.S. Highway 82 near High Rolls on the northern side of the canyon (Figures 2, 19A, 20). As elsewhere in the Sacramento Mountains, the Gray Mesa Formation at High Rolls forms cliffs and steep slopes with the characteristic step-and-ledge topography between the slope-forming Gobbler Formation below and the slope-forming Beeman Formation above (Figure 19). At the High Rolls section, the Gray Mesa Formation is ~190 m thick and mostly limestone (76% of the measured section; Figure 20). Cherty limestone (mostly wackestone) comprises 32% of the section, and algal limestone is nearly as common (27% of the section). The remaining limestone units are other types of wackestone and minor packstone and grainstone (17% of the section). The following limestone lithotypes are distinguished (Figure 20):

1. Thin-bedded limestone with even bedding, bed thickness of 5–20 cm, and a muddy texture (bioclastic

mudstone to wackestone), partly containing phylloid algae and crinoids and rarely brachiopods

2. Medium-bedded limestone (20–50 cm thick beds), mostly composed of bioclastic wackestone to packstone, containing crinoid fragments, phylloid algae, brachiopods, and rare fusulinids
3. Indistinctly thick bedded to massive limestone containing crinoidal debris, phylloid algae, solitary corals, and brachiopods (wackestone to packstone, partly floatstone to rudstone)
4. Wavy-bedded to nodular limestone (bed thickness mostly 10–20 cm) with a muddy texture (wackestone), partly crinoidal, rarely containing solitary corals
5. Nodular limestone (rare), composed of bioclastic mudstone to wackestone

In the middle and upper parts of the Gray Mesa Formation at High Rolls, chert is common and forms nodules that are mostly a few centimeters in diameter and, in some of the limestone beds, up to several tens of centimeters in diameter.

Indian Wells Canyon Section

In our Indian Wells Canyon section (Figure 13B), the Gray Mesa Formation is only about 38 m thick and separates the upper part of the Gobbler Formation from the Beeman Formation (Figure 13B, units 124–138).

The Gray Mesa Formation base is cherty limestone that rests on sandstone with low-angle cross bedding (0.6 m) and is overlain by cross-bedded sandstone containing abundant crinoid fragments, a thin shale interval (0.3 m), and 20–30 cm of bedded, dark gray, muddy limestone containing a few echinoderm fragments.

Above a covered interval of 6.9 m follows a succession of thin, micritic limestone beds that are 10–20 cm thick and separated by thin shale intervals (2.5 m). This unit is overlain by 2.8 m of limestone beds (0.3–0.6 m thick) that are separated by shale intervals with a thickness of up to 0.4 m. The limestone beds are composed of packstone and rudstone containing abundant crinoid fragments and brachiopod shell fragments. The lower two limestone beds contain chert nodules.

This interval is overlain by 2.3 m of thick-bedded to massive limestone with little chert composed of rudstone with abundant crinoid fragments and shells and 3.5 m of medium-bedded (20–40 cm thick) dark gray limestone (mostly crinoidal wackestone), followed by a 4.5 m thick succession of dark gray micritic limestone beds (10–20 cm thick) alternating with dark shale intervals (10–20 cm thick). The next unit is 6.8 m thick and composed of dark shale with one thin limestone bed intercalated. The uppermost part of the Gray Mesa Formation is composed of wavy-bedded, bioturbated cherty limestone (mostly wackestone, 3 m thick), with thin shale layers between the limestone beds.

Space History B Section

As noted above, subunit F of our Space History B section (Figure 12B) was identified by Pray (1961: fig. 21) as two tongues or lenses of Bug Scuffle limestone beds, separated by detrital facies of the Gobbler Formation. This interval is about 50% clastic and 50% carbonate, and most of the limestone beds are algal. We assign all of it (units 149–169) to the Gray Mesa Formation as the type section of the Space History Member of the Gray Mesa Formation (named for the nearby museum of that name).

The intercalated sandstone beds are massive or horizontally laminated and 0.1–0.3 m thick (Figure 12B, unit 154). Thin limestone beds are up to 1.2 m thick, even bedded or nodular, and partly cherty (units 155–159). Two limestone beds contain abundant crinoid fragments. Thicker limestone intervals are composed of indistinctly thick bedded to massive limestone containing phylloid algae, rare brachiopods, and gastropods. On top of one phylloid algal limestone (unit 165), a black chert layer is exposed that is 0.2 m thick and contains silicified fossils. On top of the uppermost phylloid algal limestone (unit 169), a subaerial exposure surface marked by rhizoliths and brecciation is developed. This is the mappable contact of the Gray Mesa Formation and overlying Beeman Formation (Figure 10; see also Figure 28).

Beeman Canyon A Section

The Beeman Canyon A section is an incomplete section of the Gray Mesa Formation that is ~15 m thick (see Figure 27). We correlate this section to the upper part of the Gray Mesa Formation at our Space History B section (Figure 12B, units 161–170). At the Beeman Canyon A section, the lower 3 m are mostly crinoidal limestone capped by a subaerial exposure surface. Black and gray shale follows, then a succession of bioclastic and algal limestone capped by a crinoidal limestone with a bioturbated top. This section can be interpreted to represent two shallowing upward cycles, units 1–5 followed by units 6–17 (see Figure 27).

SEDIMENTARY PETROGRAPHY

We undertook petrographic study of limestone samples of the Gray Mesa Formation in our High Rolls and Space History B sections (Figures 21–23). At the High Rolls section, limestones of the Gray Mesa Formation are composed of different types of wackestone (bioclastic wackestone [Figures 21A–E, 23G,H], *Komia* wackestone [Figure 22A,C], phylloid algal wackestone, crinoidal wackestone, spiculite), which grade into packstone (bioclastic packstone, *Komia* packstone [Figure 22B,D,E]), grainstone (*Komia* grainstone [Figure 22F,G]), floatstone (bioclastic floatstone [Figure 21H], crinoidal floatstone, phylloid algal floatstone [Figure 23A–D], *Komia* floatstone), and rudstone (crinoidal rudstone [Figure 21F,G]; *Komia* rudstone [Figure 22H]). *Komia* is present in the lowermost 38 m of the High Rolls

section. In the Gray Mesa Formation limestone beds, the most common fossils are echinoderms (crinoid fragments), bryozoans, brachiopods, locally, *Komia*, spicules, and red algae (*Archaeolithophyllum missouriense*; Figure 23E,F). Ostracods, smaller foraminifers, fusulinids, gastropods, trilobites, brachiopod spines, echinoid radioles, and rare calcisponges are subordinate. Some of the limestones are partly silicified, containing silicified fossils and chalcedony cement.

At the Space History B section, limestones of the Gray Mesa Formation are composed of bioclastic wackestone, grading into packstone and floatstone, phylloid algal wackestone to floatstone, and *Archaeolithophyllum* floatstone to rudstone (Figure 23A–F). All of these limestone types contain a diverse fossil assemblage, as at the High Rolls section. Thin- to medium-bedded limestones are commonly composed of bioclastic wackestone to floatstone, and indistinctly thick bedded to massive limestone is mostly phylloid algal wackestone to floatstone.

PALEONTOLOGY AND AGE

The Gray Mesa Formation in the Sacramento Mountains is very fossiliferous. Macrofossils are dominated by typical late Paleozoic shelly benthos: brachiopods, crinoids, and bryozoans. Less common macrofossils are of mollusks (mostly gastropods) and trilobites. Microfossils seen in thin section are mostly algae and foraminifers (including fusulinids), discussed below. We also extracted conodonts from some of the Gray Mesa Formation limestone beds (see below).

Pray (1961) assigned the Gray Mesa Formation (his Bug Scuffle Member of the Gobbler Formation) a Desmoinesian to early Missourian age on the basis of fusulinids that were not documented. LaPeter (2007) equated the Gray Mesa Formation at High Rolls to the Desmoinesian Cherokee and Marmaton groups in the Midcontinent, although the biostratigraphic data available to him for such a correlation was limited. Our micropaleontological data, based on fusulinids and conodonts, indicate an age range of Desmoinesian to early Missourian (Table 3) and are discussed at greater length below.

DEPOSITIONAL SYSTEMS

Pray (1961:83) stated that strata of the Gray Mesa Formation (his Bug Scuffle Member) mostly “formed in relatively shallow marine water of low turbulence.” As noted above, Algeo et al. (1991) identified cycles in the Gobbler Formation, but these are in strata of the Bug Scuffle Member of Pray (1961), so we discuss them here.

According to Algeo et al. (1991), the Gray Mesa Formation in the San Andres and Sacramento Mountains is composed of shallow marine sediments displaying 20–25 carbonate shoaling upward cycles. The shallow marine facies consist of fossiliferous wackestone and packstone containing an open-marine fossil assemblage. Argillaceous mudstones and wackestones represent deepwater environments and contain an in situ fauna of sponges and long-spined brachiopods. Cross-bedded fossiliferous grainstones formed in shallow shelf areas and occur at the top of shoaling upward cycles. Algal-plate wackestones are present as biostromes and, locally, as low-relief bioherms. The core facies is composed of massive limestone containing unbroken phylloid algae, bryozoans, and *Komia*. The flanking-bed facies is composed of fossiliferous wackestone and packstone with abundant fragments of phylloid algae. Cycle tops of the shoaling upward cycles indicate subaerial exposure. Cycles are 3–20 m thick. Though lacking detailed age control, Algeo et al. (1991) stated that cycle periods are in the range of 140,000–150,000 years, and the cycles resulted from high-frequency, high-amplitude glacioeustatic sea-level fluctuations.

However, a close examination of the cycles identified and their correlation by Algeo et al. (1991: figs. 4, 5) suggests the need for a reevaluation:

1. In their stratigraphic sections on the White Sands Missile Range in the Mockingbird Gap Hills and at Rhodes Canyon and Hembrillo Canyon (in the San Andres Mountains; Figure 1), Algeo et al. (1991: fig. 4) place the Atokan-Desmoinesian boundary very low in

FIGURE 21. (*Opposite*) Thin-section photographs of microfacies types from limestones of the Gray Mesa Formation in the High Rolls section. All photos under plane light. Width of photographs: A–C, F–H = 9 mm; D, E = 4.6 mm. (A) Bioclastic wackestone containing bryozoans, brachiopods, echinoderms, ostracods, spicules, and recrystallized skeletons. Sample HRO 18. (B, C) Bioclastic wackestone-packstone composed of abundant echinoderm (crinoid) fragments, *Komia*, brachiopods, bryozoans, ostracods, trilobites, and other skeletons. Sample HRO 1. (D, E) Bioclastic wackestone composed of a diverse fossil assemblage including fragments of *Komia*, echinoderms, ostracods, brachiopods, and smaller foraminifers embedded in micrite. Sample HRO 4. (F) Crinoidal grainstone-rudstone, poorly washed, composed of abundant echinoderm (crinoid) fragments, subordinately of bryozoans, smaller foraminifers, and ostracods. Sample HRO 15. (G) Crinoidal rudstone containing abundant crinoid and bryozoan fragments, subordinately smaller foraminifers, brachiopods, and *Komia*. Sample HRO 17. (H) Floatstone containing large fragments of crinoids, brachiopods, and bryozoans floating in fine bioclastic matrix. Sample HRO 20.

their first cycle set in the Mockingbird Gap Hills and very high in that set in Rhodes Canyon, and in their Hembrillo Canyon section, the Atokan-Desmoinesian boundary is located in their second cycle set. Consequently, their only biostratigraphic datum is placed in two superposed cycle sets. Recognizing this, Algeo et al. (1991) suggested that the boundary at Hembrillo Canyon is likely lower because of facies restrictions, but they still show it in the second cycle set at Hembrillo Canyon. Furthermore, the very different stratigraphic levels of the Atokan-Desmoinesian boundary at two different sections in the first cycle set indicate very different sedimentation rates, stratigraphic condensation, and/or unconformities. This biostratigraphic discrepancy calls into question how many cycles are in the first cycle set in the different sections at Rhodes Canyon and Hembrillo Canyon.

Bachman (1968: pl. 2) presented two stratigraphic sections in the Mockingbird Gap Hills (his sections E and F), and the top of the Atokan in his sections is ~15 m above the base of the Gray Mesa Formation (his “lower member” of the Madera Formation), much higher than in the Algeo et al. (1991: fig. 4) section in the Mockingbird Gap Hills. At the type section of the “Lead Camp Limestone” (north side of San Andres Canyon, section 19, T18S, R24E), which is near Hembrillo Canyon, Bachman and Myers (1969) showed the lower 57 m of the Gray Mesa Formation as Atokan, so the base of the Desmoinesian here is much higher stratigraphically than is indicated by Algeo et al. (1991) in their Hembrillo Canyon section. Furthermore, Kottowski et al. (1956:36, pl. 1) indicated that the lower 70 m of the Pennsylvanian section at Hembrillo Canyon is Atokan (“Derryan”), also different from Algeo et al. (1991).

2. The basis for correlating the cycle boundaries in the sections of Algeo et al. (1991) is unclear. In the sections in the San Andres Mountains, most of the cycle boundaries are at the tops of relatively thick limestone intervals, but the thicknesses of these limestones vary, as does the lithology of the beds immediately overlying a cycle boundary at different locations. Indeed, without age control, it is extremely difficult to correlate cycle boundaries between stratigraphic sections 48 and

28 km apart (also see Wilson, 1989). For example, in the Mockingbird Gap Hills, mottling characterizes the limestone at the top of the first cycle set, but no such mottling is identified in the limestones at the top of the first cycle set in the Rhodes Canyon and Hembrillo Canyon sections. There is an algal horizon and mottling at the top of the second cycle set at Rhodes Canyon, but neither feature is present at the top of the second cycle set at Hembrillo Canyon or in the Mockingbird Gap Hills. In the Mockingbird Gap Hills, a covered interval overlain by a thin-bedded limestone is at the base of the third cycle set, whereas at Rhodes Canyon an andesite sill overlain by nodular limestone is at the base of that set, and a thin interval of bedded limestone overlain by a covered interval is at the base of the third cycle set at Hembrillo Canyon. These examples can be multiplied and raise concerns about the lithostratigraphic bases behind the correlations by Algeo et al. (1991) of most or all of their cycle-set boundaries.

3. Indeed, correlation of these cycle-set boundaries becomes even more problematic as the correlations are projected eastward across the Tularosa Basin into the Sacramento Mountains (Algeo et al., 1991: fig. 5). Thus, for example, the base of the first cycle set in Mule Canyon is a covered interval above cross-bedded limestone identified as “debris flow” deposits, totally different from the base of any of the second cycle sets in the San Andres Mountains.
4. Algeo et al. (1991) recognized only three kinds of limestone: massive/even bedded, nodular or argillaceous, and cross-bedded. Their sections are drawn at a scale of ~20 m to 1 cm (note, however, that the scale in their columns varies from 18 to 25 m/cm), so that “thin” units in their sections are at least 1 meter thick. As we have noted elsewhere (e.g., Lucas et al., 2016a), lithostratigraphy needs greater detail than this, or it will miss many facies changes essential to analyzing cycles.

In conclusion, the stratigraphic sections of Algeo et al. (1991: figs. 4, 5) show a wide range of disorganized facies stacking, mostly lack age control, and present few, if any, lithological marker beds on which to base correlations. Therefore, it is difficult to accept their cyclostratigraphic analysis.

FIGURE 22. (*Opposite*) Thin-section photographs of microfacies of limestones from the Gray Mesa Formation in the High Rolls section containing abundant fragments of *Komia*. Subordinately, other fossils are present such as bryozoans, spicules, brachiopods, echinoderms, smaller foraminifers, ostracods, and fusulinids: (A, C) wackestone, (B, E) wackestone-packstone, (D) packstone, (F, G) grainstone, and (H) rudstone. All photos under plane light. Width of photographs: A, B, E–H = 9 mm; C, D = 4.6 mm. A = sample HRO 2; B, C, D = sample HRO 8; E, F = sample HRO 7; G, H = sample HRO 9.

LaPeter (2007) measured the Gray Mesa Formation (his Bug Scuffle Member) section at High Rolls with the a priori idea that its deposition was driven by glacioeustatic cycles. Therefore, he expected to see many high-frequency, high-amplitude cycles with numerous subaerial exposure surfaces and large facies offsets in the Gray Mesa strata, which he considered to be cycles characteristic of icehouse deposition. To do this, LaPeter (2007) identified 12 facies in the Gray Mesa Formation and attributed relative water depths to these facies in order to use their stacking patterns to infer cyclical changes in water depth during their deposition. His analysis identified a lower composite sequence 1 and an upper composite sequence 2. However, he identified high-frequency cycles only in the upper part of composite sequence 2; in the rest of the Gray Mesa Formation low-amplitude cycles were identified, which LaPeter (2007) considered characteristic of a greenhouse climate. To explain this discrepancy, LaPeter suggested that much of the Gray Mesa Formation was deposited in water so deep that well-developed high-amplitude cycles (and any subaerial exposure of the carbonate platform) did not form. LaPeter (2007) equated the Gray Mesa Formation at High Rolls to the Desmoinesian Cherokee and Marmaton groups in the Midcontinent. Nevertheless, the Cherokee-Marmaton interval has more cycles than does the Gray Mesa Formation at High Rolls (cf. LaPeter, 2007: fig. 28; Heckel, 2013).

Clearly, depositional cycles are present in the Gray Mesa Formation. Subaerial exposure surfaces at the tops of shallowing upward cycles may be the best evidence of a glacioeustatic driver of deposition. However, there are too few cycles in the Gray Mesa Formation in the northern Sacramento Mountains to correspond to the number of cycles in the time-equivalent interval of Midcontinent cyclothems. Therefore, as with Gobbler deposition, that of the Gray Mesa Formation also likely was affected by local tectonic and climatic events. The small number of Gray Mesa cycles thus is most likely a result of unconformities in the section, reflecting variation in the availability of accommodation space created by tectonic events, so that some of the glacioeustatic cycles are simply not preserved. Certainly, a more detailed analysis of Gray Mesa Formation deposition is needed in the

northern Sacramento Mountains to establish the roles of diverse drivers of sedimentation.

Nevertheless, the Space History Member of the Gray Mesa Formation appears to represent a significant eustatic event(s), equivalent to the major highstand of the Swope cyclothem on the Midcontinent (Table 3). This event is also seen in the carbonate-dominated Amado Member within the clastic-dominated lower part of the Atrasado Formation in Socorro County, New Mexico, and to the north (e.g., Lucas et al., 2009, 2016b). We thus term this the “Amado event” and will discuss its regional significance in detail elsewhere.

BEEMAN FORMATION

LITHOSTRATIGRAPHY

Pray (1954, 1961) named the Beeman Formation for Beeman Canyon, which is just northeast of Alamogordo (Figure 2). Its type section at Long Ridge (Figure 7) is part of our Long Ridge section (Figure 24). Pray (1961) stated that the Beeman Formation is 107–152 m thick and is a slope-forming unit that consists mostly of thin-bedded, argillaceous limestone interbedded with calcareous shale (Table 2; see Figure 28). He noted that arkosic sandstone of the Beeman distinguishes it from quartzose sandstone of the underlying Gobbler Formation, though he did point out that a few sandstones in the Gobbler Formation are arkosic (see earlier discussion in Gobbler Formation Sedimentary Petrography section). Pray (1961) stated that the Beeman Formation represents an increase in local tectonism and noted that the distribution of its facies identified a “hinge line” that separates shallow water/shelfal deposits to the east from more basinal deposits to the west. Pray (1961) regarded the strata of the Beeman Formation in and near Beeman Canyon that we studied (Figures 24–28) as representative of its “basinal” facies.

We measured complete sections of the Beeman Formation at Space History B (Figure 12B,C), Long Ridge (Figure 24), and Space History C (Figure 25) and incomplete sections in La

FIGURE 23. (*Opposite*) Thin-section photographs of microfacies of limestones from the Gray Mesa Formation near the New Mexico Museum of Space History. All photos under plane light. Width of photographs: A–F = 9 mm; G, H = 4.6 m. (A) Phylloid algal floatstone composed of recrystallized fragments of phylloid algae that are oriented parallel to the bedding. The algal thalli are embedded in peloidal micrite. Sample SH 149. (B) Phylloid algal floatstone composed of fragments of recrystallized phylloid algae, which are encrusted by cyanobacteria and *Palaeonubecularia*. Sample SH 160. (C) Phylloid algal floatstone containing abundant recrystallized fragments of phylloid algae, subordinately other fossils, including echinoderms, ostracods, and smaller foraminifers. Sample SHC 7. (D) Phylloid algal floatstone containing also bryozoans, echinoderms, brachiopods, ostracods, small gastropods, trilobites, and smaller foraminifers. Sample SH 161. (E, F) Algal floatstone-rudstone containing large, broken fragments of mostly recrystallized thalli of *Archaeolithophyllum missouriense*. Sample SH 158. (G, H) Bioclastic wackestone-packstone composed of a diverse fossil assemblage including echinoderms, brachiopods, recrystallized phylloid? algae, smaller foraminifers, ostracods, bryozoans, echinoid radioles, and small gastropods. Samples SH 163 (G) and SH 167 (H).

FIGURE 24. Long Ridge section. See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "TG 31").

Space History C

FIGURE 25. Space History C section. See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "SHC 112").

FIGURE 26. La Luz Canyon sections. See Table 1 and Figure 2 for location of sections. See Figure 8 for legend to lithologic symbols.

FIGURE 27. Beeman Canyon A and B sections. See Table 1 and Figure 2 for location of sections. See Figure 8 for legend to lithologic symbols.

FIGURE 28. Overview photograph of the Beeman Formation at the Space History B and C sections.

Luz Canyon (Figure 26) and at Beeman Canyon B (Figure 27). In general, the Beeman Formation is composed of alternating shale, siltstone, sandstone, and limestone. Within the Beeman Formation, distinct lateral facies changes are observed from east to west (Pray, 1961; Raatz, 1996; Raatz and Simo, 1998). Sandstones of the Beeman Formation contain higher amounts of detrital feldspars (subarkose-arkose) compared to the quartz-rich sandstones of the Gobbler Formation.

The Beeman Formation (Figure 4, Table 2) can be divided lithologically into a lower interval that is dominantly siliciclastic (here named the Indian Wells Canyon Member) and an upper interval that is composed of alternating shale and limestone (here named the Horse Ridge Member). The total thickness of the Beeman Formation ranges from 163 m at the Long Ridge section (the type section; Figure 24) to at least 168 m at La Luz Canyon (Figure 26), 172 m at the Space History C section

(Figure 25), and 247 m at the Space History B section (Figure 12B-C). The Indian Wells Canyon Member is 64 m thick at Space History C, 73 m at Long Ridge, 88 m at Space History B (type section), and 125 m at La Luz B. The thickness of the Horse Ridge Member measures 43 m at La Luz A, 90 m at Long Ridge, 108 m at Space History C, and 159 m at Space History B (type section).

At the type section, the Indian Wells Canyon Member is mostly composed of shale/siltstone with intercalated sandstone, rare conglomerate, and two thin limestone beds. Some lateral facies changes occur. At Space History C a thick conglomerate is exposed in the middle of this member. At Long Ridge and La Luz B intercalated limestone beds are more abundant.

The Horse Ridge Member at the type section (Space History B) is composed of alternating shale (constituting 60% of the section) and limestone (39%), one thin sandstone bed, and one thin conglomerate. Limestone includes thin limestone beds, thicker, even-bedded limestone units, and wavy to nodular limestone units. At Long Ridge a few dolomitic beds are intercalated. Also, at Space History C dolomitic beds are common in the Horse Ridge Member.

Members of the Beeman Formation

In the sections of the Beeman Formation we measured in the Long Ridge, Space History, and La Luz Canyon areas (Figures 22–24), the formation can be seen to consist of a lower interval of sandstone and shale with a few beds of limestone and an upper interval of shale and limestone beds, with very few sandstones (also see Pray, 1961; Wilson, 1972b; Raatz and Simo, 1998: fig. 3). These are units worthy of member-level distinction and can best be seen in our Space History B (Figure 12B-C), Long Ridge (Figure 24), and Space History C (Figure 25) sections of the Beeman Formation. In the Beeman Formation type section at Long Ridge A (Figure 24), units 1–56 are the lower member (73 m thick), whereas the upper member at Long Ridge B is units 1–112, 90 m thick. Therefore, we recognize and name two formal members of the Beeman Formation in the Sacramento Mountains, the Indian Wells Canyon Member and overlying Horse Ridge Member (Figure 4).

We name the lower member of the Beeman Formation the Indian Wells Canyon Member for the canyon of that name located in section 11, T16S, R11E. The type section of the Indian Wells Canyon Member of the Beeman Formation is units 171–210 of our Space History B section (Figure 12B-C). The type section is 88 m thick and composed of shale, siltstone, sandstone, and conglomerate with only two thin limestone beds intercalated. Covered intervals, which probably represent shale and siltstone units, are up to 14 m thick.

Shale, silty shale, and, rarely, siltstone are mostly greenish and gray and are present as units up to 12 m thick. At the base of the member, black, calcareous shale contains ostracods. The shale of unit 185 (Figure 12B), which overlies a thin

limestone bed with brachiopods, contains fossil plants, ostracods, conchostracans, and fish fragments.

The sandstone units are 0.2–2.6 m thick and include horizontally laminated, cross-bedded, hummocky cross-bedded, and ripple-laminated sandstone. The thicker sandstone succession (Figure 12C, units 192–197; 8 m thick) is composed of cross-bedded, rare horizontally laminated sandstone, pebbly sandstone, and conglomerate, which are also cross-bedded. Conglomerate is poorly sorted and contains reworked limestone clasts (probably reworked calcrete clasts), quartz, and chert clasts with diameters up to 8 cm. The lower limestone bed (Figure 12B, unit 184) is 0.2 m thick and contains brachiopods. The upper limestone bed (Figure 12C, unit 205) is 0.4 m thick and contains calcareous algae, bryozoans, brachiopods, and gastropods.

We name the upper member of the Beeman Formation the Horse Ridge Member, after the ridge of that name in sections 2, 3, and 11, T16S, R11E (Figure 2). The type section of the Horse Ridge Member is units 211–331 of our Space History B section (Figure 12C). The type section is 159 m thick and composed of alternating shale (60% of the section) and limestone (39% of the section) and one sandstone bed. Covered intervals (likely shale) below are up to 7 m thick, and the uppermost 52 m of the Beeman Formation at Space History B are covered (probably mostly shale/siltstone) and contain thin limestone beds (see the Space History C section, Figure 25).

Shale/siltstone units in the Horse Ridge Member at the Space History B section are up to 5.4 m thick, mostly 0.5–2 m thick. Shale/siltstone is greenish and gray and subordinately black. The greenish shale of unit 243 (4.2 m thick) contains abundant limestone nodules. Limestone nodules are also present in the thin (0.4 m) shale unit 248. Intercalated limestone units are 0.1–1.9 m thick and include the following lithotypes:

1. Individual limestone beds (mostly 0.1–0.4 m thick, rarely up to 0.7 m thick) with even bedding planes, composed of gray lime mudstone (bioclastic mudstone to wackestone, rarely floatstone), locally contain fossils such as calcareous algae, crinoid fragments, ostracods, and bryozoans. The limestone bed of unit 316 contains mud cracks.
2. Thicker limestone units, thin to medium bedded with even bedding planes, are mostly composed of lime mudstone, rarely of grainstone. Fossils locally present are algae, bryozoans, and gastropods. A few limestone beds display ripple lamination, and some limestone beds probably represent tidalites. Near the top of the Horse Ridge Member, cross-bedded grainstone is present.
3. Wavy to nodular limestone is partly bioturbated and composed of lime mudstone, locally containing algae and bryozoans.
4. On top of limestone unit 277 a thin conglomerate (10 cm thick) is exposed that contains intraclasts with diameters up to 4 cm.
5. In a few limestone units, some chert nodules are present.

In the Horse Ridge Member at the Space History B section, only one thin (0.5 m) cross-bedded sandstone is intercalated (unit 304). However, beginning in about the middle of the member, most of the overlying limestone beds are dolomicrite, so if these limestones are separated from other limestones, dolomicrite is 17% of the section, and the remaining limestones are 22%.

Type Section of the Beeman Formation: Long Ridge

The type section of the Beeman Formation (Figure 24) is located at the northwestern end of Long Ridge (Figure 2). We measured a section in two segments (A and B) through the Beeman Formation at Long Ridge, where it rests on thick-bedded limestone (wackestone) of the Gray Mesa Formation. Total Beeman thickness is 177 m. The lower 70 m of the Beeman Formation (Indian Wells Canyon Member) consist of shale intervals (many covered) and sandstone and a few intercalated thin limestone beds. Covered intervals are up to 14.5 m thick, and shale units are up to 6.3 m thick. Shale is brownish and greenish. Most shale intervals are less than 1 m thick. Approximately 60 m above the base of the section, 1.7 m of gypsum and gypsiferous siltstone, interbedded with thin sandstone beds, are exposed.

Sandstone crops out as thin beds (a few centimeters to 0.6 m thick) that are greenish, fine grained, massive, and subordinately horizontally laminated. Thicker sandstone units (up to 9.1 m thick) are composed of the following lithotypes: coarse-grained, cross-bedded pebbly sandstone with clasts up to 3 cm in size; coarse- to fine-grained cross-bedded sandstone; fine-grained, horizontally laminated sandstone; and massive, fine-grained sandstone. All sandstone lithotypes are greenish and, rarely, brownish, and locally, soft-sediment deformation structures are present.

Intercalated limestone beds are 0.1–0.8 m thick and partly display wavy bedding planes. One limestone bed contains a few chert nodules. In Long Ridge section A (Figure 24), all limestone beds are composed of gray to dark gray lime mudstone (mostly wackestone). Limestone beds of units 29 and 30 contain abundant brachiopods. Fossil wood was observed within the thick covered interval (unit 80) about 90 m above the base of the Beeman Formation.

The upper part of the type section of the Beeman Formation (Horse Ridge Member) is about 105 m thick (Figure 24, section B) and composed of alternating shale/covered intervals and limestone, rare dolomite, sandstone, and conglomerate. Shale is greenish, gray, yellowish, and, rarely, reddish (uppermost part) and partly silty (siltstone). A few shale units contain limestone nodules.

Intercalated limestone beds and limestone units are 0.1–1.6 m thick and include the following lithotypes:

1. Thin- to medium-bedded limestone with even bedding has mostly a muddy texture (bioclastic mudstone to wackestone, rare crinoidal wackestone). Rarely, brachiopods are present. One limestone bed (Figure 24, section B, unit 71) contains abundant intraclasts.

2. Less common is nodular limestone, and wavy-bedded limestone is very rare. Wavy and nodular limestone both have a muddy texture and are partly bioturbated.
3. Rare dolomite intervals are intercalated and are 0.1–1 m thick and composed of dolomicrite.
4. Approximately 23 m above the base, a breccia bed (Figure 24, section B, unit 34; 1 m thick) is composed of angular to subangular limestone clasts up to 10 cm in diameter.
5. A thin conglomerate bed (0.2 m) is intercalated 43.5 m above the base.

Greenish siltstone and fine-grained sandstone with poorly developed horizontal lamination is exposed in the upper part, underlain by a covered (shale) interval and overlain by reddish shale. Intercalated in this siltstone/fine-grained sandstone are two Cenozoic sills that are 2 and 3.2 m thick (units 94 and 98). The Beeman-Holder contact is chosen at the base of a relatively thick limestone interval above slope-forming shale at the top of the Beeman Formation.

Space History C Section

At the Space History C section (Figure 25), the Beeman Formation is 172 m thick and can be divided into a lower part (Indian Wells Canyon Member), which is 64 m thick and dominantly siliciclastic (14–78 m; Figure 25, units 14–43), and an upper part (Horse Ridge Member), which is 108 m thick and composed of alternating shale, limestone, and dolomite (78–186 m; Figure 25, units 44–173 m). The Beeman Formation base is a shale unit on top of a massive algal limestone with silicified fossils of the Gray Mesa Formation, capped by a sub-aerial exposure surface. The topmost shale unit of the Beeman Formation is overlain by algal limestone of the Holder Formation (Figure 25).

At the Space History C section (Figure 25), the Indian Wells Canyon Member (lower 64 m) of the Beeman Formation is composed of the following lithotypes:

1. Green and gray shale, 0.3–9.3 m thick, contains bivalves in unit 22 and gastropods in unit 23.
2. Sandstone appears as individual sandstone beds and thicker sandstone units (0.1–2.8 m thick), including massive, horizontally laminated, ripple-laminated, and trough-cross-bedded sandstone. Soft sediment deformation structures are present in unit 42, and oxidized plant stems (ferruginous, lacking in organic matter) are present in unit 15.
3. Conglomerate is intercalated in the middle of the lower unit as one thin conglomerate bed (0.3 m) and one thicker, cross-bedded limestone-pebble conglomerate containing fossil logs. The conglomerate has an erosional base and a channel-like geometry with a thickness ranging from 1 to 4 m.

4. Limestone is rare and occurs as one stromatolitic limestone bed (unit 35, 0.4 m thick) and an algal limestone containing brachiopods (unit 40, 1 m thick), both intercalated in shale in the upper half of the Indian Wells Canyon Member.

At the Space History C section, the Horse Ridge Member is composed of shale/covered units, intercalated limestone, dolomite, and rare thin sandstone (Figure 25):

1. Shale/covered units are 0.1–11.2 m thick. Shale of unit 66 contains limestone nodules, and shale of unit 79 contains dolomitic nodules. Shale of units 126 and 128 is gypsiferous.
2. Limestone occurs as single limestone beds or thicker bedded limestone units, and thickness ranges from 0.1 to 1.6 m. Thicker limestone units are thin to medium bedded with even bedding planes. Wavy bedding is very rare. Limestone is composed of lime mudstone, which locally contains abundant algae, crinoid fragments, and bryozoans (bioclastic mudstone and wackestone, rarely floatstone). Rarely, limestone composed of grainstone is present, particularly in the upper part of the member where some of the grainstone beds display cross bedding. A few chert nodules are present in the limestone of unit 65.
3. Like limestone, dolomitic limestone and dolomite (dolomiticrite) are intercalated as thin beds and thicker units (up to 1.9 m thick) that are thin to medium bedded. Bedding is mostly even, and wavy bedding is rarely observed. Fossils are rare and include brachiopods and bryozoans.
4. A thin, dolomitic intraclast breccia (0.6 m) is intercalated about 35 m above the base of the member.
5. Sandstone occurs as thin sandstone beds (5–10 cm thick) in unit 53 in the lower part and as a thin (0.2 m thick), ripple-laminated sandstone bed in the uppermost part of the section.

High Rolls Section

At the High Rolls section (Figure 20, units 128–130), the lowermost part of the Beeman Formation is poorly exposed, mostly covered greenish shale. The first indurated beds are bedded limestone and crinoidal limestone, suggesting that the facies of the lower Beeman at High Rolls may be more marine than it is to the west and southwest.

La Luz A Section

At the La Luz A section (Figure 26), the upper part (48 m) of the Beeman Formation (Horse Ridge Member) and lower part of the Holder Formation (53 m) are exposed (Figure 26). The succession is composed of alternating shale/covered units and limestone, subordinately of sandstone and conglomerate.

In the upper Beeman Formation, shale/covered units are mostly 0.5–3 m thick and rarely up to 8 m thick. Sandstone and conglomerate occur near the base (between 5 and 7 m) of the section. Sandstone is mostly horizontally laminated and rarely cross-bedded. The conglomerate in the lower part is about 1 m thick and intercalated in sandstone.

La Luz B Section

At the La Luz B section (Figure 26), 125 m of the Beeman Formation are exposed and composed of shale/covered units and intercalated sandstone, rare conglomerate, and limestone. This section is readily correlated to the La Luz A section by matching the very distinctive interval of sandstone and conglomerate in both sections (La Luz B, unit 51, and La Luz A, units 2–3). This means the two sections encompass a nearly complete section of the Beeman Formation that is ~166 m thick.

Shale/covered units are up to 10 m thick. Sandstone units are thin, massive to horizontally laminated sandstone beds (5–50 cm thick) and are thicker sandstone units that are 1–5 m thick and cross-bedded. Sandstone of unit 11 is a pebbly sandstone, poorly sorted, displaying cross bedding and containing rounded clasts up to 5 cm in diameter, mostly various types of carbonate clasts and chert clasts. Unit 13 is a cross-bedded sandstone, with hummocky cross bedding in the upper part and locally displaying soft sediment deformation structures. The sandstone contains some pebble layers with clasts up to 5 cm in diameter. The sandstone of unit 43 is cross-bedded.

The conglomerate of unit 5 rests on cross-bedded sandstone and is up to 3 m thick. This conglomerate is partly matrix supported, partly clast supported, composed of sandy matrix and rounded clasts up to 15 cm in diameter. The dominant clast type is quartz; subordinately, limestone and chert pebbles are present.

Limestone in the section is single limestone beds and thin-bedded limestone units. Limestone of unit 9 is massive to nodular, cherty, with a muddy texture and contains phylloid algae. Limestone of unit 24 is composed of two beds of gray lime mudstone, the lower bed having a few larger fossils (myalinid bivalves) floating in the muddy matrix. The limestone bed above the myalinid bed is 50 cm thick and contains large gastropods.

La Luz C Section

At the La Luz C section, ~26 m of the lower Beeman Formation are exposed and are composed of shale intervals, conglomerate, pebbly sandstone, fine-grained sandstone, and limestone (Figure 26). The shale is greenish and very thinly laminated. The sandstone is greenish, cross-bedded (up to 1.5 m thick) and horizontally laminated, and partly pebbly (10–50 cm thick). Pebbly sandstone to fine-grained conglomerate (up to 40 cm thick) is horizontally laminated and cross-bedded with clasts up to 3–4 cm in diameter; it is polymict, poorly sorted, containing abundant limestone clasts and, subordinately, reddish chert

clasts. The conglomerate of unit 11 contains fossils (crinoid fragments, corals).

Beeman Canyon B Section

In the Beeman Canyon B section (Figure 27), the most extensively collected plant fossil locality is in shale (unit 13) of the upper part of the lower, dominantly siliciclastic Indian Wells Canyon Member of the Beeman Formation. The Beeman Canyon B section has an exposed thickness of 22 m (Figure 27).

The lower 17 m of the section is composed of shale-siltstone and intercalated sandstone and rare limestone. Intercalated sandstone beds and units are up to 0.9 m thick, dominantly ripple laminated, rarely cross-bedded, and massive. In the lower part, thin limestone beds (0.1–0.6 m thick) are intercalated. The thickest limestone bed contains brachiopods. The plant fossils occur in shale on top of a thin limestone bed. This plant-bearing shale (unit 13) correlates with unit 206 at the Space History B section (Figure 12A).

The upper part of this section represents the lowermost part of the upper Beeman Formation (Horse Ridge Member), which is composed of alternating shale, limestone, and dolomitic limestone. The lowermost three limestone beds are nodular and contain brachiopods. Overlying carbonate beds are dolomitic or muddy limestone.

SEDIMENTARY PETROGRAPHY

We undertook petrographic study of sandstone and limestone beds of the Beeman Formation at our Space History B and C sections and at our Beeman Canyon B section (Figures 29–31). Sandstones of the Space History B and C and Beeman Canyon B sections are identical to each other with regard to textural and compositional features. Sandstone is moderately sorted and rarely well sorted, and the detrital grains are dominantly angular

to subangular and subordinately subangular to subrounded. Quartz is the most abundant grain type, represented mainly by monocrystalline quartz and subordinately by polycrystalline quartz (including a few stretched metamorphic quartz grains). Locally, some chert grains are present.

Detrital feldspars are common to abundant and represented by plagioclase and potassium feldspars. Many of the feldspar grains are altered to some degree and partly replaced by calcite cement. Rock fragments are present in small amounts and include granitic rock fragments, fine-grained (phyllitic) rock fragments composed of quartz and micas, and a few volcanic rock fragments, which are mostly altered; some of the rock fragments are of basaltic composition, composed of plagioclase laths. Locally, sedimentary rock fragments are present, such as reworked siltstone, fine-grained sandstone grains, and micritic carbonate grains. Some of the sandstones contain many chlorite grains (altered grains). Micas are rare and dominantly muscovite, rarely biotite. Accessory grains are opaque minerals, titanite, rutile, tourmaline, zircon, and apatite. Fossil fragments, such as those of echinoderms, fusulinids, brachiopods, and smaller foraminifers, are very rare. Authigenic overgrowths on detrital quartz grains also are rare. The detrital grains are cemented by coarse, poikilotopic calcite. Locally, small amounts of matrix are present (Figures 29B–H, 30A–C, E–H).

Mixed siliciclastic-carbonate sandstone (Space History C section) is moderately to well sorted. Besides mono- and polycrystalline quartz, it contains detrital feldspars, rare rock fragments, and micas. These sandstones also contain many micritic carbonate grains (sedimentary rock fragments) and some fossils, including echinoderms, fusulinids, smaller foraminifers, shell fragments, and recrystallized skeletons (Figures 29A, 30D). Sandstones of the Beeman Formation show a wide range in composition, and in the QFL diagram they plot into the field of sublitharenite and subarkose and rarely into the field of arkose and lithic arenite (Figure 18).

FIGURE 29. (*Opposite*) Thin-section photographs of sandstones from the Beeman Formation near the New Mexico Museum of Space History. All photos under polarized light. Width of photographs: A, G = 3.2 mm; B, C, E, F, H = 1.2 mm; D = 6.3 mm. (A) Mixed siliciclastic-carbonate sandstone composed of quartz, feldspars (partly replaced by calcite), metamorphic rock fragments, few altered (chloritic) grains, and calcite cement. Sample SHB 202. (B) Sandstone composed of quartz, feldspars (partly replaced by calcite), altered (chloritic) rock fragments, and calcite cement. Sample SHC 15. (C) Sandstone composed of quartz, feldspars (partly replaced by calcite), rare chert grains, few altered (chloritic) grains, matrix, and calcite cement. Sample SHC 19. (D) Poorly sorted pebbly sandstone composed of quartz, feldspars (slightly altered), sedimentary rock fragments (rarely containing fossils such as fusulinids), metamorphic rock fragments, rare fossils, matrix, and calcite cement. Sample SHC 26. (E) Sandstone containing quartz, feldspars, rock fragments, and coarse, blocky calcite cement. Sample SHC 32. (F) Fine-grained sandstone containing quartz, feldspars (slightly altered, partly replaced by calcite cement), altered (chloritic) grains, and rare phyllitic metamorphic grains. Detrital grains are cemented by quartz (authigenic overgrowths on detrital quartz grains) and blocky calcite cement. Sample SHC 37. (G) Fine-grained sandstone composed of quartz, feldspars (slightly altered, mostly potassium feldspars), rock fragments including altered (chloritic) grains, and rare micas. Detrital grains are cemented by quartz (authigenic overgrowths) and small amounts of calcite cement. Sample SHC 42. (H) Fine-grained sandstone composed of quartz, feldspars, micritic carbonate grains (sedimentary rock fragments), abundant micas (mostly biotite), and blocky calcite cement. Sample SHC 159.

Limestone of the Beeman Formation at the Space History B and C and Beeman Canyon sections is composed of different microfacies types. They are peloidal mudstone, bioclastic mudstone, and different types of wackestones and grainstone: peloidal wackestone, peloidal-intraclast wackestone, ostracodal wackestone (Figure 31C), foraminiferal wackestone (Figure 31B), rare spiculite, and bioclastic wackestone to floatstone (Figure 31A,H) and grainstone (peloidal grainstone to packstone, rare rudstone, and floatstone). Muddy textures dominate, and high-energy microfacies such as grainstone are rare.

Peloidal mudstone, bioclastic mudstone, peloidal wackestone (Figure 31D), peloid-intraclast wackestone (Figure 31E), peloidal wackestone, and peloidal grainstone/packstone (Figure 31F) contain a low-diversity fossil assemblage (mainly ostracods), and assemblages of all other microfossil types are characterized by a moderate to high diversity. All microfacies contain unidentifiable recrystallized skeletons. One grainstone type contains abundant fecal pellets (Figure 31G), some micritic intraclasts, many detrital quartz grains, a few detrital feldspars, and rare fossil fragments. Limestones of the Beeman Formation contain the same fossil groups as observed in the underlying Gobbler Formation (echinoderm fragments, bryozoans, brachiopods, bivalves, gastropods, smaller foraminifers, rare fusulinids, sponges, sponge spicules, ostracods, trilobites, brachiopod spines, echinoid radioles). However, echinoderms, bryozoans, and brachiopods are less abundant than in the Gobbler Formation.

PALEONTOLOGY AND AGE

The Beeman Formation is extremely fossiliferous, but very little has been published about its paleontology. The plant fossil assemblages from the Beeman discussed below are from green to dark gray shale beds that also yield fossils of ostracods, conchostracans, xiphosurans, insects, bivalves, microconchids, and fishes. We made no extensive effort to collect and study these

animal fossils, but we do present some preliminary observations below, pending more comprehensive collection and study.

Hunt et al. (2012) and Hunt and Lucas (2017) described an assemblage of coprolites from a shale bed of the Horse Ridge Member of the Beeman Formation in Dry Canyon (NMMNH locality 3276). These are spiral coprolites of chondrichthyan and/or sarcopterygian fishes in a marine shale. Hunt et al. (2012) identified the ichnogenera *Bibliocoprurus*, *Heteropolacoprurus*, *Hyronocoprurus*, *Kalocoprurus*, and *Liassocoprurus*. Hunt and Lucas (2017) regarded the Beeman coprolites as a typical assemblage of normal-marine Pennsylvanian coprolites.

Raatz (1996) and Raatz and Simo (1998) reported a few Missourian taxa of conodonts and fusulinids from the Beeman Formation. Wahlman and Barrick (2018) documented fusulinids that indicate the Beeman is early middle Missourian to early middle Virgilian in age. They reported some of the conodont data documented here.

We extracted calcareous microfossils and conodonts from the Beeman Formation that are documented below. They indicate that the formation ranges in age from early Missourian to early Virgilian (see Table 3 and Age and Correlation later in the Discussion section below).

DEPOSITIONAL SYSTEMS

Pray (1961) regarded Beeman deposition as reflecting an increase in tectonic activity. Closer to the Pedernal uplift (east; Figure 3), a shallow shelf facies is developed that grades into a deeper shelf facies toward the west (Pray, 1961). The Beeman Formation encompasses immature clastic sediments of a fluvio-deltaic environment (lower, Indian Wells Canyon Member), overlain by partly cyclic carbonate shelf deposits (upper, Horse Ridge Member). According to Pray (1961), Raatz (1996, 2002), and Raatz and Simo (1998), the Beeman Formation reflects a modest pulse of tectonic uplift together with rising sea level under a seasonally dry climate regime.

FIGURE 30. (*Opposite*) Thin-section photographs of sandstones from the Beeman Formation at the Beeman Canyon (A–D) and Space History (E–H) sections. All photos under polarized light. Width of photographs: A–F = 1.2 mm; G, H = 3.2 mm. (A) Siltstone to fine-grained sandstone composed of quartz, feldspars (slightly altered), carbonate grains (sedimentary rock fragments), micas, and a few altered grains. Sample BCB 2. (B) Fine-grained sandstone containing detrital quartz grains, feldspars, a few carbonate grains, and rare micas, cemented by coarse blocky calcite. Sample BCB 4. (C) Fine-grained sandstone composed of detrital quartz grains, feldspars (slightly altered), rock fragments composed of quartz and feldspar, matrix, and calcite cement, which partly replaces detrital feldspars. Sample BCB 23. (D) Fine-grained, mixed siliciclastic-carbonate sandstone composed of quartz, feldspar (partly replaced by calcite), carbonate grains, micas, and rare fossils (echinoderms, ostracods), cemented by calcite. Sample BCB 28. (E) Sandstone containing detrital quartz grains, many feldspars (partly replaced by calcite, e.g., grain in the center), rare rock fragments, and coarse blocky calcite cement. Sample SHB 173. (F) Fine-grained sandstone containing detrital quartz grains, many detrital feldspar grains that are slightly altered, some rock fragments, and rare micas. Some of the detrital quartz grains display authigenic overgrowths. Sample SHB 179. (G) Sandstone composed of detrital quartz grains, many feldspars (mostly potassium feldspars, slightly altered), rock fragments of quartz and feldspars, carbonate grains (sedimentary rock fragments), and rare micas. Sandstone is cemented by quartz, which occurs as authigenic overgrowths on detrital quartz grains. Sample SHB 192. (H) Sandstone containing quartz grains, some feldspars, and rock fragments (including volcanic rock fragments [grain in the center] and granitic rock fragments). The sandstone is cemented by coarse blocky calcite. Sample SHB 193.

Raatz et al. (1994) divided the uppermost Gobbler Formation and the Beeman Formation into three lithofacies associations (LAs): (1) sandstone-dominated Horse Ridge LA composed of green, micaceous sandstone and conglomerate interbedded with shale and siltstone interpreted as nearshore high- to low-energy deposits; (2) open-marine, carbonate-dominated Indian Wells LA composed of black and calcareous shales interbedded with argillaceous and open-marine carbonates deposited in environments ranging from deeper shelf to open, normal marine, and restricted shallow lagoon; and (3) restricted-marine, carbonate-dominated Dry Canyon LA composed of gray calcareous shale interbedded with low-diversity, fine-grained, thin, even-bedded carbonates, indicating deposition in a dominantly restricted shallow marine setting.

The uppermost Gobbler and Beeman Formations according to Raatz et al. (1994) are composed of thin transgressive and thick regressive deposits that make up most of the succession. Raatz and Simo (1998) studied in detail the Beeman Formation in the Dry Canyon area. There, they identified the deposits of the Beeman Formation as the most basinward deposits of that formation in the Sacramento Mountains. Thus, Raatz and Simo stated that most of the Beeman Formation in Dry Canyon is composed of highly restricted shallow-water to peritidal deposits with evidence of high-energy rip-ups, winnowed high-energy fabrics, peritidal lamination, a restricted low-diversity fossil assemblage, and subaerial exposure. The basinal (distal) nature is recorded by thin transgression and maximum flooding deposits of diastemic dark shales, skeletal wackestones, and some of the carbonate mudstones. In contrast to this basinal facies, the shelf facies of the Beeman Formation is dominated by open-marine limestone. Raatz and Simo (1998) termed this paradoxical relationship of the basinal facies, which is dominated by shallow-water restricted deposits, and the shelf facies, which is dominated by open-marine limestone, “inverse sedimentation.”

Inverse sedimentation refers to the interpretation of sedimentation by Raatz and Simo (1998) during what they referred to as

supersequence set 2C (Figure 32). During deposition of that stratigraphic interval, Raatz and Simo (1998) inferred that the basinal deposits consist of a relatively high volume of shallow, restricted marine deposits relative to the shelfal deposits formed to the east. They thus argued that earlier workers had misinterpreted the basinal deposits as deepwater sediments, largely on the basis of their deposition having taken place basinward of shelf deposition.

However, inverse sedimentation strikes us as an ad hoc interpretation based on a misconception that began with Pray (1961). Thus, the Missourian basinal deposits of the Orogrande Basin are in the lower part of the Panther Seep Formation in the San Andres Mountains. These are dominantly shallow marine deposits, mostly calcarenites and shallow marine limestones that indicate rapid sedimentation filling the subsiding Orogrande Basin, so that no truly deep marine, basinal facies formed during the Missourian (Kottlowski et al., 1956; Schoderbek, 1994). Thus, the proposition that an east-to-west transect through the Beeman Formation should be a transect into deeper-water facies, which underlies the idea of “inverse sedimentation,” should be reconsidered.

The lower part of the Beeman Formation is dominantly siliciclastic (Indian Wells Canyon Member) and composed of mudstone-shale with intercalated sandstone, mostly cross-bedded, subordinately horizontally laminated, ripple laminated, and rarely massive. Rarely, hummocky cross bedding is observed. Fossils in these sandstones are rare or absent, but the shales yield abundant fossils of plants and some fossils of ostracods, conchostracan, insects, and other animals that indicate dominantly freshwater deposition (see below). This succession was interpreted as fluviodeltaic sediments by Raatz and Simo (1998) and Raatz (2002), an interpretation that we also advocate.

At the Space History C section, the plant-bearing horizon that occurs 2 m above the base of the Beeman Formation is composed of greenish-gray siltstone with intercalated micritic gray limestone. This horizon lacks animal fossils and is overlain by gypsiferous sediment. We infer that the plant-bearing horizon was deposited in a restricted, probably lagoonal setting.

FIGURE 31. (*Opposite*) Thin-section photographs of different microfacies types of limestones from the Beeman Formation near the New Mexico Museum of Space History. All photos under plane light. Width of photographs: A–E, G = 4.6 mm; F = 1.8 mm; H = 9 mm. (A) Bioclastic wackestone composed of recrystallized skeletons, brachiopod fragments including spines, ostracods, echinoderms, a few foraminifers, and rare bryozoans. Sample SHB 215. (B) Foraminiferal wackestone containing abundant smaller foraminifers (mostly *Palaeonubecularia*), recrystallized skeletons, brachiopod fragments, echinoderms, and other species of smaller foraminifers. Sample SHC 163. (C) Ostracodal wackestone, bioturbated, containing abundant ostracod shells that are commonly oriented parallel to the bedding. Most of the ostracod shells are fragmented; rarely, both valves are preserved. Sample SHB 262. (D) Peloidal wackestone containing a few fossils such as ostracods, recrystallized skeletons, spicules, and smaller foraminifers. Sample SHB 217. (E) Peloid-intraclast wackestone to packstone, locally grainstone, containing abundant peloids and micritic intraclasts, a few ooids, and few fossil fragments. Sample SHB 269. (F) Peloidal grainstone containing abundant peloids, some micritic intraclasts, and rare ooids. Fossil fragments are rare. Sample SHC 112. (G) Moderately sorted grainstone composed of abundant fecal pellets, rare ooids, peloids, intraclasts, and a few fossils (mostly ostracods). Sample SHB 225. (H) Large fossils, mostly gastropods, float in peloidal wackestone to packstone. Other fossils present are smaller foraminifers and ostracods. Sample SHB 232.

Approximately 23 m above the base of the Beeman Formation at the Space History C section (Figure 25), allochthonous plant fossils occur in a 50 cm thick interval composed of brown siltstone, dark shale, and micritic mudstone. Besides plant fossils this horizon contains ostracods, conchostracans, fish scales, bivalves, and insect wings, indicating freshwater conditions. The presence of conchostracans suggests periodic drying of parts of the aquatic environment.

The different lithofacies of the Beeman Formation are partly arranged to form coarsening (shallowing) upward sequences, starting with shales that we interpret as offshore-distal prodelta deposits, overlain by alternating shale and thin sandstone beds interpreted as proximal prodelta deposits and thicker, mostly cross-bedded sandstones interpreted as distributary mouth bar deposits (e.g., Space History B section, Figure 12B, units 172–174, 182–195). Locally, fine-grained delta plain deposits are developed on top. Two thin, intercalated limestone beds composed of foraminiferal wackestone and bioclastic wackestone to floatstone document short periods of shallow, restricted to normal-marine conditions. At the Space History C section, the Indian Wells Canyon Member is thinner (63 m) compared to its thickness in the Space History B section (98 m), and a 4 m thick conglomerate is exposed in the middle of the member, representing an incised channel fill, likely a very local facies.

At the Space History B section, the Indian Wells Canyon Member is overlain by alternating shale and limestone of the Horse Ridge Member. Most of the intercalated limestones have a muddy texture with peloidal mudstone and wackestone and ostracodal wackestone being typical microfacies. These microfacies are characterized by a low-diversity fossil assemblage, indicating deposition in a shallow, low-energy, restricted marine environment (shallow subtidal). Another common microfacies is bioclastic wackestone with a moderate- to high-diversity fossil assemblage pointing to more open-marine, low-energy conditions. Individual limestone beds composed of partly cross-bedded grainstone indicate deposition in a high-energy shallow marine setting. At the Space History C section, the facies are similar, although in that section many muddy limestone beds are dolomitized (dolomicrites).

However, we are not able to distinguish between a succession composed of an open-marine, carbonate-dominated (Indian Wells) lithofacies assemblage and a restricted-marine, carbonate-dominated (Dry Canyon) lithofacies assemblage as described by Raatz et al. (1994). We also did not observe (see below) transgressive-regressive cycles as did Raatz et al. (1994) and Raatz (2002). Shale and restricted- and open-marine limestones alternate in a noncyclic pattern at both the Space History B and C sections. Indeed, the Space History B and C sections were probably deposited on a segment of the narrow Sacramento shelf on which synsedimentary tectonics was more pronounced and prevented the formation of well-developed glacioeustatic cycles, contrary to the conclusions of Raatz et al. (1994) and Raatz (2002).

Sequence Stratigraphy

Raatz (1996; see also Raatz and Simo, 1998) presented a sequence stratigraphic analysis of the upper part of the Beeman Formation based primarily on outcrop stratigraphic sections and two cores along an east–west transect through Dry Canyon (essentially along the path of U.S. Highway 82; Figure 2). Along this transect, Raatz identified seven sequence sets in the upper Beeman and argued that they record glacioeustatically driven transgressions and regressions that can be directly correlated to Midcontinent cyclothems. We are skeptical of these conclusions because of the following aspects of Raatz's analysis that do not support identifying glacioeustasy (an allogenic driver) as the primary driver of upper Beeman deposition: (1) the lack of lateral continuity of facies, (2) the lack of regular facies stacking patterns that support the cycles identified, (3) the lack of clear sequence boundaries, and (4) some problems with facies interpretations and lateral correlations.

To exemplify these concerns, we redrew the Dry Canyon West outcrop section (Figure 32) of Raatz (1996, appendix; Raatz and Simo, 1998: fig. 8) to compare it to our nearby Space History C section (Figure 25). These sections are approximately 2 km apart north–south, and the fact that they are north–south of each other is important because the basin-margin-to-basin trend in Beeman deposition has long been agreed to have been east–west (cf. Pray, 1961). Thus, the Dry Canyon West and Space History B sections are not only geographically very close to each other, but they are essentially along depositional strike, so they should show a very similar succession of facies if there were an areally extensive driver of deposition, such as glacioeustasy.

This continuity of facies succession, however, is not the case. The Dry Canyon West outcrop section (Figure 32) encompasses about 119 m of the upper Beeman, so it should overlap the Space History C section from about the 65 m level to the base of the Holder Formation (Figure 25, units 41–173). The Dry Canyon West section begins with what is termed sequence 2C-1a, mostly carbonate overlain by shale and about 8 m thick, whereas that interval at the Space History C section is shale with a prominent bed of sandstone displaying soft-sediment deformation in the middle of the interval. According to Raatz (1996) and Raatz and Simo (1998), their sequence 2C-1a is a transgression followed by a regression, but a reasonable interpretation of the comparable interval at Space History C is that it represents all or most of two transgressive cycles (Figure 25, lower cycle = units 37–41, upper cycle = units 42–48).

Sequence 2C-1b follows at the Dry Canyon West section, largely composed of “restricted limestone” and ~5 m thick. In the Space History C section, this is an interval of shale, overlain by shale interbedded with diverse limestone types (units 44–48). Raatz interprets sequence 2C-1b in Dry Canyon as regressive, whereas we have already noted that this interval in the Space History C section can be considered the uppermost part of a transgression. Sequence 2C-1c at Dry Canyon is ~4 m thick and

is dominated by “peritidal” deposits. The comparable interval at Space History C is mostly shale with two intercalated beds of limestone.

According to Raatz (1996), sequence 2C-1a consists of a transgressive-regressive cycle, 2C-1b is not interpreted by Raatz, and 2C-1c is identified as transgressive. However, given the great dissimilarities in lithologic succession and facies stacking between the Dry Canyon West (Figure 32) and Space History C (Figure 25) sections in this interval, it is difficult to believe that deposition at both sections could reflect a single allocyclic driver.

At the Dry Canyon West section, sequence 2C-2a of Raatz is ~10 m thick, all but about 1 m of which is covered, and the sequence is capped by sandstone overlain by a very thin conglomerate. The comparable interval at the Space History C section is dominantly interbedded shale and limestone with some thin sandstone beds in the lower part (units 53–54). Raatz’s interpretation of sequence 2C-2 as consisting of two to five parasequences of open-marine origin is dependent on sections other than the Dry Canyon West outcrop, which is mostly covered. In the Dry Canyon Core 2 section (Raatz and Simo, 1998: fig. 8), this entire interval is almost totally carbonate. Significantly, there is a sandstone of similar thickness and lithology in the core (Raatz and Simo, 1998: fig. 8, base of unit 15) that logically might be correlated to the sandstone at the sequence top in the Dry Canyon west section. However, Raatz correlates the sandstone in the outcrop to a limestone in the core, which keeps the thickness of sequence 2C-2a in the core and on outcrop approximately the same. This correlation equates very different facies to each other over a short distance. It well exemplifies the statement by Raatz and Simo (1998:165) that “individual bed tracing is only rarely possible for any significant distance.” However, if that is the case, what is driving sedimentation other than climate and tectonism?

At the Dry Canyon West section, sequence 2-C2b is ~5 m thick; the lower part is sandstone, and the upper part is mostly “peritidal” carbonate. The equivalent interval at the Space History C section is shale. Raatz interpreted 2-C2b as a transgression followed by a regression, but such an interpretation cannot be made of the equivalent interval at the Space History C section. Sequence 2C-3a is ~9–10 m thick and mostly covered at the Dry Canyon West outcrop. Raatz interprets it as regressive, and the limestone at the top has a subaerially exposed top (breccia, rhizoliths). The comparable interval at Space History (units 67–87) is interbedded shale and dolomicrite with no easily interpreted deepening or shallowing trends up section. However, the breccia of unit 85 could be equivalent to the subaerial surface at the top of 2C-3a at the Dry Canyon West section.

The rest of the Dry Canyon West section of the upper Beeman Formation (sequences 2C-4a through 4C-1a of Raatz) is mostly limestone beds with some interbedded shale and covered intervals (Figure 32). This interval is ~70 m thick and broadly resembles the comparable interval at the Space History C section (Figure 25, units 86–173), although we are not able to make bed-to-bed correlations between the two sections. Raatz identified at

least nine successive sequences in that interval, but at Space History C we see no facies stacking or disconformities that would allow this many (if any) sequences to be identified. Instead, this interval at the Space History C section looks like deposits of a very shallow marine system with fluctuations in carbonate and clastic deposition that may form some parasequences.

Furthermore, Raatz and Simo (1998:161) speak of “abundant evidence for subaerial exposure” in the upper Beeman that is not evident in our Space History C section. They also correlated their sequences one to one to Midcontinent cyclothems (based on Heckel, 1994) with very little age control to support those correlations. Thus, their Dry Canyon transect has nine biostratigraphic (conodont-based) control points that provide direct age data for 8 of the 53 parasequences identified, but our conodont data revise those correlations (Table 3 and see below).

This comparison raises real questions about the sequence stratigraphic interpretations of Raatz (1996) and Raatz and Simo (1998) that undermine their conclusions that the upper Beeman Formation is a succession of glacioeustatically driven depositional cycles. Thus, we see much disorganized facies stacking in the Beeman and little lateral continuity of facies, particularly those that can be interpreted as transgressive or regressive. We see little or no continuity of inferred sequence boundaries as well. This disorganized facies stacking and the lack of lateral continuity over short distances suggest a complex interplay of factors, both autocyclic and allocyclic, was an important driver of Beeman deposition.

In the Space History C section, there are well-developed cycles in the lower, dominantly siliciclastic portion of the section (seven fining and deepening upward sequences can be inferred; Figure 25) between the 21 and 80 m levels of the section. Above the 80 m level, the section consists of alternating beds of limestone/dolomite and shale/siltstone that do not form clear cycles. Perhaps units 80–81, 144–146, and 147–149 represent deepening upward successions. The remainder is alternating carbonate-shale couplets (probably parasequences) with no clear transgressive-regressive trends.

HOLDER FORMATION

LITHOSTRATIGRAPHY

Pray (1954, 1961) named the Holder Formation for Holder Ridge (section 36, T17S, R11E), and its type section is at Long Ridge, immediately overlying the type section of the Beeman Formation (Figures 7, 24). Pray (1961:76) measured a 65 m thick type section of the Holder Formation, stating that this is an incomplete section, and he estimated another succession of 30 m of Holder Formation is poorly exposed above his type section. Pray (1961) noted that the Holder Formation thickened to the north to as much as 274 m and included strata that Thompson (1942) assigned to his Fresnal Group. The lithological diversity and cyclical nature of Holder deposition was stressed early by Cline (1959) and Pray (1961). The lower Holder is mostly

algal bioherms and biostromes that form cliffs that are 15–23 m thick. The overlying Holder strata are lithologically diverse and locally include some strata of nonmarine origin. Thus, the Holder Formation is up to 274 m thick and composed of a variety of sedimentary rocks (Table 2). It is thickest near Fresnal and La Luz Canyons (at least 260 m) at the northern end of our study area (Figure 1) and thins toward the east and south.

In our study area, the Holder Formation is at least 152 m thick at the Holder bioherm section (Figure 33, top of formation not preserved), 113 m thick at the Space History D section (see Figure 34), and 85 m thick at the Long Ridge section (Figure 24). The Holder Formation can be divided into the lower, cliff-forming Little Dry Canyon Member that is overlain by the slope-forming Mill Ridge Member (Figure 4; see also Figure 33). The Little Dry Canyon Member is well exposed on both sides of Dry Canyon and nearby Little Dry Canyon. It is almost entirely composed of limestone, particularly of massive limestone representing prominent phylloid algal mounds. The maximum thickness of the Little Dry Canyon Member is approximately 23 m. The Mill Ridge Member is composed of alternating covered/shale intervals, sandstone, conglomerate, and limestone-forming mixed siliciclastic-carbonate cycles (Table 2).

Within the Holder Formation, a lateral facies change from cyclic shallow shelf deposits in the east to dark shale and sandstone deposits in the west is well documented (Pray, 1961). This dark shale-sandstone facies is similar to the Panther Seep facies in the San Andres Mountains (Kottowski et al., 1956; Schoderbek, 1994; Seals and Soreghan, 2013). According to Pray (1961), an abrupt facies change occurs in the Holder Formation from the eastern shelf facies to the basinal realm of the Orogrande Basin to the west.

Members of the Holder Formation

We divide the Holder Formation into two members, named here. The lower member is the Little Dry Canyon Member, named after the canyon of that name just northeast of Dry Canyon. The Little Dry Canyon Member is the lower, bioherm-biostrome interval of the formation first recognized by Pray (1961). It is an interval of massive phylloid-algal mounds dominated by lime mud and algae plus intermound and flanking beds of wackestone, packstone, and grainstone derived mostly from the mounds (see Figure 35). The type section of the Little Dry Canyon Member is units 20–34 of our Holder bioherm section measured on the northern flank of Beeman Canyon, just south of Little Dry Canyon (Figure 33). Here, the Little Dry Canyon Member is 23 m thick and includes at least two distinct bioherms (units 20 and 28), reef-flank facies (units 29–31), and a brecciated capping facies (unit 35).

The overlying Mill Ridge Member is named after Mill Ridge, which is just north of Dry Canyon (Figure 2). The type section of the Mill Ridge Member is units 35–149 of our Holder bioherm section (see Figure 33). Our Space History D section (Figure 34) provides a reference section where the Bursum

Formation rests directly (and likely disconformably) on top of the Mill Ridge Member. It is also a section where the underlying Little Dry Canyon Member is either absent or represented by about 2 m of limestone at the base of the Holder Formation (Figure 34, units 2–6), likely because no bioherm developed at this location.

At the type section, the Mill Ridge Member is ~128 m thick. About two-thirds of the section consists of covered intervals that are intermittently exposed to reveal shale or nodular limestone in shale. About one-third of the member at the type section is limestone, with sandstone and conglomerate being minor components (4%) of the type section. Thus, the Holder Formation changes upward from a carbonate-dominated interval with prominent phylloid algal mounds (Little Dry Canyon Member) to an overlying interval with limestone interbedded with clastic sediments, especially shale, sandstone, and conglomerate (Mill Ridge Member).

Long Ridge Section

We measured an ~85 m thick (minus ~5 m of sills) complete section of the Holder Formation at Long Ridge, with Bursum Formation red beds exposed immediately above the Holder Formation (Figure 24). Therefore, unlike Pray (1961), we consider the Holder Formation at Long Ridge to be a complete section of the formation (note that he estimated another 30 m of Holder Formation should be present, and our section adds 25 m of Holder Formation to Pray's section thickness). This, the type section of the Holder Formation, is mostly limestone (62% of the measured section), and about one-third is covered or shale (38% of the section). Shale exposed in the upper part of the section is red or variegated red and yellow. Limestones are mostly massive and algal, but various wackestones, some with nodular texture or oncoids, are present. The base of the Bursum Formation, immediately above the Holder Formation, is red mudstone with an intercalated calcrete bed capped by a characteristic Bursum limestone-cobble conglomerate (Figure 24).

Holder Bioherm Section

At the Holder bioherm section (Figures 33, 35) we measured a 173 m thick succession that includes 21 m of the upper Beeman Formation. The Holder Formation is 152 m thick and overlies a covered/shale interval of the Beeman Formation. The upper contact with the overlying Bursum Formation (Laborcita Member) is not preserved, having been weathered away.

In this section, the Holder Formation is composed of covered intervals (most likely representing shale units) and rare exposed shale, different types of limestone, rare sandstone, and conglomerate. The lower 23 m are almost entirely composed of limestone, mostly massive limestone representing a phylloid algal mound. This lower part is the Little Dry Canyon Member. The overlying succession is assigned to the Mill Ridge Member (Figures 33, 35).

Holder Bioherm

FIGURE 33. Holder bioherm section. See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "HB 20").

Space History D

FIGURE 34. Space History D section. See Table 1 and Figure 2 for location of section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "SHD 17").

FIGURE 35. Overview photograph of the Holder Formation at the Holder bioherm section. Note cliff-forming algal mounds of the Little Dry Canyon Member (above the Beeman Formation) overlain by bedded limestones and clastics of the Mill Ridge Member.

Covered (shale) intervals are 0.1–8.1 m thick. Pink shale containing limestone nodules with fusulinids is exposed in the uppermost part of the measured section. Limestone is represented by individual limestone beds and thicker limestone units that are up to 13.5 m thick. We distinguish (1) thin- to medium-bedded limestone with even bedding (up to 2.5 m thick), (2) wavy-bedded to nodular limestone (0.1–1.3 m thick), and (3) indistinctly bedded to massive limestone, which, locally, is brecciated. The thick massive algal limestone unit near the base of the Holder Formation represents an algal mound with a distinct relief (pinches out laterally) and contains some chert near the top.

Fusulinids are present in some of the bedded limestone units, commonly occurring on the tops of these units. Gastropods are present in the upper part of algal limestone unit 57. Small black oncoids (up to 1 cm in size) occur in the upper part of limestone unit 69. Larger crinoid fragments were observed at the base of limestone unit 100. The coarse-grained limestone bed (rudstone) in the upper part of the section displays cross bedding (unit 129). The most abundant fossils observed on the outcrop are calcareous algae and, subordinately, fusulinids, crinoid fragments, brachiopods, and bryozoans.

Sandstone and conglomerate are present about 43 m above the base of the Holder Formation where cross-bedded conglomeratic sandstone (0.6 m) is overlain by cross-bedded,

poorly sorted, polymict conglomerate (0.9 m) with clasts up to 20 cm in diameter in a sandy matrix. Clast types include chert, quartzite, limestone, and metamorphic (phyllitic) types. The conglomerate has an erosive base and is overlain by thin (0.2 m thick) sandstone.

Approximately 75 m above the base a clast-supported, poorly sorted conglomerate (1.5 m thick) is composed of subrounded to rounded clasts with diameters mostly around 10 cm and up to 20 cm. Most abundant are limestone clasts; less common are chert clasts. The largest clasts occur near the top of the unit. The conglomerate contains a sandy matrix. Thin lenses of cross-bedded sandstone are intercalated.

About 98 m above the base of the section several massive sandstone beds (0.4–0.7 m thick) are exposed. The sandstone interval is 2.7 m thick with a bioturbated limestone bed (0.4 m thick) intercalated in the upper part. A thin carbonate sandstone bed (0.6 m) is intercalated in the uppermost part of the section, approximately 131 m above the base.

Space History D Section

The Space History D reference section of the Mill Ridge Member (Figure 34) is about 1 km southeast of the Holder bioherm section (Figure 2). Here, the Holder Formation is ~113 m

thick and mostly shale or covered intervals likely to conceal shale (about 68% of the section thickness). About 28% of the section is limestone, and only 7% is sandstone, conglomerate, or conglomeratic sandstone. This is a complete section through the Holder Formation, which rests on shale of the underlying Beeman Formation and is overlain by a conglomerate of the basal Bursum Formation (Laborcita Member). The Holder Formation is composed of the following lithologies: shale/covered units, limestone, rare dolomite, sandstone, and rare conglomerate. Shale/covered units are 0.1–11.4 m thick and gray to pink in color. No algal bioherm (Little Dry Canyon Member) is present at the base of the formation.

Limestone units are 0.1–4 m thick and occur as thin- to medium-bedded limestone with even bedding, wavy-bedded to nodular limestone, or indistinctly bedded to massive limestone. Many limestone units contain abundant algae. Fusulinids are present in thin limestone beds at several levels throughout the section. The muddy wackestone of unit 72 contains rhizoliths and displays a subaerial exposure top. Another subaerial exposure surface was recognized on top of limestone unit 98. Chert is rare and was recognized only in the massive limestone of unit 70.

Dolomite occurs as two thin intercalated dolomitic units (0.2 and 0.4 m thick) in the lower part of the section. Sandstone is rare and occurs as 0.5–1.5 m thick units including massive, horizontally laminated, ripple-laminated, and cross-bedded sandstone. Conglomerate is intercalated as one thin, fine-grained conglomerate bed (0.2 m) in the lower part and a limestone and chert-cobble conglomerate with red lime mudstone matrix (0.7 m) in the upper part of the Holder Formation.

La Luz Canyon A Section

In the La Luz Canyon A section (Figure 26), ~53 m of Holder Formation are exposed. This portion of the lower part of the Holder Formation is mostly limestone interbedded with shale, some sandstone, and minor conglomerate. Limestone occurs as thin, individual limestone beds and as thicker limestone units (up to 6 m thick). The following lithotypes are distinguished: (1) thin, even-bedded limestone, (2) medium to thick, even-bedded limestone, (3) wavy-bedded limestone, (4) nodular limestone, and (5) indistinctly laminated to massive limestone (rare). Even-bedded limestone locally contains phylloid algae and crinoid fragments. Shale is mostly pale yellow or pink in color. Sandstone beds are laminated, and the conglomerate beds are composed of limestone cobbles.

SEDIMENTARY PETROGRAPHY

We undertook petrographic study of samples of sandstone and limestone from the Holder Formation at the Space History D and Holder bioherm sections (Figures 36–38). Sandstone of the Holder Formation commonly is mixed siliciclastic-carbonate in composition and moderately to well sorted, containing detrital grains, which are mostly subangular to subrounded (Figure 36).

This sandstone contains abundant monocrystalline quartz grains and carbonate grains (micritic intraclasts) and, subordinately, polycrystalline quartz, detrital feldspar grains (dominantly potassium feldspars), a few microcrystalline quartz (chert) grains, rare micas (muscovite), opaque grains, lithic fragments, and phosphoric grains. Fossils are a common constituent and dominated by echinoderm fragments and, subordinately, fusulinids, brachiopods, and bryozoans, with rare smaller foraminifers (Figure 36). The fossils are very fragmented. Locally, detrital quartz grains display authigenic overgrowths (Figure 36E). Many detrital feldspar grains are altered or partly replaced by calcite. Sandstone is cemented by coarse, blocky calcite (Figure 36A–H).

In pebbly sandstone, sedimentary rock fragments such as reworked siltstone, wackestone, and grainstone are present (Figure 36D). Sandstones of the Holder Formation have a higher maturity than those of the Beeman and Gobbler formations; most of the sandstones plot into the field of sublitharenite and, rarely, into the field of lithic arenite (Figure 18).

Limestone of the Holder Formation is composed of various microfacies types, among which bioclastic wackestone (Figures 37A, 38A,B), including phylloid algal wackestone, smaller foraminiferal wackestone (Figures 37B, 38C), fusulinid wackestone (Figure 37C), and crinoidal wackestone (Figure 38E) to floatstone (Figure 38D,G), and phylloid algal floatstone (Figures 37D–F, 38F; forming phylloid algal mounds) are most abundant. Bioclastic wackestone to grainstone, wackestone to packstone, grainstone to rudstone (Figure 38H), and rudstone (Figure 37G,H) are subordinate.

All microfacies contain a diverse fossil assemblage. The most abundant fossils are phylloid algae, echinoderms (crinoid fragments), bryozoans, brachiopods, and smaller foraminifers. Subordinately present are fusulinids (abundant in fusulinid wackestone to floatstone), ostracods, gastropods, bivalves, brachiopod spines, other algae such as *Epimastopora* and ?*Mizzia*, rare trilobites, and spicules. Undeterminable, recrystallized fossil fragments also are present. In phylloid algal and bioclastic wackestone to floatstone, fossil fragments, particularly phylloid algae, often are encrusted by cyanobacteria and *Palaeonubecularia* (Figure 38F), partly forming oncoids. Rarely, skeletal grains are encrusted by *Tuberitina* and bryozoans. Nonskeletal grains include micritic intraclasts and peloids and rare detrital quartz grains.

Comparison of Sandstones of the Gobbler, Beeman, and Holder Formations

In sandstones of the Gobbler, Beeman, and Holder Formations, mono- and polycrystalline quartz grains are the dominant grain types. Sandstones of the Gobbler Formation are characterized by small to moderate amounts of detrital feldspars and moderate amounts of lithic grains. Fossils are common. Most sandstones contain high amounts of calcite cement, and rarely, some quartz cement is present in the form of authigenic overgrowths.

Sandstones of the Beeman Formation contain moderate to high amounts of detrital feldspars, moderate amounts of lithic fragments, and mostly high amounts of calcite cements. Fossils are rare to mostly absent.

Holder sandstones contain abundant monocrystalline quartz, some polycrystalline quartz, small amounts of feldspars, and rock fragments. The amount of rock fragments is higher than that of feldspars. Fossils are present in all studied samples. Sandstones contain high amounts of calcite cement and, locally, some quartz cement (authigenic overgrowths).

PALEONTOLOGY AND AGE

Fusulinids and conodonts from the Holder Formation indicate a Virgilian age and are discussed in more detail under Marine Microfossils (Table 3). Kues (2004a) described macroinvertebrate fossil assemblages from two stratigraphic levels in the Holder Formation in Dry Canyon roadcuts of U.S. Highway 83. These diverse assemblages included gastropods, bivalves, brachiopods, scaphopods, nautiloids, ammonoids, fusulinids, sponges, rugose corals, bryozoans, echinoids, crinoids, trilobites, ostracods, and fish teeth. One of the assemblages is gastropod dominated, whereas the other is brachiopod dominated. Kues (2004a) judged these fossils to indicate nearshore to moderately offshore quiet, normal-marine paleoenvironments.

Kues (2004a:45), on the basis of a personal communication from W. Raatz, stated that conodonts indicate that the base of the Virgilian is ~80 m below the base of the Holder Formation. Kues also cited a personal communication from G. Wahlman, who noted that the lower 30 m of the Holder Formation “contain species of the fusulinid *Triticites* present in the Midcontinent lower Shawnee Group.” The fusulinid data of Wahlman and Barrick (2018) indicate an early middle Virgilian age for the Holder Formation, not contradicted by our more limited fusulinid and conodont data (see Marine Microfossils section).

Rasbury et al. (1998) published a U-Pb age of paleosol carbonate in the lower Holder Formation of 306 ± 2.6 Ma. Other

Holder paleosols provided much less precise ages: 286 ± 14 and 303 ± 10 Ma (Rasbury et al., 1998, table 1). The 306 Ma age is late Kasimovian (=Missourian) by current calibration (Opluštil et al., 2016), so it is too old to be an approximation of the age of part of the lower Holder Formation. The 303 Ma age is Gzhelian but has a very large error bar, and the 286 Ma age is Permian and also has a very large error bar. It thus seems that none of these U-Pb ages published by Rasbury et al. (1998) provide meaningful age control of the Holder Formation.

DEPOSITIONAL SYSTEMS

The Holder Formation is as much as 275 m thick (Pray, 1961) and, in the upper part, composed of a cyclic succession of dominantly carbonate and subordinately clastic sediments. It includes large phylloid algal bioherms near the base of the formation that formed along the shelf margin (e.g., Wilson, 1967; Raatz and Simo, 1998; Rankey et al., 1999). Locally, crustal shortening caused the formation of north-south-trending folds on the Sacramento shelf during the Virgilian and early Wolfcampian (Wilson, 1967; Toomey et al., 1977; Toomey, 1991), which are related to “pre-Abo” tectonic movements (Raatz, 2002) and influenced sedimentation on the Sacramento shelf.

Phylloid algal mounds at the base of the Holder Formation in Dry Canyon and Beeman Canyon were first described by Plumley and Graves (1953) and later in detail by Wilson (1967, 1975), Toomey et al. (1977), and Toomey (1991), among others. According to Rankey et al. (1999), the Holder Formation is composed of 22 mixed siliciclastic-carbonate, shallowing upward, high-frequency sequences. On the basis of sequence thicknesses, they estimated that the minimum amplitude of relative sea-level falls was 32 m. Phylloid algal and skeletal wackestone, packstone, etc., were deposited in subtidal open-marine settings, and claystone-siltstone was deposited in a shallow to deeper marine environment, partly at a delta front. Fossiliferous sandstone and conglomerate are interpreted as delta stream mouth bar deposits, nonfossiliferous sandstone and conglomerate are interpreted as

FIGURE 36. (Opposite) Thin-section photographs of sandstones from the Holder Formation at Beeman Canyon (A, B) and near the New Mexico Museum of Space History (C–H). A and G under plane light; B–F, H under polarized light. Width of photos: A, B, D = 6.3 mm; C = 1.2 mm; and E–H = 3.2 mm. (A, B) Mixed siliciclastic-carbonate sandstone composed of abundant quartz grains, a few feldspars, abundant carbonate grains (sedimentary rock fragments), and fossil fragments (echinoderms, fusulinids, brachiopods), cemented by calcite. Sample HB 20. (C) Sandstone composed of abundant quartz grains, rare feldspars, many carbonate grains (sedimentary rock fragments), rare micas, few fossils (echinoderms), and fine-grained carbonate cement or matrix. Sample SHD 23. (D) Coarse-grained pebbly sandstone composed of quartz, rare feldspars, many carbonate grains (sedimentary rock fragments), and fossils (fusulinids, brachiopods, echinoderms), cemented by quartz (authigenic overgrowths) and calcite. Sample SHD 24. (E) Sandstone containing abundant quartz grains, rare feldspars, many carbonate grains (sedimentary rock fragments), and fossil fragments (echinoderms, fusulinids) cemented by quartz (authigenic overgrowths, well developed around the detrital quartz grain in the center) and some calcite cement. Sample SHD 26. (F) Mixed siliciclastic-carbonate sandstone containing abundant detrital quartz grains, rare feldspars, many carbonate grains (sedimentary rock fragments), and fossils (mostly echinoderms), cemented by quartz (authigenic overgrowths) and blocky calcite. Sample SHD 29. (G, H) Fossiliferous, mixed siliciclastic-carbonate sandstone containing quartz, rare feldspars, many carbonate grains (sedimentary rock fragments), and fossil fragments including echinoderms and brachiopod shell fragments, cemented by blocky calcite. Sample SHD 76.

fluvial deposits, and red claystone-siltstone is interpreted as an indicator of subaerial exposure (paleosol, alluvial sediments).

Carr and Scott (1990) studied in detail the sandstone facies of the Holder Formation. They distinguished three main sandstone subfacies: (1) thin-bedded, fine-grained sandstone with horizontal lamination and hummocky cross stratification with graded tempestite sequences; (2) fine-grained sandstone similar to subfacies 1 but intercalated within medium- to thick-bedded, large-scale foresets up to 5 m high; and (3) medium-grained sandstone with cross bedding, ripples, and abundant marine trace fossils. They interpreted the sandstones as deposits of a storm-dominated, inner shelf environment representing sand ridges that migrated and accreted seaward during storms. Goldstein (1988) recognized from cement stratigraphic analysis that freshwater cementation occurred during 15 periods of subaerial exposure within a 180 m thick cyclic succession of the Holder Formation.

The basinal equivalent of the Holder Formation is the Panther Seep Formation, which in the southern San Andres Mountains is up to 1,019 m thick and was deposited near the center of the Orogrande Basin in low-energy, euxinic to hypersaline, subaerial to shallow marine environments (Schoderbek, 1994; Soreghan, 1994a,b; Seals and Soreghan, 2013). The different lithofacies of the Panther Seep Formation are arranged to form cycles, and most of the cyclicity was probably caused by glacioeustatic sea-level fluctuations, but sedimentation of the Panther Seep Formation was also influenced by local tectonics (Schoderbek, 1994). The Panther Seep Formation represents shallow-water deposits of the most distal part of the Orogrande Basin where subsidence (and sedimentation) rates were high. Highstand deposits are represented by dark shale, mud-dominated limestone and finely laminated gypsum (Schoderbek, 1994).

All sandstones of the Holder Formation that we studied under the microscope contain marine fossils and display a higher maturity than sandstones of the Gobbler and Beeman

Formations. The presence of marine fossils indicates deposition in a nearshore environment such as a delta-stream mouth bar (Rankey et al., 1999) or deposits of a storm-dominated inner shelf environment (Carr, 1983; Carr and Scott, 1990) and therefore indicates a degree of autocyclic control of sandstone deposition, likely due to delta switching (Carr, 1983). Holder limestone is dominantly composed of wackestone to floatstone containing a diverse fossil assemblage, indicating deposition in an open, normal-marine, low-energy shelf environment. Microfacies indicating a high-energy depositional setting, such as grainstone, are subordinate. However, plant fossils and myalinid bivalves that occur in the Holder Formation near the mouth of La Luz Canyon probably point to deposition in a slightly brackish water environment.

A distinct facies is represented by the phylloid algal mounds that are present at and near the base of the Holder Formation and that are well exposed on both sides of Dry Canyon, at Beeman Canyon and Indian Wells Canyon, thus extending over approximately 4.9 km in a northwest-southeast direction. The maximum thickness of these phylloid algal mounds is approximately 23 m (at Dry Canyon). According to Toomey et al. (1977) and Toomey (1991) and as shown by other detailed studies (see citations in Previous Studies above), the mounds show a distinct zoning. They are composed of a basal “core” facies of a shelly hardground rich in brachiopods that formed in a low-energy, subtidal environment. This shelly core is overlain by the phylloid algal mound facies composed of phylloid algal packstone and bafflestone.

The dominant phylloid alga in this mound facies is *Eugonophyllum*. The phylloid algal mound grew into increasingly shallow water, which stopped mound growth, and algal/foraminiferal bindstones containing abundant dasycladacean algae were formed. Continued shoaling caused deposition of grainstones with abundant tubular foraminifers and ostracods or thin oolitic grainstones that form the capping bed facies. Red weathering zones and in situ brecciation indicate that the bioherms

FIGURE 37. (*Opposite*) Thin-section photographs of microfacies types from limestones of the Holder Formation near the New Mexico Museum of Space History (A–C, H) and at Beeman Canyon (D–G). All photos under plane light. Width of photographs: A, C–G = 9 mm; B and H = 4.6 mm. (A) Bioclastic wackestone containing a diverse fossil assemblage of echinoderms, fusulinids, brachiopods, bryozoans, smaller foraminifers, and ostracods. Sample SHD 5. (B) Bioclastic (foraminiferal) wackestone containing abundant smaller foraminifers (mostly calcivertellids), subordinately recrystallized phylloid algae, echinoderms, small gastropods, and ostracods. Sample SHD 70. (C) Fusulinid wackestone-floatstone. Fusulinids, echinoderms, and oncoids float in fine bioclastic matrix containing a diverse fossil assemblage. Sample SHD 104. (D) Floatstone containing recrystallized phylloid algae, which are partly encrusted by *Palaeonubecularia*, fusulinids, and many small fossil fragments including echinoderms, ostracods, and smaller foraminifers. Sample HB 22. (E) Phylloid algal floatstone. Large recrystallized thalli of phylloid algae, partly encrusted by cyanobacteria, are embedded in partly peloidal micrite. Other fossils are rare. Sample HB 17. (F) Phylloid algal floatstone-rudstone composed of recrystallized phylloid algae, partly encrusted by cyanobacteria, and subordinately other fossils such as bryozoans, smaller foraminifers, ostracods, and small gastropods. Sample HB 5. (G) Rudstone (coquina) composed of densely packed shells of brachiopods and bivalves? Other fossils such as echinoderms, foraminifers, and small gastropods are rare. The shell fragments are oriented parallel to the bedding. Sample HB 7. (H) Rudstone composed of reworked intraclasts, echinoderms, shell debris, bryozoans, ostracods, and smaller foraminifers. Sample SHD 17.

were ultimately exposed subaerially. Laterally, the mound facies grades into the flanking bed facies composed of thin-bedded and steeply inclined (up to 35°) limestone beds with a diverse fossil assemblage of dasycladacean algae (*Macroporella?*), encrusting foraminifers, fusulinids, worm tubes, myalinid bivalves, and ostracods (for details see Toomey et al., 1977; Toomey, 1991). The algal mounds, which formed on the relatively narrow Sacramento shelf margin, are related to tectonic activity. According to Toomey (1991), the mounds formed during intervals when tectonic activity on the La Luz anticline produced a distinct progression of biohermal growth stages; thus, the algal mounds formed during upward shoaling phases. Wilson (1972a) showed that the Holder bioherms are diachronous, with the oldest bioherms in the southeast and the mounds becoming progressively younger toward the northwest.

Kues (2004a:45) stated that “the Holder Formation was deposited on a narrow shelf a few kilometers in width, between the Pedernal uplift to the east and the rapidly subsiding Orogrande Basin to the west.” He further noted that “the bioherms at the base of the Holder grew along the seaward edge of the shelf, and were influenced by the development of the local, north-south-trending, La Luz anticline (Wilson, 1967; Toomey, 1991).”

Holder deposition has long been considered cyclic and primarily driven by glacioeustasy (Wilson, 1972a). However, the evident facies change within the formation, from algal bioherms below to mixed carbonate/siliciclastic above, is not explained by glacioeustasy alone. In the broad Midcontinent shelf, subject to little of the tectonism of the study area in western Pangea, the entire Virgilian is organized into 50 cycles and 15 major cycle groupings (Heckel, 2003, 2013). Diverse analyses identify about 15 to 22 cycles in the upper member of the Holder Formation. These Holder strata represent only a portion of Virgilian time, and available age data are not precise enough to permit

a close correlation of the Holder cycles to the Midcontinent cycles. Furthermore, a case has been made for local tectonism (rising La Luz anticline) influencing Holder deposition (Wilson, 1972b) and for the influence of autocyclic delta switching on sandstone deposition in part of the formation (Carr, 1983; Carr and Scott, 1990). It thus seems likely that at least some of the Holder cycles, or parts of them, were glacioeustatically driven, but more age data and detailed facies analyses are needed to identify those cycles or those facies within cycles driven by local events and conditions.

BURSUM FORMATION

INTRODUCTION

In central New Mexico, the late Virgilian–early Wolfcampian Bursum Formation represents the transitional facies between the underlying, dominantly shallow marine Pennsylvanian units of mixed marine carbonates and mostly marine clastics (Atrasado, Bar B and Holder formations) and overlying nonmarine red beds of the lower Permian (Wolfcampian) Abo Formation (e.g., Lucas and Krainer, 2004; Krainer and Lucas, 2013; Figure 39). In the Sacramento Mountains, the Bursum Formation crops out over an ~27 km long, northwest–southeast outcrop belt from north of Tularosa to south of Dry Canyon (Otté, 1959a; Pray, 1961). It changes in thickness from an estimated 300 m near Tularosa to a pinchout 5.5 km southeast of La Luz Canyon (Otté, 1959a).

Bursum strata in the Sacramento Mountains represent a mixture of facies of nonmarine and marine siliciclastic sediments and shallow marine limestone, which accumulated on the narrow eastern shelf of the Orogrande Basin across the Pennsylvanian–Permian boundary (e.g., Otté, 1959a; Fly, 1986; Krainer et al., 2003; Krainer and Lucas, 2013). In general, these

FIGURE 38. (*Opposite*) Thin-section photographs of microfacies of limestones from the Holder Formation at Beeman Canyon (A, E) and near the New Mexico Museum of Space History (section Space History D; B–D, F–H). All photos under plane light. Width of photographs: A, B, D–H = 9 mm; C = 1.8 mm. (A) Bioclastic wackestone, bioturbated, composed of recrystallized, unidentifiable skeletons, echinoderms, brachiopods, smaller foraminifers, bryozoans, and rare other fossil fragments. Sample HB 15. (B) Bioclastic wackestone containing fragments of brachiopods, fusulinids, bryozoans, smaller foraminifers, ostracods, echinoderms, and calcareous algae embedded in peloidal micrite. Sample SHD 5. (C) Foraminiferal wackestone containing abundant calcivertellid foraminifers, subordinately ostracods, echinoderms, and recrystallized skeletons. Sample SHD 32. (D) Wackestone–floatstone containing a few larger fossils such as fusulinids, gastropods, calcareous algae, and shell fragments and small skeletons including smaller foraminifers, ostracods, and brachiopods. Sample SHD 19. (E) Crinoidal wackestone–rudstone containing abundant crinoidal fragments, a few bryozoans, brachiopods, trilobites, and smaller foraminifers. Sample HB 2. (F) Phylloid algal floatstone composed of recrystallized fragments of phylloid algae, smaller foraminifers, echinoderms, and ostracods. Some of the algal thalli are encrusted by *Palaeonubecularia* and cyanobacteria. Sample SHD 78. (G) Floatstone containing large gastropods embedded in matrix of bioclastic wackestone containing a diverse fossil assemblage of bryozoans, algae, smaller foraminifers, ostracods, a few intraclasts, and peloidal matrix. Sample SHD 72. (H) Grainstone–rudstone, poorly washed, containing many intraclasts, recrystallized skeletons, smaller foraminifers, echinoderms, shell fragments, gastropods, ostracods, and brachiopods. Some of the skeletons are encrusted by cyanobacteria. Sample SHD 96.

FIGURE 39. Stratigraphic nomenclature and correlation of Bursum Formation across New Mexico (after Lucas and Krainer, 2004).

strata are more marine in origin to the west (basinward) than to the east. The upper part of the Bursum Formation contains algal bioherms northeast of Tularosa that are up to 20 m high (see, e.g., Figures 42, 43A, 44A).

Because the Bursum Formation is mostly of Permian age, we do not present all of the diverse data that we have collected on the formation (those will be published elsewhere). Instead, we focus on three key sections as representative of the Bursum Formation in the northern Sacramento Mountains: Tularosa Clay Pit (see Figures 40, 44B), Fresno Canyon (see Figures 41, 44C), and Scorpion Mound (see Figures 42, 43, 44A).

LITHOSTRATIGRAPHY

Otté (1959a) coined the term Laborcita Formation for sedimentary rocks composed largely of gray and red mudstone, gray limestone, sandstone, and conglomerate that Pray (1954, 1961) termed the Bursum Formation (Figure 4). Lucas and Krainer (2004) identified the Laborcita as a member of the Bursum Formation (Figure 39). The Laborcita Member is exposed in the northern Sacramento Mountains, between the underlying Holder Formation and overlying red beds of the Abo Formation. At the type locality, the Laborcita Member is 146 m thick according to Otté (1959a). At Scorpion Mound (see Figures 42, 43), the Laborcita Member is 268 m thick, and the base is not exposed (see Figure 43). The boundary between the Laborcita Member and overlying Abo Formation is drawn at the top of the highest marine limestone of the Laborcita Member. The base of the Laborcita Member is a conglomerate of siliciclastic red beds on top of marine carbonate or clastic strata at the top of the Holder Formation (see, e.g., Figure 41).

Tularosa Clay Pit Section

In the Tularosa Clay Pit, 69 m of the lower part of the Laborcita Member of the Bursum Formation are exposed, but its basal contact is covered (Allen et al., 2009; Figure 40). The section can be divided into three parts. The lower part is 19.5 m thick and composed of gray to dark gray shale (0.2–5.1 m thick), intercalated limestone beds (0.1–0.3 m thick), gray calcareous shale (0.5 m thick), and a conglomerate bed (0.7 m thick; see Figure 44B). The shale contains fossils (mainly gastropods but also ammonoids and other mollusks; see below), particularly in the shale unit above the conglomerate bed (unit 13). Limestone beds are dark gray and composed of lime mudstone (bioclastic mudstone to wackestone). The conglomerate is clast supported and composed of well-rounded clasts, mostly 2–5 cm in diameter. The maximum clast size is 10 cm. The conglomerate is composed of quartzite, chert, volcanic, sandstone, and limestone clasts. It is overlain by a thin sandstone bed with a few pebbles near the top.

The middle part of the section (Figure 40, units 18–54) consists of siltstone and mudstone (0.4–3.4 m thick), two thin limestone beds (0.1 and 0.2 m thick), intercalated sandstone beds and sandstone units (0.1–3 m thick), one fine-grained conglomerate (0.3 m thick), and covered intervals (0.7–1.6 m thick). The siltstone and mudstone are brownish, greenish, and reddish. The thin limestone beds weather brownish, are dark gray, and display a muddy texture (mudstone-wackestone). Sandstone is fine to medium grained. Some of the sandstone beds have an erosional base. The following lithotypes are distinguished: horizontally laminated sandstone; fine-grained, ripple-laminated sandstone; massive sandstone; and, rarely, sandstone with small-scale trough

Tularosa Clay Pit

FIGURE 40. Tularosa Clay Pit section. See Figure 8 for legend to lithologic symbols. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "TCP 3"). Base of section is at Universal Transverse Mercator (UTM) zone 13, 407571E, 3660521N, and top is at 407620E, 3660464N (NAD [North American Datum] 83).

cross bedding. The sandstone of unit 20 contains a few pebbles with diameters up to 3–4 cm and partly displays soft-sediment deformation structures. The sandstone on top of the limestone bed (units 40, 41) contains crinoid fragments. The conglomerate bed (unit 30) is composed of intraclasts with diameters up to 3 cm that float in sandy matrix.

The upper part of the section (Figure 40, units 55–78) is composed of pink clay-mudstone (0.3–4.3 m thick), covered intervals (0.3–5.3 m thick), intercalated sandstone (0.2–1.7 m thick), and shale with carbonate nodules (5–10 cm, up to 20 cm in diameter) containing brachiopods (0.8 m thick). The sandstone is coarser grained than in the middle part. The sandstone beds are commonly horizontally laminated, but thicker sandstone beds are trough cross-bedded, rarely massive. In the upper part, sandstone beds (units 71, 72, 76) have pebbles with diameters up to 4 cm. In the topmost sandstone, thin conglomerate layers are intercalated, which are polymict and contain pebbles up to 10 cm in diameter. In general, the succession exposed in the Tularosa Clay Pit shows a coarsening upward trend (almost no sandstone in the lower part, fine-grained sandstone in the middle part, medium- to coarse-grained sandstone in the upper part).

Fresnal Canyon Section

The exposed thickness of the Bursum Formation at Fresnal Canyon is approximately 100 m thick, but the top of the formation is not in the section (Figure 41). The basal contact on the underlying Holder Formation is well exposed and is a conglomerate bed displaying an erosive base. The Bursum Formation is composed of various lithotypes of siliciclastic sediments and limestone that, in the lower part of the section, form cycles (Krainer et al., 2003; Krainer and Lucas, 2013). Cycles commonly start with a conglomerate bed that grades upward into alternating sandstone, siltstone and shale, and thin- to thick-bedded fossiliferous limestone, indicating a deepening upward trend (transgressive cycles). Regressive cycles are rare and thin.

Conglomerates are polymict, poorly sorted, clast supported, and up to 2.5 m thick. Coarse-grained, partly pebbly sandstone beds are massive, indistinctly laminated or cross-bedded. Fine-grained sandstone and siltstone form thin beds intercalated in shale and appear massive or horizontally laminated, rarely ripple laminated, and cross-bedded. Shale and silty shale are well exposed along the roadcut at Fresnal Canyon. The color is either gray to greenish gray or reddish to pink.

Limestone lithotypes are (1) thin-bedded limestone with even to wavy bedding (dominantly wackestone and packstone, rarely grainstone), (2) thin-bedded limestone with wavy to nodular bedding (bioclastic wackestone, rarely packstone and grainstone), (3) thick-bedded limestone composed of bioclastic wackestone to grainstone and algal wackestone, and (4) a thin carbonate bed intercalated in shale, interpreted to be of pedogenic origin (for details see Krainer et al., 2003).

Scorpion Mound Section

At Scorpion Mound (Figures 42, 43, 44A), northeast of Tularosa, the Laborcita Member of the Bursum Formation is at least 268 m thick: the base of the formation, its contact with the underlying Holder Formation, is not exposed. The Laborcita Member is overlain by nonmarine red beds of the Abo Formation. In general, we can summarize the Bursum Formation section as follows: (1) the lower 90 m of the succession are entirely siliciclastic; (2) from 90 to 135 m the succession is mixed siliciclastic-carbonate; (3) from 135 to 151 m the succession is carbonate, dominantly composed of a massive mound facies; and (4) the upper part of the member, from 151 to 168 m, is again mixed siliciclastic-carbonate.

Lower Part (Lower 90 m)

The basal 28 m are composed of greenish shale-siltstone with intercalated sandstone beds and rare, coarse-grained, pebbly sandstone (Figure 43). A few covered (shale) intervals are up to 2.9 m thick, shale-siltstone intervals are up to 1.6 m thick, intercalated sandstone beds are 0.1–0.5 m thick, and pebbly sandstone is up to 0.6 m thick. From 29 to 90 m, the succession is composed of alternating shale-siltstone, sandstone, and conglomerate. Individual units are thicker: shale-siltstone intervals are up to 7.3 m thick, sandstone units are up to 4.2 m thick, and conglomerate units are up to 2.1 m thick. Shale-siltstone units rarely contain carbonate nodules (pedogenic?).

Sandstone beds either are massive or display horizontal lamination, rare ripple cross bedding, and, in thicker units, trough cross bedding. Rarely, syndimentary deformation structures are observed.

Conglomerate units are either massive or trough cross-bedded. Massive conglomerate is either matrix supported or clast supported. Conglomerate units contain clasts that are rarely up to 30 cm in diameter; mostly, the grain size is less than 10 cm. Conglomerates are poorly sorted, and the clasts are rounded. The most abundant clast types are quartzite and rhyolite; granite is less common, and carbonate clasts are rare. Conglomerate units display a fining upward trend, and their bases are commonly erosive.

Sandstone is mostly fine to medium grained, is moderately to poorly sorted, and composed of angular to subrounded grains. The most abundant grain type is monocrystalline quartz. Polycrystalline quartz is subordinate. Rarely, rock fragments composed of quartz and feldspar, chert grains, detrital chlorite, and glauconite are present. The sandstone contains abundant detrital feldspars, which are mostly K-feldspars, and rare plagioclases. Detrital feldspars are strongly altered and partly replaced by calcite. Sandstones contain clayey matrix with chlorite and some coarse, blocky calcite cement that replaces detrital feldspars and, subordinately, quartz grains. Locally, dolomitic cement is present. Rarely, detrital quartz grains display authigenic overgrowths.

Fresnal Canyon

FIGURE 41. Fresno Canyon section. See Figure 8 for legend to lithologic symbols. Base of section is at UTM zone 13, 414536E, 3649595N, and top is at 414733E, 3649563N (NAD 27). Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, “FC 9”).

FIGURE 42. Cross section of Scorpion Mound, redrawn from Bowsher (1986: fig. 32). The algal mound complex overlies westward prograding siliciclastic sediments deposited on the shelf between the Pedernal uplift to the east and the Orogrande basin to the west.

Mixed Siliciclastic-Carbonate Succession (90–135 m)

From 90 to 135 m (Figure 43), the succession is composed of shale-siltstone with intercalated sandstone and limestone. Shale-siltstone units and covered (shale) intervals are up to 5.2 m thick but mostly less than 1 m thick. Sandstone units are up to 1.3 m thick, and limestone units are up to 1.8 m thick.

The following sandstone types are observed: massive, horizontally laminated, ripple laminated, and trough cross-bedded. Sandstones display the same textural and compositional features as sandstones lower in the section.

Limestone is gray to dark gray and micritic and occurs as individual limestone beds (5–30 cm thick) and bedded limestone units. It is composed of various microfacies types such as bioclastic mudstone, peloidal packstone, bioclastic wackestone, foraminiferal wackestone, oolitic grainstone, and oncoidal rudstone.

Bioclastic mudstone and peloidal packstone contain a low-diversity fossil assemblage composed of ostracods. The other microfacies types are characterized by a more diverse fossil assemblage composed of fragments derived from brachiopods, bivalves, gastropods, echinoderms, ostracods, smaller foraminifers, and rare trilobite fragments. Grainstones contain small amounts of detrital quartz and feldspar grains. Foraminiferal wackestone contains abundant calcivertellid foraminifers, subordinate ostracods, echinoderms, and other bioclasts. Oncoids are formed by *Girvanella*, *Archaeolithophyllum lamellosum*, and sessile foraminifers. The foraminiferal assemblage includes *Calcivertella* (abundant), *Eotuberitina*, *Globivalvulina* ex gr. *bulloides*, *Glomospiroides*, *Pseudovermiporella*?, and *Syzrania*.

Mound Facies (135–151 m)

The succession from 135 to 151 m is mainly composed of massive limestone (mound facies; Figures 42, 43). A thin

interval (0.8 m) of gray, bedded limestone is overlain by 11.2 m of massive limestone (algal mound), followed by 0.6 m of cover (shale), 0.6 m of bedded limestone, and three thick beds of algal limestone measuring 3 m thick, overlain by a thicker interval of mostly covered gray shale. The bedded gray limestone at the base of the massive mound facies is composed of bioclastic wackestone.

The most common microfacies of the massive limestone of the mound facies is phylloid algal wackestone to floatstone that is composed of large (up to several centimeters long), completely recrystallized phylloid algae (Figure 45B,D–F). The algae are still partly in life position and partly toppled, and locally, the algal thalli are oriented parallel to the bedding, indicating short distances of transport (phylloid algal floatstone to rudstone; Figure 45A,C). Some algal fragments are encrusted by *Girvanella*, *Palaeonubecularia*, *Tubiphytes*, and bryozoans.

The phylloid algae are embedded in a matrix that is composed of (1) fine-grained bioclastic wackestone to grainstone (Figure 45C), (2) peloidal wackestone, and (3) peloidal grainstone (Figure 45B,D). Fine-grained bioclastic wackestone (Figure 45G) contains small skeletons of algae, smaller foraminifers, ostracods, bryozoans, shell debris, small gastropods, and rare echinoderms. Peloidal wackestone and grainstone contain only a few fossils, in particular ostracods and smaller foraminifers. Locally, in situ brecciation is observed (Figure 45H).

Peloidal grainstone and bioclastic wackestone are intercalated in the mound facies. The peloidal grainstone contains a few skeletons of ostracods and smaller foraminifers, rare small echinoderm fragments, bryozoans, and other small, fragmented skeletons. The bioclastic wackestone is diverse, containing shell fragments of bivalves, gastropods, brachiopods, bryozoans, echinoderms, ostracods, smaller foraminifers, and, locally, large calcisponges. Sponges are partly encrusted by *Girvanella* and *Archaeolithophyllum*. Many skeletons are encrusted by

FIGURE 43A. Scorpion Mound section of the Bursum Formation. See Figure 8 for legend to lithologic symbols. Base of section is at UTM zone 13, 407280E, 3661550N; and top is at 407859E, 3662204N (NAD 27). The lower to middle parts of the Bursum Formation are portrayed here. Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, “SM 3”).

FIGURE 43B. Scorpion Mound section of the Bursum Formation. The upper part of the Bursum Formation and base of Abo Formation are portrayed here. See Figure 8 for legend to lithologic symbols. Base of section is at UTM zone 13, 407280E, 3661550N; and top is at 407859E, 3662204N (NAD 27). Stratigraphic levels at which samples for petrographic analysis were collected are indicated by acronyms and numbers (for example, "SM 3").

FIGURE 44. Photographs of outcrops of the Bursum Formation. (A) Mound facies of the Bursum Formation at Scorpion Mound. (B) Lower, shale-dominated part of Bursum Formation section at Tularosa Clay Pit. (C) Mixed red-bed clastics and carbonates of the lower part of the Bursum Formation at Fresno Canyon.

Girvanella. The matrix is micritic, pelmicritic, and locally pel-sparite. Stromatactis structures (centimeter-large voids) are filled with micrite and cement.

The massive mound facies laterally extends over several hundred meters, and the top of the mound complex locally displays a relief of up to 2 m that is filled with bedded limestone, locally overlain by a 2 m thick phylloid algal mound facies (Figure 42). The bedded intermound facies is composed of mainly bioclastic wackestone to packstone and also of phylloid algal wackestone and fusulinid wackestone. All microfacies types contain a diverse fossil assemblage (phylloid algal fragments, shell fragments of bivalves, gastropods, brachiopods, echinoderms, ostracods, smaller foraminifers, fusulinids, bryozoans, very rare trilobites, rare *Epimastopora*).

The mound complex extends from Scorpion Mound 1.5 km NE of Tularosa and approximately 5 km to the north (entrance of Coyote Canyon). Cys and Mazullo (1977), Mazullo and Cys (1979), Mazullo (1988), and Bowsher (1986) distinguished four facies in the mound complex: (1) aragonite massive facies, (2) phylloid algal-aragonite-cement facies, (3) phylloid algal wackestone-packstone facies, and (4) breccia facies (see Krainer et al., 2007). The core of the mound is composed of the aragonite massive facies and is overlain by the phylloid algal-aragonite-cement facies and, finally, by the phylloid algal wackestone-packstone facies. Short periods of subaerial exposure caused the formation of fissures and local brecciation of the mound facies.

The bedded limestone that overlies the massive mound facies is composed of bioclastic wackestone containing a diverse fossil assemblage: shell fragments of bivalves, gastropods, brachiopods, bryozoans, echinoderms, ostracods, *Tubiphytes*, calcisponges, and many smaller foraminifers (*Calcivertella*, *Deckerella*, *Endothyra*, *Eotuberitina*, *Geinitzina*, *Palaeonubecularia*, *Spireitlina* cf. *bashkirica*, *Syzrania*, *Tetrataxis*).

Upper Part (151–268 m)

The upper part of the Bursum Formation at Scorpion Mound (Figure 43) is mainly composed of shale-siltstone, which from 151

to 187 m in the section is gray and partly fossiliferous and from 187 to the top of the Bursum Formation is mainly reddish. In the lower part, between 170 and 195 m, thin, massive, and horizontally laminated, fine-grained sandstone beds (0.1–0.3 m); one planar cross-bedded pebbly sandstone interval (1.6 m); and even-to wavy-bedded and nodular limestone intervals (0.3–1.4 m) are intercalated. A few covered (shale) intervals are present, too.

The uppermost part of the section (195–268 m) is composed of thicker shale-siltstone units (0.5–7.6 m thick, mostly >1 m). Two red shale-siltstone intervals contain carbonate nodules in the upper part (pedogenic nodules?). Intercalated beds are (1) dark red arkosic sandstone, up to 1.4 m thick, trough cross-bedded and massive, commonly with an erosive base; (2) matrix-supported conglomerate, 1 m thick, intercalated in coarse pebbly sandstone; (3) coarse pebbly sandstone with abundant fossils; (4) pedogenic limestone horizon (0.2 m); and (5) limestone intervals (0.7–2.6 m) that are thin bedded (even-wavy), nodular, or thick bedded to massive.

Sandstone is moderately to poorly sorted, and the grains are mostly subangular to subrounded. The most common grain type is monocrystalline quartz (including rare volcanic quartz). Polycrystalline quartz is subordinate. Other grain types are rare and include granitic rock fragments (quartz and feldspar), sedimentary rock fragments, chert, and quartzite. The sandstone contains abundant detrital feldspars that are strongly altered and partly to completely replaced by calcite. Most detrital feldspars are K-feldspars, and plagioclase is rare. Other rare grains are detrital chlorite grains and very rare glauconite grains. The detrital grains are cemented by calcite.

Limestone is composed of (1) phylloid algal wackestone, (2) bioclastic wackestone, (3) bioclastic grainstone/packstone, (4) peloidal-bioclastic packstone/wackestone/grainstone, (5) mixed siliciclastic-carbonate sandstone, and (6) caliche carbonate. The composition of the phylloid algal wackestone is the same as that of the massive mound facies. The bioclastic wackestone is composed of a diverse assemblage, including abundant phylloid algal fragments. Bioclastic wackestone of the wavy-bedded to nodular limestone contains a low-diversity fossil assemblage

FIGURE 45. (Opposite) Thin-section photographs of the mound facies of the Bursum Formation at Scorpion Mound. All photos under plane light. (A) Phylloid algal floatstone to rudstone composed of recrystallized and broken fragments of phylloid algae (probably *Eugonophyllum*) embedded in peloidal micrite. The algal thalli are aligned parallel to the bedding, indicating some transport. Sample SM 18; width of photograph is 17 mm. (B) Phylloid algal floatstone. Large phylloid algae (probably *Eugonophyllum*), probably in life position, are embedded in peloidal grainstone and wackestone. Sample SM 20B; width of photograph is 20 mm. (C) Phylloid algal rudstone composed of broken fragments of phylloid algae that are aligned parallel to the bedding (transported) and embedded in wackestone to poorly washed grainstone composed of skeletal grains, intraclasts, peloidal micrite, and some cement. Sample SM 21A; width of photograph is 21 mm. (D) Phylloid algal floatstone composed of large thalli of phylloid algae (probably *Eugonophyllum*), probably in life position, embedded in peloidal wackestone and grainstone with abundant small fossil fragments. Sample SM 21A; width of photograph is 26 mm. (E, F) Phylloid algal floatstone. Broken fragments of phylloid algae (probably *Eugonophyllum*) are oriented parallel to the bedding and embedded in peloidal micrite. E = sample SM 22A; width of photograph is 21 mm. F = sample SM 23A; width of photograph is 12 mm. (G) Bioclastic wackestone containing a diverse fossil assemblage of echinoderms, shell fragments (bivalves?), abundant foraminifers, ostracods, and rare bryozoans. Sample SM 22A; width of photograph is 14 mm. (H) In situ brecciated phylloid algal floatstone. Sample SM 21A; width of photograph is 17 mm.

of ostracods, microconchids (“spirorbids”), and rare other skeletons.

Bioclastic grainstone to packstone contains a fossil assemblage similar to that of the diverse wackestone, but rare fusulinids, a few micritic intraclasts, very rare glauconite, and angular detrital quartz grains (~10%) are also present. Peloidal-bioclastic packstone/wackestone/grainstone is poorly sorted and contains larger skeletons of recrystallized mollusk shell fragments and phylloid algae, echinoderms, and smaller fossils such as foraminifers (mostly *Calcivertella* but also *Palaeonubecularia* and *Globivalvulina*), ostracods, and shell debris. Nonskeletal grains are represented by abundant peloids, a few micritic intraclasts, and some detrital quartz grains. Locally, the rock contains micrite (wackestone) or calcite cement (grainstone/packstone).

The mixed siliciclastic-carbonate sandstone is poorly sorted and contains large skeletons (mostly 5–10 mm in diameter) of recrystallized shells derived from bivalves, gastropods, and brachiopods, rare echinoderms and bryozoans, brachiopod spines, smaller foraminifers, algae, ostracods, abundant micritic intraclasts and quartz grains (30–40%), rare chert grains, and calcite cement. Caliche carbonate is nonlaminated and reddish gray, composed of rounded to irregular micritic grains displaying dark micritic envelopes that are partly black to dark red. Rarely, recrystallized skeletons of probable gastropods are present. The pore space is filled with calcite cement.

Fossils from limestone of the upper part of the Bursum Formation at Scorpion Mound recognized in thin section are *Calcivertella*, *Endothyra*, *Geinitzina postcarbonica*, *Globivalvulina* ex gr. *bulloides*, *Nankinella*, *Palaeonubecularia*, *Schubertella*, *Spireitlina*, *Syzyrania*, *Tetrataxis*, *Eugonophyllum*, *Epimastopora*, *Koivaella*, and *Tubiphytes*.

We undertook petrographic study of selected clastic and carbonate rocks of the Bursum Formation (Figures 45, 46). The limestone is composed of various microfacies types, most of them containing a diverse fossil assemblage. Mudstone is rare; most common is wackestone, which occurs as bioclastic wackestone (Figure 46A), phylloid algal wackestone, foraminiferal wackestone to floatstone (Figure 46D), oolitic wackestone, and oncolitic wackestone to floatstone (Figure 46B). Wackestone locally grades into packstone (mostly bioclastic packstone [Figure 46C] and rare peloidal packstone [Figure 46E]), and algal wackestone grades into floatstone. Mudstone is composed of gray micrite containing a few caliche peloids/ooids and a few ostracods (Krainer et al., 2007).

Grainstones are represented by coarse-grained oncolithic grainstone, bioclastic grainstone (Figure 46G), oolitic grainstone (Figure 46F), and peloid/oncoid grainstone (Krainer et al., 2007). From outcrops north of Laborcita Canyon, Fly (1985, 1986) described foreset-bedded sandy grainstones, blue-green algal dismicrites, digitate algal stromatolites, and phylloid algal bindstones.

Siltstones of the Laborcita Member are mixed carbonate-siliciclastic-bioclastic in composition, with varying amounts of siliciclastic grains and bioclasts. Bioclasts are present in all studied siltstone samples in small amounts. Siltstones are moderately to well sorted, and in some samples, poor sorting was observed. The matrix is mostly micrite or pelmicrite, but siltstones cemented by calcite are present as well.

Sandstones are mixed siliciclastic-carbonate in composition, and rarely, quartzarenite is present. Detrital grains are angular to subrounded, but in some medium- and coarse-grained sandstone

FIGURE 46. (*Opposite*) Thin-section photographs of microfacies of the Bursum Formation of the Scorpion Mound section. All photos under plane light. (A) Bioclastic wackestone to floatstone composed of bivalve and gastropod shell fragments, brachiopods, a few echinoderms, ostracods, and smaller foraminifers embedded in micritic matrix. Sample SM 10; width of photograph is 9 mm. (B) Oncoidal floatstone containing small oncoids composed of a nucleus (commonly echinoderm and bryozoan fragments) that is encrusted by *Girvanella* and subordinately by *Palaeonubecularia*. Sample SM 22; width of photograph is 9 mm. (C) Bioclastic wackestone to packstone, fine grained, containing abundant smaller foraminifers (mostly calcivertellids), ostracods, rare fragments of calcareous algae, echinoderms, and bryozoans. Sample SM 42; width of photograph is 4.6 mm. (D) Fusulinid wackestone to floatstone. Besides abundant fusulinid tests, fragments of phylloid algae, a few bryozoans, ostracods, and smaller foraminifers are present. Sample SM 39; width of photograph is 9 mm. (E) Wackestone to packstone composed of abundant peloids, some micritic intraclasts, recrystallized skeletons, ostracods, smaller foraminifers, and a few other fossils. Sample SM 31; width of photograph is 4.6 mm. (F) Grainstone composed of ooids, quartz grains, recrystallized skeletons, intraclasts, rare peloids, a few ostracods, and smaller foraminifers. Sample SM 12; width of photograph is 4.6 mm. (G) Bioclastic grainstone containing recrystallized skeletons, smaller foraminifers (mainly calcivertellids, globivalvulinids), a few fragments of echinoderms and phylloid algae, a few micritic intraclasts, and quartz grains. Many fossil fragments display thin micritic envelopes. Sample SM 28; width of photograph is 4.6 mm. (H) Mixed siliciclastic-carbonate sandstone, fossiliferous, containing a few large skeletons (recrystallized mollusk fragments), brachiopods, rare fragments of echinoderms, bryozoans, smaller foraminifers, algae, ostracods, many micritic intraclasts, some peloids, and abundant detrital quartz grains. Sample SM 32; width of photograph is 9 mm.

the grains are rounded to well rounded. Most of the sandstones are moderately sorted. Siliciclastic grains are more abundant than carbonate lithoclasts, and small amounts of bioclasts are present in most samples.

The dominant siliciclastic grain type is mono- and polycrystalline quartz. Very rarely, detrital feldspars are present. Most sandstones contain abundant recrystallized micritic carbonate grains, and some contain small bioclasts such as shell fragments, echinoderms, and fusulinids (Figure 46H). Sandstones are frequently cemented by calcite replacing quartz, feldspar, and chert grains (Krainer et al., 2007).

Sandstone of the Bursum Formation in the Scorpion Mound section is composed of abundant quartz, abundant detrital feldspar grains (subarkose), and a few rock fragments. Carbonate grains are absent. In a few sandstone beds glauconite grains occur. Like the sandstones, the fine-grained conglomerates are mixed siliciclastic-carbonate in composition and poorly sorted. Fine-grained conglomerates contain abundant bioclasts. The matrix consists of blocky calcite cement. At Scorpion Mound, conglomerate is siliciclastic and contains rounded clasts of quartzite and rhyolite, rare carbonate clasts, and sandy matrix.

PALEONTOLOGY AND AGE

The Bursum Formation in the northern Sacramento Mountains is very fossiliferous. Macroinvertebrate fossils are mostly brachiopods, bivalves, and gastropods (Otté, 1959a; Dutro and Yochelson, 1977; Kues, 1991a, 1991b), and there is one important ammonoid locality in the formation at the Tularosa Clay Pit (Böse, 1920; Miller, 1932; Furnish and Glenister, 1971; Boardman et al., 1994). Few plant fossils have been reported (C. B. Read in King, 1942), but there are diverse vertebrate fossils known, mostly of temnospondyl amphibians and eupelycosaurs (Vaughn, 1969; Lucas et al., 2014a, 2018).

In the northern Sacramento Mountains, most limestone samples of the Bursum Formation contain a diverse microfossil assemblage, including skeletons of calcareous algae (phylloid algae, mostly *Eugonophyllum*, subordinate dasycladacean algae *Epimastopora*, *Anthracoporella*, and other algae), bivalves, brachiopods, bryozoans, echinoderms, fusulinids, gastropods, ostracods, smaller foraminifers, rare trilobites, and *Tubiphytes* (Krainer et al., 2003). In the mound facies, the algal thalli are completely recrystallized and are assigned to the genus *Anchicodium*. Locally, skeletons are encrusted by tubular foraminifers and cyanobacteria, partly forming oncoids.

Fossils of the mound facies (Scorpion Mound) are phylloid algae, *Eugonophyllum* sp., *Eugonophyllum johnsonii*, *Epimastopora*, *Efluegelia*, rare calcisponges, *Tubiphytes*, *Palaeonubecularia*, *Frondicularia*, *Nostocites*, *Ammovertella*, *Calcitornella*, *Calcivertella*, *Climacammina*, *Cornuspira*, *Diplosphaerina*, *Endothyra*, *Endothyranella*, *Eolasiiodiscus*, *Eotuberitina*, *Geinitzina*, *Globivalvulina* ex gr. *bulloides*, *Hedraites*, *Nodosinelloides* ex gr. *longissima*, *Nodosinelloides* ex gr. *mirabilis*, *Nodosinelloides potievskayae*,

Nodosinelloides sp., *Pseudovidalina*, *Spireitlina conspecta*, *Syzrania*, *Tetrataxis*, and *Vervilleina*.

Fossils of the massive mound facies north of Scorpion Mound are *Pseudofusulina*, ?*Schwagerina*, *Calcivertella*, *Climacammina*, *Deckerella*, *Endothyra* ex gr. *bowmanni*, *Eotuberitina*, *Geinitzina* sp., *Geinitzina postcarbonica*, *Globivalvulina* ex gr. *bulloides*, *Hedraites*, *Nodosinelloides ampla*, *Nodosinelloides longissima*, *Nodosinelloides potievskayae*, *Spireitlina*, *Syzrania*, *Tetrataxis*, *Vervilleina*, *Claracrusta catenoides*, *Efluegelia*, *Frondicularia*, *Epimastopora*, and *Eugonophyllum*.

The lower part of the Bursum Formation at the type section north of Fresno Canyon was considered to be of late Virgilian age by Otté (1959a). He drew the base of the Wolfcampian at the base of the first limestone bed containing the fusulinid *Schwagerina*, which occurs 26 m above the base. According to Steiner and Williams (1968), the Laborcita Member is wholly Wolfcampian in age based on fusulinids. However, the stratigraphic extent of their fusulinid assemblages (Steiner and Williams, 1968: text-fig. 1) encompasses only most of the Bursum section, from about 30 m above its base to about 130 m above its base. The fusulinid assemblage consists of large, globose species of *Triticites* (*T. creekensis* Thompson and *T. ventricosus* Meek and Hayden), small species of *Schwagerina* (*S. campensis* Thompson and *S. emaciata* (Beede)), and *Dunbarinella americana* Thompson. Otté (1959a:102–104) listed fusulinid identifications provided by Thompson that indicated that except for the lowermost 60 to 90 ft (18 to 27 m), the fusulinid faunas of the Laborcita are of Bursum age, or early Wolfcampian. Non-fusulinid foraminifers also indicate an early Wolfcampian age (Krainer et al., 2003).

Frederick et al. (2018) described a conodont fauna from the Tularosa Clay Pit in the lower part of the Laborcita Member (Figure 40) that they interpreted to be latest Pennsylvanian (latest Gzhelian) in age. The stratigraphic level bearing conodonts can be correlated with the lower part of the Hughes Creek Shale in the northern Midcontinent region, which lies just below the base of the Permian, as identified by the first occurrence of *Streptognathodus isolatus* in the overlying Red Eagle Formation. The well-studied ammonoid fauna of the Tularosa Clay Pit indicates a latest Gzhelian–Asselian age and is assigned to the *Subperrinites* Zone; conodonts are consistent with this age assignment and suggest further refinement to latest Gzhelian. The stratigraphic position of the Tularosa Clay Pit fossils was also revised by Frederick et al. (2018) to low in the Bursum Formation, approximately 60 m above its base. Ritter (2020) reported the earliest Permian conodont *Streptognathodus isolatus* from the lower part of our Scorpion Mound section of the Bursum Formation.

Therefore, part of the Bursum Formation is Virgilian (which we consider Pennsylvanian), and part is Wolfcampian (which we consider Permian). Clearly, more biostratigraphic research is needed to establish the precise position of the Pennsylvanian–Permian boundary in the Bursum Formation at its diverse outcrops in the northern Sacramento Mountains.

DEPOSITIONAL SYSTEMS

Regional synsedimentary tectonic movements of the Ancestral Rocky Mountain (ARM) orogeny strongly influenced Bursum sedimentation, resulting in conspicuous lateral variations in lithofacies and thickness (Krainer and Lucas, 2009, 2013). The widespread Bursum Formation thus represents significant tectonic activity of the ARM orogeny during the late Virgilian–early Wolfcampian.

In the northern Sacramento Mountains, the Laborcita Member of the Bursum Formation, which is composed of carbonate and clastic rocks, represents a transitional facies between the dominantly carbonate and marine strata of the underlying Holder Formation and the nonmarine red beds of the overlying Abo Formation. The succession is interpreted to consist of cyclic sequences of nonmarine to transitional marine siliciclastics and shallow marine carbonates that were deposited on a narrow shelf between the Orogrande Basin to the west and the Pedernal uplift to the east during early Permian time (e. g., Otté 1959a; Delgado, 1977; Fly, 1985, 1986; Krainer et al., 2003, 2007; Krainer and Lucas, 2013).

Krainer et al. (2003) interpreted the conglomerate facies at Fresno Canyon as fluvial channel-fill deposits. Some Bursum conglomerate beds contain fossils indicating a shallow marine origin. Sandstone and siltstone contain marine fossils that also indicate a shallow marine depositional environment. Greenish-gray shales are interpreted to have formed in a marine environment, whereas pink and reddish shales probably formed in a nonmarine, coastal plain environment. Fossiliferous limestones formed during relative sea-level highstands without siliclastic influx in a shallow, open- and normal-marine environment. Dominantly muddy textures indicate deposition under predominantly low-energy conditions (Krainer et al., 2003).

The different lithotypes of the Laborcita Member of the Bursum Formation are arranged to form alternating cycles of limestone and siliclastic sedimentary rocks. Otté (1959a) pointed out that these lithologies change abruptly laterally and do not extend over long distances. An ideal cycle begins with a conglomerate and grades upward into alternating sandstone, siltstone and shale, and thin- and thick-bedded fossiliferous limestone, indicating a deepening upward trend (transgressive cycles). Grainstone on top of some limestones, overlain by greenish-gray shale, indicates the beginning of a regressive event, although shallowing upward (regressive) cycles are rare and very thin.

The algal mounds in the upper part of the Laborcita Member are particularly well studied (Otté and Parks, 1963; Cys and Mazullo, 1977; Parks, 1977a; Mazullo and Cys, 1979; Cross and Klosterman, 1981; Bowsher, 1986; Mazullo, 1988). The core facies of the mound complex (aragonite-massive facies) is assumed to have formed in a shallow, low-energy near-shelf edge environment as extensive botryoidal masses of aragonite meadows on the seafloor. The overlying phylloid algal aragonite-cement and phylloid algal wackestone-packstone facies represent

the transitional stage from the inorganic boundstone stage to the organic phylloid algal mound stage. The mound formation was periodically interrupted by subaerial exposure and related formation of fissures and brecciation of the mound facies.

PALEOBOTANY OF THE GOBBLER, BEEMAN, AND HOLDER FORMATIONS

GEOLOGICAL CHARACTERISTICS OF THE FOSSIL PLANT LOCALITIES

As part of this study, fossil plants were collected from the upper part of the Gobbler Formation, from several stratigraphic levels in the Beeman Formation, and from the upper part of the Holder Formation. The collections from the Beeman Formation come entirely from the lower member (Indian Wells Canyon Member). The relative stratigraphic positions of these collections are shown in Figure 47 and in Tables 4 and 5. The geological attributes of each of the collecting sites is described below in stratigraphic order, from the oldest to the youngest.

USNM 43879

Located in the uppermost Gobbler Formation, USNM locality 43879 is from unit 148 of the Space History B stratigraphic section (Figures 12, 48). It is the oldest fossil plant collection in the Alamogordo study area, of late Desmoinesian (Middle Pennsylvanian) age. The collection was made from a brown, micaceous, fine sandstone or coarse siltstone, approximately 1 m below the base of the Gray Mesa Formation. Only a small sample was collected at locality 43879 because the plant material was fragmentary. All appeared to be allochthonous, and comminuted plant debris was abundant and intermixed with the larger plant fragments. Co-occurring animal fossils included ostracods, pectinacean bivalves, and more elongate, asymmetrical types of clams (likely myalinids), which suggests some brackish influence.

USNM 43870

USNM locality 43870 was collected approximately 2 m above the base of the Beeman Formation. It is from unit 14 of the Space History C stratigraphic section (Figures 25, 47, 49). The Beeman Formation base is marked by its sharp contact on limestone at the top of the Gray Mesa Formation. The ~2 m thick interval from the top of the limestone to the plant-bearing strata consists of interbedded gypsiferous layers, micritic limestones, and greenish-gray shales, of variable thicknesses. The plants are found in a 50 cm thick zone of greenish-gray to brown siltstone and interbedded, micritic, silty limestone. The plant-bearing zone is overlain by gypsiferous sediment. Larger plant fragments are concentrated in vuggy, irregularly bedded, coarser-grained, shaley layers that also are rich in muscovite. Finer-grained, shaley layers contain more fragmentary plant remains. The plants are not

FIGURE 47. Stratigraphic distribution of the fossil plant localities. See text for discussion.

TABLE 4. Taxa identified in excavations in the Alamogordo area. Excavations (USNM localities) are organized into 10 stratigraphic levels, from oldest to youngest. Frequency of occurrence (number of quadrats in which a taxon was identified) is given for each taxon in the collection made from each excavation. See text for details. A dash (—) indicates a particular taxon was not present at a particular locality; UN = Unidentified.

Locality information and taxa ^a	USNM locality									
	43430	43838	43869	43868B	43868A	43891	43871	43881	43880	43839
Stratigraphic level	9	8	8	8	8	8	7	6	5	4
Informative quadrats	20	56	25	51	288	498	34	28	34	103
Empty quadrats	9	46	15	46	80	368	7	19	9	96
<i>Sigillaria brardii</i>	—	—	—	—	—	2	—	—	—	2
<i>Lepidophylloides</i>	—	—	—	—	—	1	—	—	—	—
Cf. <i>Lepidostrobohyllum</i>	—	—	—	—	1	3	—	—	—	—
<i>Calamites</i> stems	—	—	5	1	7	9	1	—	—	14
<i>Annularia</i> sp.	—	—	—	—	—	—	—	—	—	—
<i>Annularia spinulosa</i>	—	—	1	4	4	7	—	—	—	—
<i>Annularia</i> cf. <i>carinata</i>	—	—	—	1	1	—	—	—	—	—
<i>Annularia spicata</i>	1	—	—	—	1	2	—	—	—	2
<i>Asterophyllites equisetiformis</i>	—	—	—	—	1	—	—	—	—	—
<i>Calamostachys tuberculata</i>	—	—	—	—	1	6	—	—	—	—
<i>Sphenophyllum</i> sp.	—	—	1	2	2	—	—	1	—	2
<i>Sphenophyllum oblongifolium</i>	—	—	—	—	3	8	—	—	—	1
<i>Sphenophyllum verticillatum</i>	—	—	—	—	2	1	—	—	—	—
<i>Sphenophyllum miravallis</i>	—	—	—	—	—	—	—	—	—	—
<i>Sphenophyllum angustifolium</i>	—	—	—	—	—	—	—	—	—	—
<i>Sphenophyllum</i> strobilus	—	—	—	—	—	—	—	—	—	—
Marattialean foliage	—	—	—	1	—	—	—	3	—	—
Marattialean narrow polymorphopterid	—	—	—	—	19	9	—	—	—	—
Marattialean broad foliage	—	2	1	—	—	12	—	—	—	—
polymorphopterid	—	—	—	—	—	5	—	—	—	—
<i>Pecopteris permica</i> (small pinnules)	—	—	—	—	—	—	—	—	—	—
<i>Diplazites unitus</i>	—	—	—	—	—	—	—	—	—	—
<i>Sphenopteris</i> sp. (small ferns)	—	1	2	—	4	5	1	1	1	—
<i>Sphenopteris biturica</i>	—	1	—	—	—	2	—	—	—	—
<i>Oligocarpia gutbieri</i>	—	—	—	—	—	5	—	—	—	—

(Continued)

TABLE 4. (Continued)

Locality information and taxa ^a	USNM locality																			
	43430	43838	43869	43868B	43868A	43891	43871	43881	43880	43839	43836	43836D	43836C	43836E	43836F	43836A	43834	43870	43879	
<i>Corynepteris angustissima</i>	—	—	—	—	3	4	—	—	—	—	—	—	—	—	—	—	—	—	1	—
<i>Renaultia</i> sp.	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	4	—	—
Pteridosperm UN	—	—	—	—	—	—	4	1	1	—	—	—	—	—	—	1	4	—	3	—
<i>Dicksonites pluckenettii</i>	—	—	—	—	1	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
<i>Neuropteris</i> sp.	—	—	—	—	8	10	—	—	—	9	—	—	10	1	—	—	—	—	—	—
<i>Neuropteris ovata</i>	—	—	—	—	—	—	—	—	—	22	—	—	—	—	—	—	—	—	—	—
<i>Macroneuropteris scheuchzeri</i>	—	9	—	8	34	29	—	4	—	1	—	—	1	—	—	4	4	3	—	—
Neuropterids, high vein density	—	—	1	2	3	2	—	—	—	—	—	—	4	—	—	—	—	—	—	—
<i>Neurodontopteris auriculata</i>	—	2	—	1	—	30	—	—	—	—	—	—	1	—	—	—	—	—	—	—
<i>Alethopteris schneideri</i>	—	—	—	—	1	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—
<i>Reticulopteris</i> sp.	—	—	—	—	—	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—
Cf. <i>Laveineopteris</i>	—	—	—	—	—	5	—	—	—	—	—	—	—	—	—	—	—	—	—	—
<i>Neuropteris</i> / <i>Odontopteris</i> / <i>Cyclopteris</i> sp.	—	—	—	—	1	—	—	—	—	—	—	12	1	1	1	1	—	—	1	—
<i>Odontopteris schlotheimii</i> or <i>cantabrica</i>	—	1	—	—	4	6	—	1	—	—	—	—	1	1	—	1	—	—	—	—
<i>Odontopteris subcrenulata</i>	1	—	—	2	5	—	7	—	—	—	1	—	1	—	1	—	—	—	—	—
<i>Odontopteris lingulata</i>	—	—	—	—	—	2	—	—	—	—	—	—	—	—	—	—	—	—	—	—
<i>Odontopteris brardii</i>	—	—	1	—	4	—	—	—	—	—	—	—	—	—	2	—	—	—	—	—
<i>Odontopteris</i> , small triangular pinnules	—	1	—	—	1	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
<i>Cordaites</i> foliage	11	6	8	14	—	95	11	3	8	35	4	4	12	1	13	20	18	1	—	—
<i>Cordaites</i> foliage, coarse veined	5	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
<i>Cordaitanthus</i> sp.	—	2	—	—	—	2	1	—	—	2	1	—	—	—	2	7	1	—	—	—
<i>Walchia</i> n. sp.	—	12	3	7	26	101	7	1	—	1	—	1	—	—	—	1	—	6	—	—
<i>Walchia piniformis</i>	—	8	1	—	14	27	6	2	2	—	—	4	2	1	1	1	2	1	—	—
<i>Walchia</i> morphotype A	—	—	—	—	—	7	—	—	—	—	—	—	—	—	—	—	—	—	—	—
<i>Walchia</i> sp.	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	—	—	—	—
Cf. <i>Culmitzschia</i>	—	—	—	—	—	2	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Conifer pith cast	—	—	—	—	—	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—
Conifer reproductive organs	—	—	—	1	11	9	1	—	—	1	—	—	—	—	—	—	—	1	—	—
<i>Dicranophyllum</i> sp.	—	—	—	2	3	23	—	—	—	—	—	—	2	—	—	3	—	—	—	—

TABLE 4. (Continued)

Locality information and taxa ^a	USNM locality																			
	43430	43838	43869	43868B	43868A	43891	43871	43881	43880	43839	43836	43836D	43836C	43836E	43836F	43836A	43834	43870	43879	
Cf. Ginkgophyte	—	—	—	—	—	1	—	—	—	—	—	—	—	—	—	—	—	—	—	
<i>Taeniopteris</i> , medium	3	2	1	2	14	18	—	—	—	—	—	—	6	—	—	—	—	—	—	
<i>Taeniopteris</i> , wide veins	—	—	—	—	2	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
<i>Sphenopteris germanica</i>	—	—	—	—	2	—	1	—	1	—	—	9	25	1	—	1	—	—	—	
UN Noeggerathalean foliage	—	—	—	—	—	—	—	1	3	—	—	—	—	—	—	—	—	—	—	
<i>Charliea manzanitana</i>	—	2	—	—	2	10	—	1	—	—	—	—	—	—	—	—	—	—	—	
Noeggerathalean reproductive organs	—	—	—	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	
<i>Autunia naumannii</i>	—	—	—	—	1	—	—	—	—	—	—	—	1	—	—	—	—	—	—	
<i>Dichophyllum flabellifera</i>	—	—	—	—	—	5	—	—	—	—	—	—	—	—	—	—	—	—	—	
UN fringed object	—	—	—	—	—	—	—	—	—	—	—	2	1	—	—	—	—	—	—	
UN segmented objects	—	—	—	—	—	—	—	2	4	—	—	—	—	—	—	—	—	—	—	
Seeds	3	5	—	—	53	58	—	6	9	9	1	2	13	9	3	3	7	2	—	
Cf. pollen organs	—	1	—	—	1	4	—	—	—	—	—	—	—	—	—	—	—	—	—	
Axes	8	8	5	14	58	65	17	14	12	22	1	15	29	7	1	9	3	11	4	
Woody axes	—	—	—	—	19	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Roots	—	—	1	—	3	—	—	—	—	—	—	2	—	—	—	—	—	12	—	
Comminuted plant debris	13	—	15	1	5	14	27	13	22	—	—	30	4	—	—	—	2	—	3	
Charcoal	1	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Conchostracans	—	—	—	—	—	—	—	—	—	2	4	—	4	2	18	29	14	—	—	
Snails	—	—	1	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Ostracods	—	—	—	12	29	129	—	—	—	48	4	—	—	—	4	15	7	—	3	
Spirorbids	—	1	—	2	18	29	—	—	—	—	1	—	5	1	4	5	2	—	—	
Clams	—	—	—	—	—	—	—	—	—	—	1	—	2	—	1	3	1	—	1	
Pectinids	—	8	—	—	—	—	—	—	—	—	—	—	—	—	—	2	—	—	1	
Linguloids	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	—	—	—	
Fish scales	—	—	—	—	—	—	—	—	—	1	—	—	—	1	—	1	3	—	—	
Horseshoe crabs	—	—	—	—	—	—	—	—	—	—	—	—	2	—	—	—	—	—	—	
Insect wings	—	—	—	—	—	2	1	—	—	—	—	—	—	—	—	—	—	—	—	

^a Taxa are grouped according to higher level taxonomic affinity, following standard organization for late Paleozoic floristic papers: pteridophytes first (ordered lycopsids, sphenopsids, marattalean ferns, small ferns), seed plants second (ordered medullosan pteridosperms, cordaitaleans, coniferophytes, ginkgophytes, cycadophytes, noeggerathaleans, callipterid peltasperms), and then various object categories that are not referable to specific taxa, plant parts first, then animal groups.

FIGURE 48. Gobbler Formation, unit 148, Space History B stratigraphic section. USNM locality 43879. (A, B) Two views of the excavation (white arrow) and section.

FIGURE 49. Beeman Formation, unit 14, Space History C stratigraphic section. USNM locality 43870. (A) General view of excavation in context. (B) Close-up view of excavation.

consistently horizontally disposed; many are slightly oblique to bedding or have undulatory surfaces. Plant fragments also occur in the micritic limestone beds within this interval. The entire interval was penetrated by fibrous, woody roots at some point after its deposition, although no evidence of paleosol development was noted. The depositional environment from which this collection was made appears to have been a restricted, possibly lagoonal, standing-water deposit with periodic influx of sediment. No animal fossils were identified in the sediments.

USNM 43834

Located in a roadcut on the north side of La Luz Canyon Road (Figure 50), USNM locality 43834 is at the approximate level of unit 4 of the La Luz C section (Figure 26). Plant remains were found in the lower 2 m of an organic shale bed that rests directly on a sandstone bed. In addition to plants, the shale also contains coprolites, clams, conchostracans, and insect wings. On the basis of the character of the lithologic succession, the fossil

FIGURE 50. Beeman Formation, La Luz C stratigraphic section. USNM locality 43834. (A) Overview of fossiliferous beds in roadcut. (B) Close-up of excavation. (C) Anthracosiid clam fossil, possibly in the genus *Palaeonodonta* Amalisky, occurring in siltstone matrix with plant remains. Scale bar = 1 cm.

content, and, by projection, this locality is at the same stratigraphic level as collections 43836A, B, and F, a more detailed description of which is given below.

USNM 43836

One of the two most intensively collected fossil plant localities in the Alamogordo area, USNM locality 43836 is composed of six separate collections (A–F), plus an uncollected flora from

the lowermost sandstone bed (Figure 51). These beds are in the lower portion of the Beeman Formation (Indian Wells Canyon Member), unit 3 of the La Luz B measured section, ~160 m below the base of the Holder Formation (Figure 26). The plant-bearing beds are exposed in a small, steep arroyo (Figure 51) on the south side of La Luz Canyon. The exposure transitions through about 10 m from a sandstone at the base to a dark, fine-grained, organic-rich clay shale, which coarsens upward to buff siltstone with fine-grained sandstone interbeds. The section

FIGURE 51. Beeman Formation, unit 3, La Luz B stratigraphic section. USNM locality 43836. SS = sandstone at the base of the arroyo exposure. A–F = locations of separate excavations and subcollections.

FIGURE 52. Beeman Formation, unit 3, La Luz B stratigraphic section. USNM locality 43836. Sedimentological features. (A) Ripples in basal sandstone. (B) Mud cracks on parting in basal sandstone. (C) Transition between basal sandstone and overlying organic shale. (D) Laminations in organic shale from excavation 43836C.

is capped by a limestone that contains marine invertebrates. Five separate collections were made from these strata, and floral remains were noted in the basal sandstone and photographed in the field but could not be collected.

Total thickness of the basal sandstone is not determinable on outcrop but is >50 cm (Figures 26, 51). Where bedding surfaces are exposed, thus revealing internal features of the

basal sandstone, the sandstone is thin bedded (<10 cm thick beds), with ripples (Figure 52A) and mud cracks (Figure 52B) visible on some surfaces. The contact with the overlying dark gray shale is transitional through an interval of approximately 15 cm, composed of alternating, thin platy beds of sandstone, with thin interbeds of siltstone and claystone (Figure 52C). The only plant fossils observed in the sandstone

were cordaitalean leaves, many intact, concentrated at the contact between the main sandstone body and the transition zone to the organic shale.

Collections A, B, and F are from the lower 2 m of an organic shale, which lies above the thin transition zone from sandstone. This organic shale is finely laminated at the base, with the laminae thickening upward. Primarily claystone in the lower portion, it coarsens upward slightly. In addition to sparse, allochthonous plant fossils, the organic shale also contains bivalves, microconchids, conchostracans, and insect wings. This shale clearly represents a change from active flow, capable of transporting sand, to quiet, nearly still water, probably of nonbrackish, nonmarine salinities. Both the underlying sandstone and the organic shale may have been deposited under a seasonally dry climatic regime, given the mud cracks in the former and the conchostracans in the latter (which reproduce in ephemeral pools and disperse eggs in windblown dust). This basal part of the deposit may represent a transition from active stream to abandonment and conversion to a lagoon or lake, with periods of quiet standing water (suggested by the presence of microconchids, which attach to floating plant remains).

Collections C and E are from approximately 2 m above the base of the basal sandstone. Somewhat siltier than the lower, darker shale, these beds are also more thickly laminated; laminae are couplets ~50 mm thick composed of a thin, ~1 mm thick, organic-rich band and a thicker, lighter-colored band (Figure 52D). There is no evidence of bioturbation. Allochthonous plant remains, at site E in particular, are concentrated in a thin layer. The remains of xiphosurids (horseshoe crabs) occur in these beds (Lucas et al., 2014b), and late Paleozoic limulids

are generally considered to have lived in freshwater environments (Lamsdell, 2016).

Collection D is from an interval of siltstone with thin sandstone interbeds or laminae. The collection was made ~6 m above the basal sandstone at the base of the exposure. These strata are thin bedded to laminated, as throughout the deposit, but the sediment is overall more coarse grained than below. The beds contain abundant comminuted plant debris; macrofossils are fragmentary and allochthonous. No animal body fossils were found, but horizontal/bedding-plane burrows are present occasionally.

USNM 43839

USNM locality 43839 (Figure 53) is located approximately 23 m above the base of the Beeman Formation in the Space History B section, unit 185, 50 cm above unit 184, a limestone bed (Figure 12C). The stratigraphic interval in which the plant-bearing beds occur is complex on the small scale. Plant fossils (mainly cordaitalean leaves) occur in three different, successive mudrock lithotypes: a brown siltstone; a dark, clay-rich shale; and a hard, micritic mudstone, with a total thickness of 30 cm. Most of the plants are in the finely laminated to thinly bedded, dark shale, which is hard, displaying only weak fractures along horizontal surfaces. Plant material is sparsely distributed in the various lithologies. This mixed-lithology interval lies immediately above a thin, brown sandstone, which is above the gypsiferous interval at the base of the unit. The plant-bearing matrix also contains ostracods, fish scales, bivalves, and insect wings,

FIGURE 53. Beeman Formation, 23 m above the base, 50 cm above unit 184, Space History B section, unit 185. USNM locality 43839. (A) General view of excavation in context. (B) Closer view of excavation.

an assemblage that neither confirms nor rules out occasional brackish-water salinities; conchostracans also are present, the reproductive biology of which suggests periodic drying of parts of the water body.

USNM 43880 and 43881

USNM localities 43880 and 43881 were collected from the same slope, separated vertically by somewhat more than 1 m (Figure 54). They are in the La Luz Canyon B section (Figure 26). USNM 43880 is from the upper part of unit 30; USNM 43881 is from the lower part of unit 31.

USNM 43880 is in a massive, micaceous, brown siltstone to fine sand. Plant remains are allochthonous and associated with abundant comminuted plant debris, which makes up most of the plant remains. The organic debris is distributed through several centimeters of matrix, as if the original sediment were deposited as a slurry.

USNM 43881 is a massive brown siltstone with blocky fracture, no clear lamination, and no muscovite. Comminuted plant debris is common, and intact macrofossil remains all are

allochthonous. Plant remains are somewhat angularly disposed in the matrix, as if deposited rapidly. Thus, although this bed does not have the aspect of an organic-rich slurry, as did the plant-bearing beds of USNM 43880, it nonetheless appears to have been deposited in a pulse-like manner, with the plant remains incorporated in traction load, at least in the final phases of sediment deposition. Both of these deposits appear to represent parts of channel-fill sequences.

USNM 43871

USNM locality 43871 is located stratigraphically approximately 43 m above the base of the Beeman Formation, below the first major Beeman sandstone. It is in the Space History B section, unit 198 (Figure 12C). The plant-bearing beds consist of laminated brown siltstone with fine sandstone laminae or interbeds (Figure 55). The plant material is allochthonous, mostly small debris, with abundant comminuted plant debris on nearly all observed bedding surfaces. One insect wing was found, but no remains of shelly invertebrates and no indications of marine or brackish influence were encountered.

FIGURE 54. Beeman Formation, units 30 and 31, La Luz Canyon B stratigraphic section. USNM localities 43880 and 43881. (A, B) Two views of the general area, taken from different positions. Note large conglomerate boulder in B, displaced from ridgeline.

FIGURE 55. Beeman Formation, unit 198, 43 m above the base of the formation, Space History B stratigraphic section. USNM locality 43871. (A) General view of excavation in context. (B) Closer view of excavation.

USNM 43868 and USNM 43891

In combination, USNM localities 43868, 43869, and 43891 constitute the largest collection made in the Alamogordo area, from the same outcrop in Beeman Canyon (Figures 56, 57). USNM 43868 and 43891 were collected at two separate times from the same place on the outcrop. USNM 43869 was from slightly higher in the section than the other two. All three are in the Beeman Canyon B measured section (Figure 27, unit 13), which we correlate to unit 206 in the Space History B section (Figure 12C). It must be noted that several excavations, in addition to those described below, were made on this outcrop, at different levels, but no identifiable plant remains were recovered from these additional excavations. Comminuted plant debris was found, and continued excavation of this site might, indeed, reveal other beds rich in identifiable macrofossil plant remains.

USNM 43868 was subdivided into two excavations (Figure 57). USNM 43891 was expanded from these initial excavations into three side-by-side excavations. The dominant fossiliferous lithology was a brown to gray siltstone. Beds of this composition were interbedded with thinner beds of more indurated siltstone, possibly cemented by carbonate, and with variably thick, buff sandstone. Some of the sandstone beds pinch out on outcrop, suggesting deposition in small channels. Beds composed of other lithologies also may have been of channel form, but this geometry was not as obvious on outcrop as it was for the more resistant and prominent sandstone beds.

Plant-bearing siltstones are blocky, massive, and not clearly laminated or bedded. The rock fractures irregularly, differentially, and often subhorizontally to the plant-bearing layers. Plant remains are allochthonous and sparse in the matrix; at the outcrop face, we estimated that 95% of shale surfaces excavated

were devoid of plant remains. Comminuted plant debris is rare. The rock contains abundant remains of ostracods; microconchids were observed frequently, attached primarily to cordaitalean foliage but also to *Taeniopteris* specimens, both foliage types characterized by broad laminae that may have floated in standing water or in areas with little turbulence. Rare remains of high-spired gastropods and cockroach wings were observed.

The depositional environment of this deposit, as well as that of USNM 43869, described below, appears to have been a lagoonal or deltaic setting, with variable sediment influx, implying seasonal changes in stream transport capacity. The animal component suggests nonmarine to possibly slightly brackish water conditions or periodic influx of marine waters into a dominantly freshwater setting, with intervals of quiet, relatively still water.

USNM 43869

As noted above, USNM locality 43869 is located on the same outcrop as USNM 43868 and 43891 and immediately adjacent to them (Figures 56, 57). The base of the excavation was immediately above an iron-rich siltstone that also marked the top of USNM 43868. Plant remains occurred in a thin-bedded, brown siltstone. All plant material in the collection is allochthonous and fragmentary, intermixed with abundant comminuted plant debris, very little of which is identifiable.

USNM 43838

USNM locality 43838 (Figure 58) is located on the south flank of Horse Ridge, equivalent to unit 206 of the Space History B section (Figure 12C), and thus is approximately 65 m above the base of the Beeman Formation. Plant macrofossils occur in a buff to brown

FIGURE 56. Beeman Formation, Beeman Canyon B section, unit 13, equivalent to unit 206, Space History B stratigraphic section. USNM localities 43868, 43869. USNM locality 43891 was collected later, from the same location as 43868. (A) General view of exposure down length of arroyo wall, NW-SE axis. (B) Orthogonal view of southwest-facing outcrop.

FIGURE 57. Beeman Formation, Beeman Canyon B section, unit 13, equivalent to unit 206, Space History B stratigraphic section. USNM localities 43868, 43869. (A) Excavations 43836A and 43836B. These were later expanded to a single excavation and given the number 43891. (B) Excavation 43869.

to gray, iron-stained, fine- to medium-grained siltstone, approximately 1 m in thickness. This siltstone is relatively finely laminated (Figure 58C), the laminae consisting of a darker, presumably more organic rich, layer and a somewhat lighter colored layer, indicating rhythmicity and thus possibly indicative of seasonal sedimentation. Plant remains in this deposit were allochthonous and, although fragmentary, were accompanied by little comminuted plant debris. Microconchids are present in small numbers on cordaitalean leaves. The plant-bearing beds are situated immediately below a thin bench of gypsum, part of a sequence of siltstone and interbedded thin layers of powdery gypsum. Above this gypsiferous interval, but at a position we were unable to measure exactly because of the lack of outcrop, was a brown sandstone found as allochthonous blocks, on the surfaces of which cordaitalean leaves were identified, together with less common specimens of woody axes and samaropsid-type winged seeds.

USNM 43430

USNM locality 43430 is located in the upper Holder Formation near the mouth of La Luz Canyon, in a roadside exposure (Figure 59). Plant remains were found in a siltstone. The base of the deposit is covered by talus and vegetation, precluding determination of thickness. The top is marked by a sharp, erosional contact with yellow, massive sandstone, which is at least 3 m thick. The flora is fragmentary and allochthonous. Animal fossils associated with the plant remains include bivalves, leaving open the possibility of deposition in slightly brackish water, consistent with the inferred coastal-plain setting. Above the sandstone is a sequence of marine limestones and shales. The limestones

contain fusulinids (*Triticites*) indicative of a Virgilian age. On the basis of its position in the measured section, this collection is the youngest Pennsylvanian-age flora we were able to find in the Alamogordo area.

DESCRIPTION OF THE FLORA

LYCOPSIDS

SIGILLARIA BRARDII Brongniart. Three specimens of *Sigillaria brardii* and one suspect specimen were identified at two collecting localities in the Beeman Formation, both of Missourian age. In each instance, the specimens were part of large collections, among the three largest of the total Alamogordo assemblage. One specimen is a narrow strip consisting of four leaf bases, three of which are vertically stacked, all widely separated by broad, rugose interareas (Figure 60A). The other two specimens are disorganized masses of detached leaf bases, probably reflecting in situ decay and disassociation (Figure 60B); the leaf bases are much smaller than in the more intact fragment. These specimens are all likely attributable to the subsigillarian group. That of Figure 60A has helically disposed leaf bases, not the distinct vertical rows typical of the eusigillarian group of species. In addition, the leaf bases are distinctly horizontally elongate, with abruptly constricted lateral angles that terminate acutely, and have a notched apex and broadly rounded base and oval, angularly disposed parichnos strands that flank the leaf trace. These are all typical *S. brardii* traits. A third suspect *S. brardii* (Figure 60C) has a leaf-cushion-shaped base, the details of which are obscured by what appears to be an attached leaf

FIGURE 58. Beeman Formation on south flank of Horse Ridge, equivalent to unit 206, Space History B stratigraphic section, USNM locality 43838, approximately 65 m above the formation base. (A) General view of excavation (white arrow) in context. (B) Closer view of excavation. Limestones of Holder Formation crop out in ridge on skyline. (C) Fossiliferous sediment showing lamination.

or possibly a sporophyll. The dimensions of this specimen are greater than the others, but in its general aspect, it appears to be sigillarian.

These occurrences indicate that sigillarian trees were present on the landscape, perhaps in wetter patches along the margins of streams or lagoons, but seemingly exceedingly rare. *Sigillaria brardii* is the main species of this genus known from the Late Pennsylvanian of Euramerica (e.g., Wagner and Álvarez-Vázquez, 2010), although it may, in fact, be a complex of closely related and morphologically similar forms, given its long stratigraphic

range and broad geographic distribution. *Sigillaria brardii* is also reported from the Permian in both Euramerica (Chaney and DiMichele, 2007) and China (Wang et al., 2009a). Other species of *Sigillaria* also survived into the Late Pennsylvanian but appear to have been much rarer floristic elements. For example, Castro (2005) reports and illustrates *Sigillaria rugosa* Brongniart from the middle Stephanian of the La Magdalena coal field in Spain, and Wagner and Álvarez-Vázquez (2010) report several additional species from the Stephanian C, also in Spain. These indisputable occurrences of sigillarians other than *S. brardii* are

FIGURE 59. Upper Holder Formation, La Luz Canyon, roadside exposure. USNM locality 43430. (A) General view of excavation (white arrow) in context. (B) Closer view of excavation.

interesting and in need of further investigation from an ecological perspective. In our collecting in North America, the only species of the genus we have encountered, as of this writing, is *S. brardii*.

LEPIDOPHYLLOIDES Snigirevskaya. Leaves of lycopsids were identified in collections from two localities; the oldest, a small collection in the lowermost Beeman Formation, contains five specimens, and the other in the middle Beeman, part of the largest single collection made in the Alamogordo area, thus the most likely to sample rare elements. The scarcity of such leaves is congruent with the rarity of lycopsid remains, either of the arborescent lycopsids or of smaller-bodied forms. These leaves are narrow, about 2 mm in width, with a single medial vein that is somewhat sunken in the lamina (Figure 60D). All of the specimens collected are fragmentary.

Lycopsid Megasporophylls (Perhaps). Several specimens were found (Figure 61) that may possibly be lycopsid megasporophylls. These objects, triangular in shape and of similar size, are distinctly different in overall form. All have strongly developed mid ridges that may represent a single, central vascular strand. Each also has a small oval scar on the midline, near the base, at the level where the upper and lower lateral margins converge, creating a sharp angle. This scar may represent the point of attachment of a sporangium.

Several specimens, one of which is illustrated, have straight, converging lateral margins, both above and below the point at which the upper and lower lateral margins converge. The tip is narrow and sharply acuminate (Figure 61A). These specimens resemble *Lepidostrobophyllum triangulare* (Zeiller) Bell in both their small size and distinctly triangular shape. A specimen of different morphology (Figure 61B), also triangular, is more shield shaped and has rounded margins, both above and below

the point at which the upper and lower margins converge. The apex is rounded and blunt. This specimen also has several small, circular marks on its surface that may represent arthropod damage. Possibly the most distinctive form, represented by a single specimen (Figure 61C), is also distinctly triangular in shape, with a basal protrusion. However, the lateral margins are curved gracefully and fringed by hairs, and the tip is elongate, drawn out, and acuminate. The final morphotype (Figure 61D) is narrower and more elongate than the others, thus not distinctly triangular in shape. It also has a basal protrusion and curving lateral margins.

The diversity in form of these suspected megasporophylls raises the issue of lycopsid diversity in the Alamogordo collections. At two of the four collecting localities where sporophyll remains were found, there are additional lycopsid remains, either *Sigillaria brardii* or lycopsid leaves. *Sigillaria brardii* occurs at another site, but without any associated reproductive remains, and sporophylls were recovered from two other excavations, but without associated vegetative lycopsid remains. One of these is from USNM 43868A, which is a replicate of USNM 43891, from which both lycopsid vegetative and reproductive remains were found. The other, a pair of *Lepidostrobophyllum triangulare*-like sporophylls found on the same hand specimen (not illustrated), is from USNM 43836A. This is one of seven separate collections from the lower Beeman organic shale bed, none of the others of which had any lycopsid remains. Sporophylls of this type are not typical of *Sigillaria strobili*, suggesting other lycopsids on the landscape, perhaps small isoetalean forms, the vegetative remains of which either were not transported into the depositional environments or were not recognized as such by us.

FIGURE 60. Lycopsid vegetative organs. (A) *Sigillaria brardii* stem fragment. USNM specimen 706336, USNM locality 43891. (B) *Sigillaria brardii* detached leaf cushions. USNM specimen 713224, USNM locality 43839. (C) Likely *Sigillaria brardii* with attached leaf base or sporophyll. USNM specimen 713066, USNM locality 43891. (D) Lycopsid leaf. USNM specimen 713212, USNM locality 43870. Scale bars = 1 cm.

SPHENOPSIDS

Arborescent, calamitalean sphenopsids are represented in 12 of 19 Alamogordo area collections. *Sphenophyllum* Brongniart, plants with a scrambling, climbing, or thicket-forming growth habit, although less abundant as measured by frequency of quadrat occurrences, nonetheless, has the same overall frequency as the calamitaleans, occurring at 12 of the 19 sampling locations. The distributions of the two groups strongly overlap. Both are typical parts of Pennsylvanian wetland assemblages.

Calamitalean Stems. The most common sphenopsid remains are calamitalean stems (Figure 62A–E), present at 9 of the 19 collecting sites. Their average frequency is 8.24%. Most of these stems can be assigned either to *Calamites cistii* Brongniart or *C. suckowii* Brongniart. In each, the stem consists of ribs that alternate at nodes, which are not swollen. In the former species, there are no scars at the nodes (Figure 62A,C). In the

latter, small oval scars occur just below the apex of a rib and are generally present across the entire node (e.g., Figure 62B). Many specimens are ambiguous. The species-level classification is almost certainly artificial. In some instances, nodal plates are preserved in transverse view (Figure 62D,E), rarely with leaves still in attachment (Figure 62D).

Calamitalean stem remains are commonly referred to as “pith casts.” DiMichele and Falcon-Lang (2012) challenged this interpretation and argued that these were the remains of stems whose thin outer primary tissues decayed away. Data supporting this interpretation are presented in that paper and were further developed by Falcon-Lang (2015).

ASTEROPHYLLITES Brongniart. A single, questionable specimen of *Asterophyllites* (Figure 62F) was found at USNM locality 43868A. This is the second-largest collection in terms of the number of specimens made in the Alamogordo area and thus more likely to capture rare material than smaller collections. The

FIGURE 61. Cf. *Lepidostrobophyllum*. (A) Cf. *L. triangulare*. USNM specimen 713213, USNM locality 43870. (B) USNM specimen 706320, USNM locality 43891. (C) USNM specimen 706326, USNM locality 43891. (D) USNM specimen 706325, USNM locality 43891. Scale bars = 1 cm.

FIGURE 62. Calamitalean remains. (A) *Calamites cistii*. USNM specimen 713098, USNM locality 43891. (B) *Calamites suckowii*. USNM specimen 713216, USNM locality 43839. (C) *Calamites cistii*. USNM specimen 713215, USNM locality 43839. (D) Calamitalean nodal plate with foliage in attachment. USNM specimen 699092, USNM locality 43868B. (E) Calamitalean nodal plate. USNM specimen 713214cp, USNM locality 43839. (F) *Asterophyllites* sp. USNM specimen 699124, USNM locality 43868A. Scale bars = 1 cm.

genus is characterized by narrowly linear, upswept leaves, with acuminate tips and single veins, clustered at axis nodes. The rarity of this plant in the Alamogordo collections is mirrored in equivalent-age strata from Socorro County to the north (DiMichele et al., 2017a), which have been collected much more extensively. In this pattern it differs significantly from deposits in north central Texas, where *Asterophyllites equisetiformis* (Schlotheim ex Sternberg) Brongniart was the most common form of calamitalean foliage in all plant-bearing, terrestrial, wetland facies, but

particularly those formed in swamps (Tabor et al., 2013). Thus, as in Socorro County, New Mexico (DiMichele et al., 2017a), we speculate that its near absence in the Alamogordo collections may reflect a slightly more arid climate in western reaches of Pangea during the Late Pennsylvanian.

ANNULARIA Sternberg. Three principal forms of the calamitalean foliage *Annularia* were identified in the Alamogordo collections (Figure 63). These occurred at 8 of the 19 collection sites. Frequency, where the genus was present, averaged 2.82% of

FIGURE 63. *Annularia* foliage. (A) *A. spinulosa*. USNM specimen 699138, USNM locality 43868A. (B) *A. spinulosa*. USNM specimen 713099, USNM locality 43891. (C, D) *A. cf. mucronata*. USNM specimen 699090, USNM locality 43868B. (E) *A. cf. mucronata*. USNM specimen 706295, USNM locality 43868A. (F–H) *A. spicata*. USNM specimen 713202, USNM locality 43870. (I) *A. spicata*. USNM specimen 713227, USNM locality 43839. (J) *A. spicata*. USNM specimen 713112, USNM locality 43836A. Scale bar in E also applies to H–J. Scale bars = 1 cm.

the sample quadrats. Reproductive organs, which we have attributed to *Calamostachys tuberculata* (Sternberg) Weiss, were found in two excavations, at which one of the more common forms of *Annularia* foliage, *A. spinulosa* Sternberg, was also found.

The most commonly encountered calamitalean foliage in the collections can be attributed to *Annularia spicata* (Gutbier) Schimper. Leaves in this species are small; entire whorls can be <5 mm in diameter. Leaves are fusiform in shape with acute apices. Whorls are symmetrical and round in shape (Figure 63F–J).

The second most commonly encountered calamitalean foliage in the Alamogordo collections is attributed here to *Annularia spinulosa* Sternberg. Leaves of this species are narrow and

fusiform in shape, with tapering acuminate apices and single veins (Figure 63A,B), attaining lengths >20 mm and widths of ~2 mm. Leaf whorls of the Alamogordo specimens contain approximately 20 leaves, which is slightly fewer than characteristic of the species. In leaf and whorl shape, however, they conform fairly well to the original description, which has been discussed by Barthel (2004). Specimens of this form have long been attributed to *A. stellata*, which Barthel (2004) argues should be replaced by *A. spinulosa*, a taxonomic determination followed here.

Two specimens illustrated in Figure 63C–E have leaves that are more spatulate in form, with bluntly rounded, mucronate apices. In the larger of the two (Figure 63C,D), the whorl,

although incomplete, appears to be radially symmetrical, which, in combination with the other characteristics, is typical of *Annullaria mucronata* Schenk. In the smaller (Figure 63E), the leaves are of different lengths, and the whorl is asymmetrical. Preservation of the small specimen is considerably incomplete, and it appears to have been slightly distorted. Thus, it simply may be a smaller specimen of *A. mucronata*, or it may be attributable to something else, such as *A. carinata* (Gutbier) Schimper.

CALAMOSTACHYS Weiss. Several specimens of a calamitalean strobilus similar to *Calamostachys tuberculata* (Sternberg) Weiss occur at two collecting sites, from which the largest of the Alamogordo collections was made (USNM localities 43898A and 43891). Both are from the same beds in Beeman Canyon, in the upper part of the lower Beeman Formation. The strobili (Figure 64) are large, with a wide central axis, 5 mm or more in width, bearing long bracts and sporangiophores in alternating whorls. The bracts are exserted, extending several millimeters beyond the sporangiophores. Sporangia are elongate and numerous, projecting back toward the stem, as is typical for calamitaleans. All of

these features are visible in USNM specimen 713094, the part and counterpart of which are illustrated in Figure 64B,C.

SPHENOPHYLLUM Brongniart. *Sphenophyllum* is a sphenopsid reconstructed as having various growth habits, varying from sprawling ground cover, to shrub or thicket former, to climber (e.g., Batenburg, 1982; Bashforth and Zedrow, 2007). Analyses of its vascular system (Wilson, 2013) indicate that it had very high hydraulic conductivity due to the large size of its tracheids, with a water-flow capacity equivalent to that of some angiosperms; among the consequences of this is reliance on almost permanently wet substrates due to the risk of cavitation should the water supply be interrupted. The plant is a common element of Pennsylvanian wetland floras but is poorly represented in New Mexico fossil floras equivalent in age to the Alamogordo assemblages. For example, it is absent, as far as is presently known, from the large Kinney Brick Company Quarry flora (e.g., Mamay and Mapes, 1992; DiMichele et al., 2013) and is reported as rare in the many collections from Socorro County (DiMichele et al., 2017a). In contrast to these other areas, *Sphenophyllum*

FIGURE 64. *Calamostachys tuberculata*. (A) USNM specimen 706299, USNM locality 43868A. (B, C) USNM specimen 713094, USNM locality 43891, part and counterpart. Scale bar in B = 1 cm and applies to all images.

remains were relatively common in the Alamogordo collections, considering its growth habit and ecological preferences, where it was found in 13 of the 19 excavations. Considering only the collections in which it occurs, *Sphenophyllum* vegetative or fertile remains occur at a frequency of 7.92%.

Other than specimens not identifiable below the generic level, the most readily identifiable, abundant, and widespread *Sphenophyllum* remains are those of *S. oblongifolium* (Germar and Kalfuss) Unger. This species is considered diagnostic of the Cantabrian stage or younger (Wagner, 1984), which is consistent with the age estimate of the Beeman Formation based on conodonts. The Alamogordo specimens (Figure 65A,B) are typical, consisting of six leaves per whorl, two each forming opposite pairs and two shorter leaves forming a bib. Leaves have an elongate, triangular shape with straight lateral margins and a toothed, sometimes deeply incised distal margin. Open-dichotomous veins branch multiple times, and all reach the terminal margin.

The second most abundant *Sphenophyllum* species, on average, is a form we attribute to *S. miravallis* Vetter emend Boersma (Hettterscheid and Batenburg, 1984). The taxonomy of this plant is complicated, but we believe that *S. miravallis* is the correct name for specimens described elsewhere as *S. miravallense*, *S. crenulatum* Knight, *S. crenulatum* (Maithy) Srivastava and Rigby, and *Parasphenophyllum crenulatum* (Knight) Castro (non *Parasphenophyllum crenulatum* Maithy). DiMichele et al. (2017a) mistakenly identified *S. miravallis* specimens as *S. verticillatum* (Schlotheim) Zeiller. As discussed by Hettterscheid and Batenburg (1984), the plant has a high degree of variability in leaf shape and characteristics. In the form most easily recognized, leaves in a whorl are of equal length, and whorls are round in shape (Figure 65C is a specimen possibly attributable to *S. miravallis* that still has leaves attached). Individual leaves (Figure 65D,E) are wedged shaped with straight lateral margins and a broadly rounded distal margin. The distal margin is fringed by small teeth that may have acuminate extensions (Figure 65E) in which one or more veins terminate. Veins curve outward and terminate in the rounded portions of the distal margin.

The final form identified with reasonable certainty is *Sphenophyllum verticillatum* (Schlotheim) Zeiller (Figure 65I,J), a species that is very similar in morphology to *S. emarginatum* (Brongniart) Brongniart. Leaves are generally found isolated, although the specimen illustrated in Figure 65I may represent a complete leaf whorl of *S. verticillatum*. The species is characterized by wedge-shaped leaves with straight lateral margins and a flat distal margin. Veins all terminate at the distal margin. The species is characteristic of Late Pennsylvanian (Stephanian) floras in western Europe (Remy and Remy, 1959; Wagner and Álvarez-Vázquez, 2010).

A single specimen was found that is possibly attributable to *Sphenophyllum angustifolium* (Germar) Göppert, bearing lacinate leaves at right angles to the stem. Two slender axis specimens possibly attributable to *Sphenophyllum* are illustrated in Figure 65G,H. Another specimen, similar to that in Figure 65G, is illustrated in Figure 65F. Both of these specimens have leaves

or bracts attached at the closely spaced nodes. However, close examination of Figure 65F suggests that some of these are leaves that may have rounded leaf apices, as found in *S. miravallis*.

FERNS

Marattialean Fern Foliage. We recognize six groups of marattialean fern foliage, five tentatively assigned to genera/species. These groups are (1) *Polymorphopteris* sp. sterile foliage, (2) cf. *Danaeites* sp. sterile and fertile foliage, (3) *Diplazites unitus* sterile foliage, (4) specimens of sterile foliage with once-forked ultimate veins, (5) specimens of sterile foliage with very small pinnules and widely once-forked ultimate veins, comparable to "*Pecopteris*" *jongmansii*, and (6) undifferentiated sterile marattialean foliage consisting primarily of fragmentary or poorly preserved specimens. We also recognize specimens of fertile foliage that are assignable broadly to *Asterotheca* but for which preservation is too poor to be confident in that assignment (given the variety of marattialean foliage with circular synangia).

The cautiousness in making these assignments reflects, most critically, the fragmentary nature and generally small size of the specimens, making it difficult to discern many of the crucial diagnostic characteristics. Second, few specimens were found for most of these forms, making it difficult to assess the range of variability in the sample and thus to see the full expression of many characteristics. Last, even when dealing with well-preserved, large specimens, there are challenges in the identification of marattialean fern foliage as meaningful biological species, particularly when sterile. Cleal (2015) recommends abandonment of the widely used name *Pecopteris* for sterile foliage and urges the use of reproductive organs as the main guide to generic taxonomy; Mosbrugger (1983) used this same approach to reassign a number of *Pecopteris* species, although many practitioners still rely on older usage, particularly where identifications are tentative and affinities with reproductive organs uncertain. The need for names for sterile foliage is addressed further by Pšenička et al. (2009) and Wittry et al. (2015). Large numbers of species have been described on the basis of only vegetative characteristics (e.g., Boureau and Doubinger, 1975), many representing different parts of fronds of the same natural species, reflecting complex morphological changes within the fronds, depending on location and stage of development. This variability is compounded by what appears to have been high natural diversity in the group, which expanded ecologically during the Late Pennsylvanian, in particular, many species co-occurring locally and regionally (e.g., Wagner, 1985; Popa and Pšenička, 2010).

Among the most commonly found, identifiable forms of marattialean sterile foliage in the Alamogordo collections are specimens of cf. *Polymorphopteris* Wagner, identified on the basis of venation and pinnule shape (Figure 66). Specimens of this type were found in three collections from two collecting horizons. These specimens have narrow pinnules, 2–4 times longer than wide, with straight sides and bluntly rounded tips; in some specimens, the pinnules may taper toward the apex (e.g.,

FIGURE 65. *Sphenophyllum*. (A) *S. oblongifolium*. USNM specimen 706293, USNM locality 43868A. (B) *S. oblongifolium*. USNM specimen 713095, USNM locality 43891. (C) *S. miravallis*. USNM specimen 713161, USNM locality 43836. (D) *S. miravallis*. USNM specimen 713155, USNM locality 43836D. (E) *S. miravallis*. USNM specimen 713156, USNM locality 43836D. (F) Foliated axis. USNM specimen 713159cp, USNM locality 43836. (G) Axis, stem or strobilus. USNM specimen 713167, USNM locality 43836F. (H) Axis, stem. USNM specimen 713188, USNM locality 43836F. (I) *S. verticillatum*. USNM specimen 713097, USNM locality 43891. (J) *S. verticillatum*. USNM specimen 713205, USNM locality 43870. Scale bars = 1 cm. Scale bar in D also applies to E.

Figure 66A) or have slightly crenulate margins (Figure 66C). Pinnules are not confluent, have a slightly inclined insertion on the rachis (where preserved in attachment), and are broadly attached across their bases. Midveins have a straight to slightly

inclined insertion, are prominent, and traverse at least 90% of the pinnule length. Lateral veins fork at their point of attachment to the midvein, and in most specimens, both derived veins fork again prior to the midpoint of the lamina, between the

FIGURE 66. Cf. *Polymorphopteris* foliage. (A) USNM specimen 713104, USNM locality 43891. (B) USNM specimen 713106, USNM locality 43891. (C) USNM specimen 706300, USNM locality 43868A. (D) USNM specimen 706306cp, USNM locality 43868A. (E) USNM specimen 713202, in dashed box, USNM locality 43870. Scale bars = 1 cm.

midvein and margin, although this may vary; in some pinnules, some of the lateral veins may remain unforked, or the position of the second division may lie beyond the lamina midpoint. Ultimate pinnae taper gradually to a small, rounded apical pinnule (Figure 66A), and there may be a slight pinnule confluence near the pinna apex. Pinnules of this form are most often found as

isolated, poorly preserved specimens (e.g., Figure 66E), recognizable because of their distinctive venation. The genus *Polymorphopteris* has a long stratigraphic range, extending from the late Middle Pennsylvanian through the entire Late Pennsylvanian. The type species, *P. polymorpha* Wagner, extends through this entire range, and several additional species appear in the

Late Pennsylvanian, described from Europe (e.g., Wagner and Álvarez-Vázquez, 2010). Foliage of this type is the most commonly encountered form of marattialean foliage in the floras we have investigated from Missourian-age strata in New Mexico (e.g., DiMichele et al., 2017a, 2017b).

Initially, we identified this material, which appears to be of one common species, as *P. polymorpha* but later decided it was best to leave it undetermined below the generic level. It also must be cautioned that the reproductive organs associated with this genus are not known and that Krings and Schultka (2010) described seeds in attachment to foliage with polymorphopterid

venation (although not the same in detail as the material described herein or elsewhere in New Mexico floras).

A second, relatively common form of marattialean foliage in the Alamogordo collections is tentatively assigned to *Danaeites* Goeppert based on pinnule shape and venation and association with reproductive organs of distinctive morphology (Figure 67). Only pinna fragments were found, so the pinna architecture of these specimens is unknown. The sterile and fertile pinnules are the broadest of any marattialean foliage found in the Alamogordo collection and are generally <2 times longer than wide. Pinnules are broadly attached to the rachis, and the fertile specimens may be

FIGURE 67. Cf. *Danaeites* vegetative (A–D) and fertile (E–G) foliage. (A) USNM specimen 713107cp, USNM locality 43891. (B, C) USNM specimen 713107p, USNM locality 43891. (D) USNM specimen 699094, USNM locality 43868B. (E) USNM specimen 713101, USNM locality 43891. (F) USNM specimen 713197, USNM locality 43838. (F) USNM specimen 706303, USNM locality 43868A. Scale bars = 1 cm.

slightly contracted at the base. Pinnule sides are slightly rounded and thus subparallel, terminated by a broadly rounded apex. The midvein is prominent and extends through most of the pinnule length, breaking up below the apex. Lateral veins are difficult to observe in most sterile specimens. A part and counterpart of the best-preserved specimen are illustrated in Figure 67A,B; at the highest magnification (Figure 67C), it can be seen that the lateral veins fork widely immediately at their point of insertion to the midvein and then fork again some distance from that point of insertion on both limbs and somewhat asymmetrically. This is not the typical pattern for the common American species *Danaeites emersonii* Lesquereux, in which the lateral veins are generally closely spaced, low angled, and very straight in their path to the margin. Fertile specimens (Figure 67E–G) have distinctly ovoid to horizontally elongate synangia, known from the Pennsylvanian only in *Danaeites*, linking this form to modern groups of the Marattiales. This characteristic is best shown in the specimen illustrated in Figure 67E.

Danaeites has a long stratigraphic range, extending from the late Middle Pennsylvanian through the late Permian (Liu et al., 2001; Wagner and Álvarez-Vázquez, 2010; Blake and Gillespie, 2011). Blake and Gillespie (2011) noted the abundance of this genus in the Virgilian (late Late Pennsylvanian/Stephanian C) and early Wolfcampian (Asselian) of the Appalachians, a pattern of occurrence that led Read and Mamay (1964) to name their uppermost Pennsylvanian floral zone (number 12) for *Danaeites*. In a review of the distribution of later Pennsylvanian floras in the Appalachian region, Wagner and Lyons (1997) review, recount, and reinterpret many occurrences of *D. emersonii*, which often occurs in association with *Polymorphopteris polymorpha*, among other plants characteristic of moisture-stressed microhabitats, such as conifers and *Taeniopteris*, as part of mixed assemblages composed of wetland and seasonally dry elements. Regionally, Mamay and Mapes (1992) illustrate fertile material with typical *D. emersonii* syngial morphology from the Missourian-age Kinney Quarry, New Mexico.

Among additional, possibly identifiable, forms of marat-tialean foliage is a relatively common form that is typified by elongate, straight-sided pinnules with blunt apices and inclined insertions on the rachis (Figure 68A,B). Pinnule midveins are strong and extend to near the pinnule apex; in larger pinnules, all lateral veins are once forked, close to the midvein. In rarely occurring pinnae with preserved apices, pinnules decrease sharply in size near the apex, and the last two or three are confluent; the apical pinnule is small and rounded. These specimens are similar to “*Pecopteris*” *cyathea* Brongniart (moved by Mosbrugger [1983] on the basis of association with a distinct form of marat-tialean reproductive structure to *Cyathocarpus cyatheus* (Brongniart) Mosbrugger, a decision mirrored in the above-referenced paper by Cleal [2015]). The relatively strong inclination of the pinnules is somewhat at odds with this attribution.

Another morphotype, represented by only a single, fragmentary specimen, has attributes suggestive of the species “*Pecopteris*” *jongmansii* Wagner or *P. arborescens* (Schlotheim) Brongniart (as with *P. cyathea*, this latter species was moved by Mosbrugger [1983]

to *Cyathocarpus* Weiss as *C. arborescens* (Brongniart) Weiss). Pinnules are very small and have an orthogonal insertion on the rachis, straight sides, and a bluntly rounded apex (Figure 68C,D). Venation is very difficult to see because of poor preservation; however, when viewed at high magnification (Figure 68D), the midvein extends to near the apex, and lateral veins appear to be simple and unforked, except near the pinnule base, where the first basiscopic vein may be once forked. This specimen bears comparison with *P. jongmansii* of Wagner (see more complete description in Wagner, 1985), who describes the pinnules as small, closely spaced, and orthogonally inserted, with straight midveins and mostly unforked ultimate veins. Forking of the lateral veins would not be in conformance with the understanding of *P. arborescens*, although the small pinnule size is appropriate, and the species is known from Missourian-age strata (Wagner and Lyons, 1997; Wagner and Álvarez-Vázquez, 2010); Gillespie and Pfefferkorn (1979), however, indicate restriction to the Virgilian in the Appalachians.

A final, tentatively identified fragment of vegetative marat-tialean foliage is *Diplazites unitus* (Brongniart) Cleal. The single specimen is a small fragment (Figure 68E). Pinnules are fused laterally at the base but are free through most of their length. The most distinctive feature is the strongly upswept, fasciculate venation. This venation is not of the stereotypical form found in lobatopterid marat-tialean ferns, in which the lateral veins branch alternately in a candelabra-like fashion (Wagner, 1958), although there are other forms of lobate pecopterids with various forms of fasciculate lateral venation (see discussion in Pšenička et al., 2009).

Specimens with reproductive organs intact were sometimes encountered, although not as commonly as might be expected, given the significant reproductive capacity of marat-tialean tree ferns. The details of these reproductive organs were insufficient to permit assignment, other than to note that the synangia tended to be round in shape, rather than elongate. An example of such a specimen is illustrated in Figure 68F. On the basis of pinnule shape and insertion, this specimen may belong to the cyathea group of marat-tialean ferns.

Small Ferns. Small ferns, although never abundant, are diverse and were found at 10 of the 19 collecting sites. For the most part, these are identifiable only to the level of the genus, *Sphenopteris*, which encompasses more than ferns but is useful to characterize small, generally lobate types of foliage. A few specimens were identifiable with some confidence to species that are widespread throughout Euramerica.

Among the forms of suspected known identity are a few specimens similar to *Oligocarpia gutbieri* Göppert (Figure 69A–C). These specimens occur at two localities, both in the Beeman Formation. The pinna fragments have elongate pinnules with a slightly angular insertion. Rachises are smooth, with slight longitudinal striation. Pinnule midrib insertions vary from slightly decurrent to orthogonal to the rachis; midribs are thin and slightly flexuous. Lateral veins are sparse and thin, fork at least once at a variable distance from the midrib, and may fork a second time. These veins also are somewhat flexuous. The pinnule laminae appear to be thin on the basis of the ghostlike

FIGURE 68. Miscellaneous marattialean foliage. (A) “*Pecopteris*” (*Cyathocarpus*) *cyathea*. USNM specimen 713166, USNM locality 43836F. (B) “*Pecopteris*” (*Cyathocarpus*) *cyathea*. USNM specimen 713206p, USNM locality 43870. (C, D) “*Pecopteris*” *jongmansii*. USNM specimen 713102, USNM locality 43891. (E) *Diplazites unitus*. USNM specimen 713185, USNM locality 43834. (F) Fertile marattialean foliage. USNM specimen 713103, USNM locality 43891. Scale bars = 1 cm.

appearance on the rock surface. This species is relatively common in the Late Pennsylvanian and is illustrated in numerous floras (e.g., Wagner, 1985; Castro, 2005).

A form of *Sphenopteris* recovered in two excavations has small, inclined, triangular, basally confluent pinnules that become progressively more fused toward the apex of the

penultimate rachis (Figure 69D,E). Pinnule margins may be wavy to slightly toothed. Although only one vein enters the pinnule base, a strongly developed midvein is not present; ultimate veins are strongly acroscopically directed and fork several times in the larger pinnules. We attribute these specimens to *S. biturica* Zeiller (see extensive illustrations in Wagner, 1985).

FIGURE 69. Small ferns I. (A) *Oligocarpia* cf. *gutbieri*. USNM specimen 713207p, USNM locality 43870. (B) *Oligocarpia* cf. *gutbieri*. USNM specimen 713089, USNM locality 43891. Brackets indicate pinnules enlarged in C. (C) Pinnules from B enlarged to show shape and venation. (D) *Sphenopteris* cf. *biturica*. USNM specimen 713087cp, USNM locality 43891. (E) *Sphenopteris* cf. *biturica*. USNM specimen 713087p, USNM locality 43891. Scale bars = 1 cm.

They also are somewhat similar to *S. mathetii* Zeiller; however, in this latter species, the pinnules are somewhat more rounded or lobed, although very similar in general aspect and with similar venation. The two species have been combined by some authors.

A third form to which we feel a name can be attached is very similar but perhaps not identical to the zygopterid fern *Corynepteris* cf. *angustissima* (Sternberg) Němejc (Figure 70A,B; Brousmiche, 1983). Although mainly a late Middle Pennsylvanian form and widespread in wetlands (e.g., Opluštil et al., 2007; Tenchov, 2010), it also occurs in the Stephanian/Late Pennsylvanian. The Alamogordo specimens were found in three separate excavations from two stratigraphic levels. All are very similar and quite distinct from all kinds of small-fern foliage. On the ultimate pinnae, small pinnules are arrayed in very close proximity. In some specimens, these pinnules appear to be fused into a continuous lamina (Figure 70A), whereas in others, the pinnules are somewhat to completely free, particularly at the base of the pinna (Figure 70B). In the latter case, the pinnules are slightly inclined acroscopically and have straight lateral margins and a toothed, nearly straight, apical margin. Venation is quite sparse, consisting of a midvein that is inclined, but only slightly decurrent, which proceeds about 2/3 of the way through the lamina and forks below the apical margin at least once. The pinnae themselves are rather widely spaced on the penultimate rachis, so that laminae from adjacent pinnae do not touch.

The final possibly identifiable small fern foliage is referred to the genus *Renaultia* Zeiller (Figure 70C). This tiny scrap could, of course, be derived from any number of small ferns. It bears some similarity to forms of *Renaultia* as illustrated and described by Brousmiche (1983), although that study is focused on stratigraphically older species than in the Alamogordo collections. Several small scraps of this morphotype were found at one locality, the lowest in the Beeman Formation. The specimens consist of minute pinnules of a rounded aspect, with slight marginal undulations. Pinnules lack a clear midrib; the thin veins are initiated at a high angle and slowly diverge from the rachis before sweeping up to the margin. The apical pinnule is simply a small bump below which the lateral pinnules fuse to one another.

The remainder of the small ferns were classified simply as *Sphenopteris* Sternberg. These consist mainly of small bits of foliage of various morphologies, from those with rounded pinnules and sparse upswept, open dichotomous venation (Figure 70D–F) to laciniate forms (Figure 70G). These various morphotypes of small fern scraps occurred at 10 of the 19 collection sites, at an average frequency of 4.7% where they occur, overall at about 2.5% (Table 4). In either instance, this is a relatively high level of occurrence given the unlikelihood of small fern preservation (Scheihing, 1980)—such plants rarely abscise foliage and often live in microhabitats sheltered from wind (damage and transport), greatly reducing the probability that they will be naturally sampled. Thus, the frequency of small ferns seen

in the Alamogordo collections, combined with the abovementioned relative frequency of *Sphenophyllum* (compared to other Pennsylvanian-age deposits from New Mexico thus far examined), suggests periodic development of a luxuriant ground cover of smaller plants, further suggesting an adequate, if not relatively high, volume of freshwater input, likely via rainfall.

Fertile specimens of small ferns were rarely encountered. Two specimens suspected to be fertile are illustrated in Figure 70H and 70I. That in Figure 70H has 5–6 sporangia (the exact number is difficult to determine) clustered in circular arrays at the tips of pinnules. The specimen illustrated in Figure 70I appears to have sporangia located in clusters, possibly on the underside (abaxial side) of pinnules, perhaps paralleling veins in their orientation; this is difficult to discern, and it could be argued that this is not even a fertile specimen. In each of these instances, the pinnule laminae appear to have been thin, delicate in construction, and thus preserved only as reduction-halo “ghosts” on the rock surface.

NOEGGERATHIALES

The Noeggerathiales are an enigmatic group of heterosporous pteridophytes of uncertain, higher-order taxonomic relationships. A recent study of anatomically preserved forms concluded that their closest affinities appear to be with some of the basal groups of ferns or with progymnosperms (Wang et al., 2017). In tropical regions of Pangea, the group is known from strata of both Carboniferous and Permian age (Pfefferkorn and Wang, 2016). In general, noeggerathialeans are found rarely in Euramerican Carboniferous floras (e.g., Šimůnek and Bek, 2003; Wagner and Álvarez-Vázquez, 2010), leading to the inference that the group was centered in “uplands” (e.g., Cridland and Morris, 1963) or “extrabasinal lowlands” (e.g., Pfefferkorn, 1980), areas where preservation potential approached nil (Leary and Pfefferkorn, 1977). This inference therefore requires the transport of noeggerathialean remains to basinal wetlands in order to effect preservation; comment on the inadequacy of the “upland/extrabasinal” model to explain the fossil record has been made elsewhere (DiMichele, 2014). Undoubtedly, remains of plants that grew in seasonally dry habitats sometimes were transported into basinal settings, but where well-preserved remains are present, it is almost without question that the plants were growing in the basins, proximate to depositional environments, at times when climatic conditions suited them. Pfefferkorn and Wang (2016) note that during the Permian, noeggerathialeans became more common in basinal habitats, including seasonally dry settings in Euramerica (e.g., Mamay, 1968) and wetland habitats in Cathaysia (Wang et al., 2009b, 2014).

CHARLIEA MANZANITANA Mamay. *Charliea* Mamay is a multispecies genus that ranges throughout the Pennsylvanian in floras dominated by xeromorphic elements. There was some debate about the phylogenetic affinities of *Charliea*, but it was settled once Tidwell and Ash (2003) demonstrated that the laminate foliar elements were leaves on a plagiotropic

FIGURE 70. Small ferns II. (A) *Corynepteris* cf. *angustissima*. USNM specimen 713098, USNM locality 43870. (B) *Corynepteris* cf. *angustissima*. USNM specimen 713090, USNM locality 43891. (C) Cf. *Renaultia* sp. USNM specimen 713211, USNM locality 43870. (D) *Sphenopteris* sp. USNM specimen 713085, USNM locality 43891. (E) *Sphenopteris* sp. USNM specimen 713086, USNM locality 43891. (F) *Sphenopteris* sp. USNM specimen 706305, USNM locality 43868A. (G) *Sphenopteris* sp. Field photograph. USNM locality 43834. (H) Fertile *Sphenopteris* sp. Arrows indicate the position of the two best-preserved sporangial clusters. USNM specimen 706304, USNM locality 43868A. (I) Fertile *Sphenopteris* sp. Arrow indicates a suspect sporangium or cluster of sporangia. Others of similar form can be seen elsewhere on the specimen. USNM specimen 713088, USNM locality 43891. Scale bar in E also applies to C–G. Scale bar in I also applies to H. Scale bars = 1 cm.

branching system, not leaflets of a compound, frond-like leaf. This characteristic links *Charliea* to other species of the Noeggerathiales with similar branch architecture and basic leaf organization.

Charliea manzanitana Mamay was first described from the Kinney Brick Company Quarry of New Mexico (Mamay, 1990), which, like most of the deposits described here, is of Missourian age (Lucas et al., 2011). Leaves of *C. manzanitana* (Figure 71A–D) are broad but elongate, 2 to 3 times longer than wide. They are strongly constricted at the base and have lateral margins that curve outward in the lower 1/3 of the leaf and then are approximately parallel, flaring outward somewhat in the

upper 1/4 to 1/3 of the leaf, where the lamina divides into 3, rarely more, teeth (Figure 71B,C), which might also be described as elongate, sharply pointed lobes. It is this division into acuminate pointed, terminal lobes that permits *Charliea* to be separated with certainty from other noeggerathialeans. The typical *C. manzanitana* leaf is approximately 2–3 cm in length. The leaves lack a midvein; several veins enter the base and undergo one to several dichotomies in the lower 20% of the leaf, after which vein dichotomies are rare to absent. The veins are widely spaced, parallel, and terminate in the lobes (Figure 71B,C). Isolated leaves are the most commonly encountered remains of the plant (Figure 71D).

FIGURE 71. *Charliea manzanitana*. (A) USNM specimen 713064, USNM locality 43891. (B) USNM specimen 713063, USNM locality 43891. (C) USNM specimen 699120, USNM locality 43868A. (D) USNM specimen 713193, USNM locality 43838. Scale bars = 1 cm.

FIGURE 72. Noeggerathiales and *Autunia naumannii*. (A) Noeggerathialean foliage, cf. *Noeggerathia*. USNM specimen 713175, USNM locality 43880. (B) Noeggerathialean reproductive structure. USNM specimen 713138, USNM locality 43836C. (C) *Autunia naumannii*. USNM specimen 699121, USNM locality 43868A. Scale bars = 1 cm.

Noeggerathialean Foliage, Possibly *NOEGGERATHIA* Sternberg. Several specimens of unidentifiable foliage, possibly attributable to the Noeggerathiales, were found at two collecting localities, both in the Beeman Formation (Table 4, Figure 72A). These specimens consist of flabellate foliage with a wedge shape, with diverging lateral margins and what may be a rounded distal margin with small, rounded teeth (these margins are not fully preserved). Multiple veins enter the base of each leaf and diverge gradually, the divergence increasing toward the distal margin, the veins branching one to several times, all reaching the distal margin.

We attribute these specimens to the Noeggerathiales in part because of their mutual arrangement on an unseen axis. The leaves are strongly acropetally oriented, overlap slightly, and seem to be helically arranged. Two rows are possibly visible in the illustrated specimen, one folded onto the other. The specimen is similar in many respects to *Noeggerathia foliosa* Sternberg, as emended by Šimůnek and Bek (2003). However, this species is early Bolsovian/Moscovian, middle Middle Pennsylvanian

in age, whereas the Alamogordo specimens are of Missourian/Kasimovian age, Upper Pennsylvanian. Nevertheless, a relationship at the generic level may be possible, although a confident assertion is unwarranted, given the nature of the fossil remains.

Noeggerathialean Reproductive Structure. A single specimen of what is most likely a highly retted noeggerathialean reproductive structure (Figure 72B) was found at the largest of the lower Beeman Formation organic shale deposits (USNM locality 43836C), an allochthonous plant fossil assemblage. Most of the tissue has been removed, leaving only the more sclerenchymatous skeletal framework. This is, in our experience, the most commonly encountered and highly recognizable form of preservation of this plant group in allochthonous deposits. Noeggerathialean strobili consist of stacked, disklike or partial disklike organs (e.g., Bek and Šimůnek, 2005; Wang et al., 2009b, 2017). Most studies of these organs in addressed preservation illustrate nicely preserved material, as well they should, because botanical studies require maximal structural and organizational information (e.g., Bek and Šimůnek, 2005). However, this “chicken-wire”

architecture of the strobiliar components can sometimes not be seen, even in well-preserved specimens (e.g., Wang et al., 2004).

PELTASPERMALES

AUTUNIA NAUMANNII (Gutbier) Kerp. Two small fragments, most likely attributable to *Autunia naumannii*, were recovered in two excavations in the Beeman Formation; one is illustrated in Figure 72C. These specimens consist of pinna fragments in which the ultimate pinnae are pinnule-like in gross morphology; they are wedge shaped, strongly decurrent basally, and incised acroscopically. Clear pinna rachises are not readily visible, which may reflect derivation from distalmost positions on the frond, and can lead to ambiguities in identification. Three to four individual pinnules are borne on each terminal pinna. The pinnules are elongate, narrow, wedge shaped, and basally strongly decurrent. The only venation visible in the pinnules is the midvein, which is slightly sunken, and proceeds the entire length of the pinnule, terminating in a shallow notch in the rounded pinnule apex. These characteristics closely parallel the emended diagnosis of *A. naumannii* by Kerp (1988).

Autunia naumannii can appear similar in overall form to some kinds of *Sphenopteris* foliage; compare it, for example, with *S. germanica*, which is found in the same deposits. *Autunia naumannii* also can be difficult to distinguish from species of *Lodevia* in fragmentary preservation, especially when the preserved remains are small, as is the case here. In the Alamogordo specimens, the absence of a clear rachis in the pinnae and the lack of any obviously developed intercalary pinnules make the separation of these two forms difficult. However, the clearly marked midveins in the pinnules, terminating in an apical notch, are characteristics that aid in final determination. In addition, the Alamogordo specimens do not bear close enough resemblance to any described species of *Lodevia* to be assigned confidently to that genus.

Kerp (1988) notes the occurrence of *Autunia naumannii* in middle to late Stephanian strata of central and eastern Europe. Consequently, the occurrence of this species in Missourian strata of western Pangea, although extending the geographic range of the taxon, is not an exception with regard to known stratigraphic distribution.

PTERIDOSPERMS

ALETHOPTERIS cf. *SCHNEIDERI* Sterzel. Two fragmentary specimens attributable to *Alethopteris* cf. *schneideri* were identified in the collections from the Alamogordo area. USNM specimen 699099 (Figure 73A) is from locality USNM 43868A, which is in the upper part of the lower Beeman Formation; USNM specimen 713148 (Figure 73B) is from locality USNM 43836E, which is in the lower Beeman Formation. Only the latter specimen has a terminal pinnule, but the lateral pinnules and venation in the two are similar, suggesting identity. In both, the lateral veins are somewhat widely spaced, at about 30 veins per centimeter of margin, and the veins are slightly flexuous, rather than stiff

and straight. The pinnules are inserted angularly, with a midvein that runs about 3/4 of the distance through the pinnule. The acroscopic side of the pinnules is deeply incised, and the basiscope side is elongated and decurrent. The pinnules are barely or even not confluent. The most distinctive feature is found on the specimen from locality 43836E (USNM specimen 713148), which has a broad terminal pinnule with a bluntly rounded tip; the immediately subjacent pinnules are not fused with it.

These specimens are similar in most respects to *Alethopteris schneideri* Sterzel. The only reason to give pause to this determination is the stratigraphic distribution of *A. schneideri*, which, as known today, is restricted to the European Rotliegend (thus uppermost Pennsylvanian/lower Permian). As noted by Barthel (2006:51, translation by HK), *A. schneideri* is “so far only known from central Europe, from Karniowice, the Boskovic Furrow to the Saar-Nahe Basin as well as from Busaco. The species apparently only occurs in the Rotliegend and is thus of stratigraphic importance.” However, with the recent recognition of many “early,” disjunct stratigraphic appearances of plants from widely different phylogenetic lineages, termed “Methuselah taxa” by Looy et al. (2014), such a stratigraphic discontinuity may not be impossible.

An additional comparable species, though generally attributed to the Middle Pennsylvanian (e.g., Gillespie and Pfefferkorn, 1979), to which these New Mexico specimens might be assigned is *Alethopteris sullivantii* (Lesquereux) Schimper. Again, as noted by Barthel (1976:100, translation by HK), “of all *Alethopteris* species *Alethopteris schneideri* Stz. can only be compared with *Alethopteris sullivantii* (Lesquereux) Schimper from the Pennsylvanian of the USA and the Westphalian D of England (Sterzel, 1918; Wagner, 1968). However, the pinnules of the latter are anadromous, basally less deeply constricted, often even still fused with each other.” There also are differences in the cuticular features, but we were not able to examine those in the New Mexico specimens and must rely on gross morphology alone.

MACRONEUROPTERIS SCHEUCHZERI (Hoffmann) Cleal, Shute, and Zodrow. *Macroneuropteris* Cleal, Shute, and Zodrow was segregated from *Neuropteris* on the basis of the characteristics of its cuticle (Cleal et al., 1990). *Macroneuropteris scheuchzeri* (Figure 74) is particularly recognizable among Pennsylvanian pteridosperm foliage types because of several distinctive macromorphological characteristics. The first of these is the generally large size of the pinnules, measuring several centimeters in length and often more than a centimeter in width. The second characteristic to be observed, but perhaps the most definitive, is the presence of small, hairlike features on the surface, apically directed but at a slightly outwardly directed angle from the midrib (e.g., Figure 74B), causing them to cross the lateral veins; the density of these features varies from quite dense, to sparse, even to absent or difficult to find. Long interpreted as hairs, Zodrow (2014) demonstrated that their chemistry differed from that of the cuticle and proposed that they were self-assembled, extracuticular deposits. Inspired by this assessment, Laveine and Oudoire (2015) extended the morphological aspects of the analysis and demonstrated that

FIGURE 73. *Alethopteris schneideri*. (A) USNM specimen 699099, USNM locality 43868A. (B) USNM specimen 713148, USNM locality 43836E. Scale bars = 1 cm.

the “hairs” were, in fact, not surficial but, rather, were subepidermal features and most likely to be some form of resin duct. The venation of *M. scheuchzeri* is characteristically fine, with a thin, but well-marked, midvein that extends most of the distance from the cordate base to just below the bluntly rounded to acuminate pinnule apex, where the midvein divides into numerous, dichotomizing secondary veins. Secondary veins are fine, steeply arching away from the midvein, contacting the margin at about a 45° angle, at between 35 and 50 veins per centimeter of margin. At the base, individual *M. scheuchzeri* pinnules may bear 1 or 2 (and in exceptional instances, up to 4) small, ovoid, subsidiary pinnules that can be entirely free from the main pinnule or partially fused with it (Laveine and Legrand, 2008). Furthermore, it is possible to find so-called heteromorphic pinnules that are further divided into elongate lobes that take the shape of a full pinnule but without a clearly defined midvein (Zodrow, 2003: fig. 10a; Laveine and Legrand, 2008: fig. 3; Figure 74E). The most common way to find *M. scheuchzeri* specimens is as isolated pinnules. Evidently, the pinnules were caducous and abscised from the frond as it aged. Fronds were relatively small, on the order of a meter in total length, perhaps more; frond architecture has been analyzed

by Laveine and Brousmiche (1982), Zodrow (2003), and Laveine and Belhis (2007). Another distinctive feature of the plant is the short spines present on the ultimate rachises, which mark the former positions of pinnules; this morphology (Figure 74F) renders these axis segments readily identifiable.

Macroneuropteris scheuchzeri is the most commonly occurring and abundant pteridosperm in the Alamogordo collections. It is present in 10 of 19 samples, at an average frequency of 4.87% across the entire suite of samples and 9.25% in the 10 samples in which it is present (Table 4). All occurrences are in the Beeman Formation.

Examination of the distribution of *Macroneuropteris scheuchzeri* relative to coal beds (immediately above, immediately below, nonassociated) in Pennsylvanian siliciclastic deposits of eastern U.S. coal basins indicated a consistent association and a preference for habitats with well-saturated substrates (Stull et al., 2012). This statistically significant association suggests that the specimens in this collection, from a lithological sequence in which there are no coal beds, probably grew in the wettest parts of the local Missourian-age landscapes. Farther to the east, in north central Texas, in strata of Virgilian age, *M. scheuchzeri*

FIGURE 74. *Macroneuropteris scheuchzeri*. (A) USNM specimen 713116, USNM locality 43868A. (B) USNM specimen 706341, USNM locality 43891. (C) USNM specimen 713196, USNM locality 43838. (D) USNM specimen 699091cp, USNM locality 43868B. (E) USNM specimen 698909, USNM locality 43891. (F) USNM specimen 713117, USNM locality 43836A. Scale bars = 1 cm.

reaches its maximum abundance in organic-rich, siliciclastic swamp deposits, where it may be the overwhelmingly dominant plant (DiMichele et al., 2005; Tabor et al., 2013; Xu et al., 2018).

There has been some debate about the taxonomy of *Macroneuropteris scheuchzeri* due to the often very large size of pinnules found in the United States, particularly in Middle Pennsylvanian

strata, which have been assigned to *Neuropteris* (*Macroneuropteris*) *decipiens* Lesquereux. This matter is fully discussed by Gastaldo (1977). In brief, pinnules of *M. scheuchzeri*, in the sense of the European type species, are generally smaller in size than *M. decipiens*, often have rather elongate-acuminate tips rather than more broadly rounded, blunt apices, have higher vein

densities (high 30s to low 40s compared to upper 20s), lack the intercalary pinnules found in the American form, and are thought to lack small, blunt, crenulate pinnules sometimes found in the American form. Furthermore, as Gastaldo (1977) notes, Darrah (1969) believed *M. decipiens* to be restricted to the late Middle Pennsylvanian of the Illinois Basin. Gastaldo (1977) argues for the continued recognition of these two forms as subspecies. In light of these considerations and the observation that the range of *M. scheuchzeri* terminates at the end of the Middle Pennsylvanian in Europe (Laveine and Belhis, 2007), on the eastern side of the Trans-Pangean mountain range, it is possible that the Late Pennsylvanian form, found in the United States (the Canadian Maritimes are linked to Europe paleogeographically), could be an additional subspecies. For the most part, like the type concept of *M. scheuchzeri*, pinnules from the Late Pennsylvanian of the United States are smaller than those that characterize the upper end of the *M. decipiens* size range, and vein densities are in the upper 30s to near 50, thus higher than in *M. decipiens*. However, they also have blunt, slightly tapering apices and may be somewhat larger than the typical European, Middle Pennsylvanian forms. These features can be seen in the specimens from the Alamogordo area.

NEURODONTOPTERIS AURICULATA Potonié. Specimens attributable to *Neurodontopteris auriculata* were found in four excavations at two stratigraphic levels in the Beeman Formation, reaching a maximum frequency of approximately 6% in the largest (most numerous specimens) of our excavations, the combined USNM localities 43868, 43869, and 43891. We have reported this species from several other locations in the Late Pennsylvanian or early Permian of New Mexico (e.g., DiMichele and Chaney, 2005; DiMichele et al., 2013, 2017b), including from the large Missourian-age, Late Pennsylvanian collections from the Kinney Quarry. It was not found in the Late Pennsylvanian collections from Socorro County, New Mexico (DiMichele et al., 2017a). On the other hand, fairly characteristic specimens of *N. auriculata* have been identified in collections from north central Texas, both in the Markley Formation of Virgilian, Late Pennsylvanian age (DiMichele et al., 2005) and in the Sanzenbacher flora, of early Permian age (DiMichele et al., 2018a).

Neurodontopteris auriculata is quite variable morphologically, which has resulted in a rather complex synonymy. The morphology and taxonomy of the genus and species were reviewed recently by Laveine and Dufour (2013). The most complete specimen in the Alamogordo collections (USNM 698912, Figure 75A) is a fragment of a large pinna that shows clearly the lobation of some pinnules in their lower portion, accompanied by odontopterid venation, which Laveine and Dufour (2013) speculated may be common in this species under some conditions and at some locations. Terminal pinnules of pinnae are large and deltoid in shape (Figure 75B,C) and are fused to some degree with the immediately subtending lateral pinnules, which have odontopterid venation. This type of venation, and a broad attachment to the rachis, may be seen in several pinnules below the terminal, changing to more typically neuropteroid attachments several pinnules removed (Figure 75B,C). Pinnules of this

species show clear midveins or a distinct bundle of veins running up to 3/4 of the pinnule length (Figure 75D,E). The secondary veins arch toward the pinnule margin but at a relatively steep angle and do not meet the margin perpendicularly; these veins fork repeatedly in a somewhat irregular manner, which is one of the features that make isolated *N. auriculata* pinnules identifiable. Pinnules generally are large, usually slightly asymmetrical, sometimes with slightly wavy margins, especially well marked near the base, and with one (basiscopic) to several small, basal pinnule lobes (auriculae), one on each side. Pinnule apices vary from broadly rounded to tapered and bluntly pointed.

The specimens from Alamogordo are similar to those from Texas and likely are the same species. They are not, however, an exact fit to the concept of *Neurodontopteris auriculata*, as discussed by Laveine and Dufour (2013). These western Pangean specimens have venation that is somewhat too coarse and often rather irregular in the patterns of vein forking, pinnules that frequently are lobate, and terminal pinnules that are not free and clearly individuated. Thus, there is a possibility that they constitute a distinct *Neurodontopteris* species.

NEUROPTERIS OVATA Hoffmann. *Neuropteris ovata* is one of the most widely recognized Euramerican forms of Pennsylvanian pteridosperm foliage. Consequently, much has been written about it, and it has been studied in most of its paleobotanically accessible aspects, including such things as frond morphology (Zodrow and Cleal, 1988) and anatomy (Beeler, 1983), taxonomy (Cleal et al., 1990), chemistry (D'Angelo and Zodrow, 2016), physiology (Wilson et al., 2017), aspects of its cuticle as a measure of atmospheric CO₂ composition (Cleal et al., 1999; Montañez et al., 2016), and its role as a stratigraphic indicator (Cleal, 1984). The “species” has been subdivided into several forms, which are similar, nearly indistinguishable in fact, in gross morphology but can be distinguished using the micromorphology of cuticles (Cleal and Zodrow, 1989). In light of the recognition of these distinct forms, it is possible, even probable, that *N. ovata*, as based on such gross morphological features as frond architecture, pinnule shape, venation characteristics, etc., is a complex of closely related species.

The specimens from the Alamogordo area come from only one locality, in the Beeman Formation. All are quite fragmentary, most consisting of small pinna fragments or isolated lateral pinnules; no pinna terminal pinnules were found (Figure 76). The remains, nonetheless, have the typical characteristics of *Neuropteris ovata*. Pinnules are about twice as long as wide. The base is basiscopically slightly auriculate, and the apex is asymmetrically rounded, tending to be curved gradually on the basiscopic side and variably more truncated on the acroscopic side—this shape is well represented in the illustrated specimens. Lateral pinnule margins are slightly curved, with a variably developed degree of concavity on the acroscopic side. The midvein extends more or less to the middle of the pinnule, where it divides into a spray of fine ultimate veins. Ultimate venation is fine and gently curves toward the margin, which veins meet at an angle; lateral veins fork one or more times on their path. It is typical for most *N.*

FIGURE 75. *Neurodontopteris auriculata*. All specimens from USNM locality 43891. (A) USNM specimen 698912P. (B) USNM specimen 713077. (C) USNM specimen 713081. (D) USNM specimen 713082. (E) USNM specimen 713083. Scale bars = 1 cm.

ovata pinnules to have a cordate base, with all pinnule veins originating from a common, slightly basally decurrent midvein. Near the apex of a pinna, however, pinnules tend to be adherent to the rachis over a broader portion of their base, becoming odontopteroid in aspect. Although no terminal pinnules are present, the fusion of the near-terminal lateral pinnules and the loss of distinctiveness of the midvein can be seen in Figure 76A.

Because of similarity, it is worth noting briefly how *Neuropteris ovata* can be differentiated from *Neurodontopteris auriculata* when examining relatively small, fragmentary specimens. The venation in *N. ovata* is finer, denser, and more regular in its path to the pinnule margin than seen in the typical *N. auriculata* pinnules. In the latter, the venation can often appear somewhat disorganized or somewhat irregular. In addition, the pinnules of *N. auriculata* have less regular margins, often with some slight undulation, lending a somewhat irregular shape to the pinnules. Size is not a reliable taxonomic indicator in this instance, but in general, the pinnules of *N. auriculata* can be considerably larger than those of *N. ovata*.

Neuropterids of Uncertain Affinity: Large Pinnules. Figure 77A–C illustrates examples of isolated, generally fragmentary, large pinnules, found in small numbers at several excavations in the two most heavily sampled collecting

sites, both in the lower member of the Beeman Formation. Although generally similar, it is uncertain that such pinnules all represent the same taxon. Figures 77D,E and 78 illustrate the problem and will be discussed below. The smaller class of such pinnules includes specimens generally about 3 cm or more in length and 1–1.5 cm in width. They have rounded, noncordate bases; a narrow attachment point; straight to slightly curved, but gradually converging, lateral margins; and apices that vary from bluntly rounded to gradually acuminate. All have distinctively fine venation. A midvein is poorly developed, although a concentration of veins can be seen in the midline of the pinnule, extending 2/3 or more of the distance to the apex before dissolving in a spray of ultimate veins. The lateral veins dichotomize several times and follow a relatively steeply arching path to the margin, which they contact at an angle. There are about 40 veins, more or less, per centimeter of margin.

The affinities of these specimens are unclear. Initially, we considered that they might be *Neurocallipteris planchardii* (Zeiller) Cleal, Shute, and Zedrow. However, the vein densities are too high, and the midvein is too poorly developed to fall within the morphological scope of that species, even though pinnule size and shape are generally conformant. At this time, we identify pinnules of this form simply as undetermined neuropterids.

FIGURE 76. *Neuropteris ovata*. All specimens from USNM locality 43839. (A) USNM specimen 713225p. (B) USNM specimen 713220p. (C) USNM specimen 713218. (D) USNM specimen 713217. (E) USNM specimen 713219. (F) USNM specimen 713220cp. Scale bars = 1 cm.

FIGURE 77. Neuropteroids of large size and uncertain affinity I. (A) Neuropterid with high vein density, cf. *Neurocallipteris planchardii*. USNM specimen 713093, USNM locality 43891. Arrow points to one of several microconchids on the pinnule surface. (B) Neuropterid with high vein density, cf. *Neurocallipteris planchardii*. USNM specimen 713140, USNM locality 43836C. (C) Neuropterid with high vein density. USNM specimen 699089, USNM locality 43869. (D) Possible odontopterid terminal pinnule. USNM specimen 699110, USNM locality 43868A. (E) Possible odontopterid terminal pinnule of *Odontopteris subcrenulata* type. USNM specimen 698907, USNM locality 43891. Scale bars = 1 cm.

The specimens illustrated in Figure 77D,E are of approximately the same size and shape as those described above. They have similar marginal vein densities, ultimate vein trajectories, and midvein characteristics, although the midvein in

Figure 77D is better defined than in other examples. However, these pinnules have small lobes developed at their bases and veins disposed in such a manner as to suggest the fusion of an odontopterid lateral pinnule with an elongate terminal segment

FIGURE 78. Neuropteroids of large size and uncertain affinity II. All specimens from USNM locality 43891. (A) USNM specimen 713079. (B) USNM specimen 698908. (C) USNM specimen 698906cp. (D) USNM specimen 706328. Scale bars = 1 cm.

at a pinna tip. We thus interpret pinnules of this form as likely of odontopterid affinity, possibly *Odontopteris subcrenulata*, described below.

Continuing with the consideration of large neuropteroid pinnules, Figure 78A–C illustrates still larger pinnules, all found isolated and dispersed. Such specimens differ distinctly from those illustrated in Figure 77A–E. They not only are larger, but they have well-developed midveins and even higher lateral vein densities, approximately 50 veins per centimeter of margin. In some respects, they are similar to *Macroneuropteris scheuchzeri* in size, shape, and vein density. But in size, they are more like the *M. scheuchzeri* forma *decipiens* (Lesquereux) Gastaldo (often described as a variety of *M. scheuchzeri*; e.g., Cleal and Shute, 2012). This species was originally described as *Neuropteris decipiens* Lesquereux from the Mazon Creek, Middle Pennsylvanian flora of Illinois, which, as discussed by Sellards (1908:414), generally has 1/2 to 1/3 lower vein densities than typical *M. scheuchzeri*. Modern studies of sun versus shade leaves suggest that such vein density differences may appear as a result simply of leaf distance from the base of the tree, as a consequence of a drop in hydraulic pressure over distance (Zwieniecki et al., 2004), and as a result of the inability of leaves near the top of the plant to expand as greatly as those in a more favorable hydraulic environment, near the crown base; one of the effects of this difference in the oaks that Zwieniecki et al. examined was greater distance between veins in the more basally located leaves. Perhaps such factors may have been at play with *M. scheuchzeri*. As a final note, however, these large pinnules lack any evidence of the resin ducts (“hairs”) typical of *M. scheuchzeri* foliar laminae. As with the specimens discussed above, we identify these large pinnules only to the level of neuropteroids.

Few cyclopterid pinnules were found. One is illustrated in Figure 78D. Its affinities are undoubtedly with one of the medullosan pteridosperm taxa.

ODONTOPTERIS Sternberg and MIXONEURA Weiss. Described below are specimens of several distinct species or species groups that we assign to either *Odontopteris* or *Mixoneura*. Recently, Laveine and Oudoire (2016) revised the diagnosis of *Odontopteris* on the basis of frond architecture, restricting it to those species known to have bifurcate outer semipinnate fronds, generally less than 2 m in length. They further segregated the new genus *Cyrrillopteris* Laveine and Oudoire from *Odontopteris* sensu stricto on the basis of its frond morphology, which is bifurcate bipinnate. Laveine and Oudoire (2016) encompass in *Mixoneura* those species traditionally assigned to *Odontopteris* known to have bifurcate outer/inner semipinnate fronds (see also Laveine and Delbecq, 2011). However, there is a complex history surrounding the use of the name *Mixoneura*, which is summed up in Wagner and Castro (1998), Laveine and Dufour (2013), and Laveine and Oudoire (2016).

The frond architecture in these plants also may be developmentally variable, depending on the growth stage of the parent plant, as suggested by Krings et al. (2006). Laveine and Oudoire (2016) take exception to the conclusions of the Krings et al. study, but our reading of the latter paper suggests that the two sets of

authors are not in conflict in their interpretations. The need to divide the odontopterids, sensu lato, into two or even three lineages appears to be substantiated by such data as exist. Readers, therefore, hopefully will excuse us for hedging our bets with the use of these names, particularly *Mixoneura*, which remains in need of a thorough revision, including cuticular studies.

ODONTOPTERIS with Small Pinnules. A single specimen possibly attributable to *Odontopteris* sensu stricto (Figure 79A) was found in excavation USNM 43868A. This excavation was the second largest, in terms of specimens recovered, making it likely that some less common elements of the flora might be found. This specimen is relatively small, in the range of *O. cantabrica*, but its venation is too dense to permit it to be referred to that species. In addition, assuming it to be the terminal pinnule of a pinna, the tip is too narrowly triangular, almost pointed, to be assigned to *Mixoneura*. The well-developed midvein/rachis; nearly complete fusion of subapical pinnules, resulting in an undulatory margin; and lack of clearly developed midveins in the confluent pinnules are consistent with an *Odontopteris* sensu stricto generic assignment. Further classification is not possible.

ODONTOPTERIS BRARDII (Brongniart) Sternberg. A small number of specimens that we attribute to the *Odontopteris brardii* group of species or possibly even to *Cyrrillopteris* (Figure 79B–D) were found in three excavations from the Beeman Formation. The *O. brardii* species group consists of *O. brardii*, the type of the genus, sensu stricto, *O. reichiana* Gutbier, *O. minor* Brongniart, *O. minor* f. *zeilleri* Potonié, and *O. nemejcii* Šimůnek and Cleal, among other less well characterized or less fully known species. Laveine and Oudoire (2016) have synonymized all these species under the generic type *O. brardii*. A considerable amount is known about these species (e.g., a selection of more recent studies includes Barthel, 1961, 1962; Kerp and Krings, 2003; Šimůnek and Cleal, 2004; Krings et al., 2006, in addition to Laveine and Oudoire, 2016). Because our specimens are all fragmentary, it would be very difficult to differentiate them to any of the species typically included in the group, although they compare rather closely to several, most notably *O. brardii* itself, although the pinnules are somewhat less acute at the apex than is typical of that species. Others include, in particular, *O. nemejcii*, which Šimůnek and Cleal (2004) suggested as being restricted to more-moisture-limited habitats in intermontane regions. Southern New Mexico, during the Pennsylvanian, had many areas of local tectonism, so it is possible that these plants were transported into the depositional setting from locally more elevated areas.

In the few Alamogordo specimens recovered (Figure 79B–D), pinnules are of triangular shape, with converging lateral margins and a bluntly rounded apex. They are variable in size, acroscopically inclined, and subopposite to alternate in their attachment to the supporting rachis, which is by the full width of their base. Pinnules may be slightly confluent basally and, near the apical part of the pinna, slightly constricted acroscopically. In the small specimens at our disposal, lateral pinnules have a clearly developed

FIGURE 79. *Odontopteris* species. (A) *Odontopteris* sp. USNM specimen 699111, USNM locality 43868A. (B) *Odontopteris* cf. *brardii*. USNM specimen 699105, USNM locality 43868A. (C) *Odontopteris* cf. *brardii*. USNM specimen 713165, USNM locality 43836. (D) *Odontopteris* cf. *brardii*. USNM specimen 699106, USNM locality 43868A. (E) *O.* cf. *cantabrica*. USNM specimen 713149, USNM locality 43836E. (F) *O.* cf. *schlotheimii*. USNM specimen 699112, USNM locality 43868A. (G) *O.* cf. *schlotheimii*. USNM specimen 698024, USNM locality 43836A. (H) *O.* cf. *schlotheimii*. USNM specimen 713093, USNM locality 43891. (I) *O.* cf. *schlotheimii*. USNM specimen 713173, USNM locality 43881. Scale bars = 1 cm (all specimens at same magnification).

midvein, and two or three principal veins appear to enter the lamina; additional veins may enter the lamina from the rachis in pinnules near the pinna apex and in large pinnules (Figure 79D). Just how this venation is interpreted depends on how one wishes to characterize the ultimate vein insertions—from its point of origin, the acroscopic vein initially runs close to, and only gradually departs from, the rachis, and some of the more basiscopically located veins may originate from the base of the acroscopic vein. This feature is easily obscured by preservation state, thus making some of the basiscopic veins appear to have independent insertions on the rachis (see description and discussion in Šimůnek and Cleal, 2004). In any case, the acroscopic vein undergoes more bifurcations than those veins closer to the basiscopic end of the pinnule. The ultimate veins may appear rather thickened or even doubled in some pinnules, a phenomenon possibly caused by the presence of secretory canals in the pinnule lamina. The terminal pinnule (Figure 79B,C) is small but variable in size, elongate and bluntly tapered at the apex, and fused to a relatively small degree with the immediately subjacent lateral pinnules. The pinna rachis is coarsely longitudinally striated (see particularly Figure 79B), the most visible striae possibly being secretory canals, as suggested by Laveine and Oudoire (2016).

ODONTOPTERIS with Sparse Veins. There is a group of *Odontopteris* species that are similar in having sparse pinnule venation and variable pinnule morphology. Included are *Odontopteris schlotheimii* Brongniart, *O. aequalis* Lesquereux, and *O. cantabrica* Wagner. Two of these species, *O. aequalis* and *O. cantabrica*, are reported primarily from the upper part of the Middle Pennsylvanian and lower Upper Pennsylvanian (Cantabrian in European terminology; Desmoinesian and Missourian in U.S. regional terms). *Odontopteris schlotheimii* occurs primarily in the lower Permian but also has been identified in Upper Pennsylvanian strata, with suggested occurrences in the upper Middle Pennsylvanian (summarized in Barthel and Amelang, 2011; see, particularly, Lesquereux, 1880, who mentions, but does not illustrate, occurrences in the shale above the coal at Morris, Illinois, which is the Francis Creek Shale, above the Colchester Coal, that hosts the Mazon Creek flora). When examined side by side, these species are very difficult to tell apart and may constitute a clade of related species or even a single, long-ranging species. Detailed study of cuticular features may shed light on this problem, and some steps have been taken in that direction (Cleal et al., 2007). Note that these species have been considered to belong more properly in *Mixoneura*; because the use of the name *Odontopteris* continues in several recent publications (e.g., Cleal et al., 2007; Barthel and Amelang, 2011), we continue to use that generic name here.

Cleal (1997) and Cleal et al. (2007) suggested that *Odontopteris aequalis* and *O. cantabrica* may be synonymous, the “species” differing only in size. They note, however, this may be difficult to determine because of the fragmentary state of the type of *O. aequalis* (Lesquereux, 1866). Wittry (2006), however, noted that *O. aequalis* is the most common odontopterid in the well-known Mazon Creek flora, of late Middle Pennsylvanian (Desmoinesian) age. The Francis Creek Shale is the type stratum

of *O. aequalis*. Wittry illustrated many additional specimens, which he attributed to *O. aequalis*, displaying a considerable range of pinnule morphological variation. If his identifications are accepted, they document comparability of *O. aequalis* and *O. schlotheimii*, which Barthel and Amelang (2011), in an analysis of a large suite of *O. schlotheimii* specimens from a single locality, demonstrated to have frond dimorphism and considerable variation in pinnule shape. Barthel and Amelang (2011) note, again, the similarities between *O. schlotheimii* and *O. cantabrica*. This suite of observations (those of Lesquereux, Cleal et al., Wittry, and Barthel and Amelang) points yet further to the similarity among these three species, in a sense closing the circle of comparability in form and in stratigraphic distribution.

This taxonomic problem is not simply one of biological interest to specialists. It raises some significant matters regarding the use of *Odontopteris cantabrica* in biostratigraphic zonation. This species occurs widely, but rarely, in Euramerican strata and has been considered diagnostic of the base of a terrestrial Cantabrian stage, defined in western Europe (Wagner and Winkler-Prins, 1985). The Cantabrian, as currently constituted, spans the upper part of the Middle Pennsylvanian and lower to middle parts of the Upper Pennsylvanian. *Odontopteris cantabrica*, furthermore, is the namesake of a macrofloral zone (Wagner, 1984), widely applied to European terrestrial strata (e.g., Cleal, 1997; Cleal et al., 2003; Wagner and Álvarez-Vázquez, 2010; Wagner and Castro, 2011; Opluštil et al., 2016, 2017, 2018), marking the lower part of the Cantabrian Stage in Europe. Similar morphotypes also have been identified in the United States and Canada in this same stratigraphic interval as *O. cantabrica* or *O. aequalis* in more recent literature (Gastaldo, 1977; McComas, 1988; Cleal, 1997; Wittry, 2006; Zedrow et al., 2012; Moore et al., 2014; DiMichele et al., 2017a) and have been used to identify the base of the Cantabrian in North America. If *O. cantabrica* should prove to be the same species as *O. aequalis* and *O. schlotheimii*, the two former names will be submerged as later synonyms of *O. schlotheimii*, and the stratigraphic utility of this species will be diminished.

Sparse-veined odontopterids, attributable to this *O. cantabrica/aequalis/schlotheimii* group, were found in seven excavations from three stratigraphic levels in the Beeman Formation. Specimens are rare and, in the larger collections, are present at frequencies between 1% and 2%. A selection of specimens is illustrated in Figure 79E–I. In these specimens, all pinna fragments, several preserving all or part of the terminal pinnule, the pinnules are ovoid with bluntly rounded apices and outwardly curving lateral margins. In some specimens the lateral pinnules are variably confluent; in other specimens the pinnules converge toward the base and are largely nonconfluent. In all specimens in which an apical pinnule is present, the immediately subjacent pinnules are strongly fused, creating a continuous lamina with an undulatory margin. It appears that where pinnules are fully free, two separate veins enter the base of the pinnule and that the majority of ascending, ultimate veins arise from the more acroscopic of the two principal veins. In the single specimen we have attributed to *O. cantabrica* (Figure 79E), the terminal pinnule tip is triangular, tapering to a

blunt point. In the specimens we have attributed to *O. schlotheimii* (Figure 79F–I), the terminal pinnules, to the extent they are preserved, appear to be relatively large, broad, and bluntly rounded. Both of these morphologies fall within the scope of specimens attributed to *O. schlotheimii* by Barthel and Amelang (2011).

Both morphological types (*Odontopteris cantabricalaequalis* and *O. schlotheimii*) were encountered in the floras from Socorro County (DiMichele et al., 2017a, 2017b), equivalent in age to those reported here. Those most like *O. cantabrica* occurred only in the Bartolo Formation, which is Middle Pennsylvanian, upper Desmoinesian (lower Cantabrian in European regional terminology). Those from the Missourian were identified as *O. schlotheimii* and have characteristics similar to the specimens so-identified here. Note, however, that some of the specimens identified as *O. schlotheimii* from the higher (Virgilian) strata in the Abo Pass section (DiMichele et al., 2017b) are similar to *Neurodopteris auriculata* and may belong to that species instead (except the specimens from unit 35, the oldest of the stratal units described in that paper).

MIXONEURA (ODONTOPTERIS) LINGULATA (Göppert) Wagner and Castro. *Mixoneura (Odontopteris) lingulata* is a mixoneuroid odontopterid that has pinnae with very long terminal pinnules that usually occupy about half or more of the pinna length and are lobed at the base or can be subtended by a pair or a few pairs of smaller, round pinnules. The large terminal pinnules may taper toward the apex. The number of typically odontopterid lateral pinnules in a pinna varies with the position in the frond but is usually not more than 4. The frond is bifurcated, and the internal pinnae, those between the limbs of the bifurcated main rachis, are shorter than the external ones. Moreover, the pinnae in the middle portions of the frond are longer than the basal and apical ones. Kerp and Fichter (1985: pl. 21, fig. 1) show a specimen that is quite similar to those shown here (Figure 80).

Only two *Mixoneura lingulata* specimens were identified in the Alamogordo collections, both from USNM locality 43891, the largest of the Beeman Formation collections. Both specimens are illustrated here (Figure 80).

Mixoneura (Odontopteris) lingulata is quite similar to *M. (Odontopteris) subcrenulata* (Rost) Zeiller. Some authors regard *M. lingulata* as a synonym of *M. subcrenulata*, but we believe they are clearly different. *Mixoneura (Odontopteris) subcrenulata* has much shorter terminal pinnules that are about 1/4 to 1/3 of the total pinna length (in larger pinnae even less), and the number of small, typically odontopterid lateral pinnules in a pinna is usually much higher (up to 8 to 10) than in *M. lingulata*. The terminal pinnule of *M. subcrenulata* is often more rounded than that of *M. lingulata*.

MIXONEURA (ODONTOPTERIS) SUBCRENULATA (Rost) Zeiller. Fossil plant remains that we have assigned to *Mixoneura (Odontopteris) subcrenulata* are the most commonly encountered odontopterid in the Alamogordo collections (Figure 81A–E). Such remains have been identified in 7 of the 19 adpression collections, in both the Beeman Formation,

Missourian age, and the Holder Formation, Virgilian age. *Mixoneura subcrenulata* generally occurs in low frequency, although reaching 20% in USNM 43871. Most of the fossil remains are fragmentary, consisting of isolated lateral or terminal pinnules (Figure 81B–E). The species also has been identified in collections from most of the other areas where Upper Pennsylvanian terrestrial strata crop out in New Mexico (e.g., DiMichele et al., 2017a). As noted by Wagner and Álvarez-Vázquez (2010), the species has “extrabasinal connotations,” which can be restated as preferring seasonally dry habitats, and thus environments where preservational potential is relatively low compared to that of wetland-dwelling species, but also habitats that even if present in the fossil record, are not found as readily as wetland habitats associated with coal beds and thus made readily available to researchers and collectors.

Mixoneura (Odontopteris) subcrenulata has pinnae with a series of well-developed, free, apically rounded pinnules and a large, tongue-shaped terminal pinnule with a blunt apex. Pinnule apices are rounded, and the distinction between terminal and lateral margins may be unclear, particularly in more apical parts of pinnae and where pinnules are small. Several decurrent principal veins enter the base of a pinnule from the supporting rachis. These are often organized in such a manner that there is a central concentration of veins, a vein bundle, that gives the appearance of a midvein, but this generally dissipates into a spray of veins by 1/2 or less of the distance to the pinnule apex (Figure 81A,D,E); terminal pinnules, on the other hand, have a well-developed midvein, basically an extension of the rachis (Figure 81A–C). The secondary veins arch steeply toward the pinnule margin, fork multiple times, and form a relatively dense venation.

As discussed above, *Mixoneura (Odontopteris) subcrenulata* is similar to *M. lingulata*, but the latter has fewer pinnules, and the terminal pinnule is longer, more slender, and less blunt. In Europe, *M. lingulata* is restricted to the Rotliegend, whereas *M. subcrenulata* is common in the Stephanian; it has been identified in a number of lower Permian, Lower Rotliegend equivalent plant fossil collections in the western United States (e.g., DiMichele et al., 2018a). In fragmentary preservation, it also may be difficult to separate *M. subcrenulata* remains from those of neuropterids (e.g., Figure 81F) or from those of *O. schlotheimii* or *O. cantabrica* (Figure 81G,H), which also possibly should be assigned to *Mixoneura* rather than *Odontopteris*. The latter two species have less dense venation than *M. subcrenulata*, but as in the specimens illustrated, that venation may be obscure or not sparse enough to make identity certain one way or the other.

Cf. DICKSONITES PLUCKENETII (Schlotheim ex Sternberg) Sterzel. *Dicksonites pluckenetii* is a pteridosperm of the family Callistophytaceae (e.g., Meyen and Lemoigne, 1986; Galtier and Béthoux, 2002). The plant has a scrambling or facultatively climbing habit (Galtier and Béthoux, 2002). It is one of the most completely known plants of the Pennsylvanian interval, thanks to its abundant preservation in coal balls (Rothwell, 1981).

FIGURE 80. *Mixoneura* (*Odontopteris*) *lingulata*. Both specimens from USNM locality 43891. (A) USNM specimen 713091. (B) USNM specimen 713092. Scale bars = 1 cm.

A single, fragmentary specimen was found in one of the larger collections, from the Beeman Formation, which we assign tentatively to this species (Figure 82A). The assignment is tentative because the fragment, although well preserved, is small and may not reveal the full variability in pinna and pinnule morphology necessary to identify it with confidence. Pinnules of this specimen are triangular and slightly falcate in shape, with an acuminate apex and curved, smooth to possibly slightly undulate margins. They possess a well-developed, if thin, midvein and sparse lateral veins that are one forked, becoming unforked near the apex. The midvein insertion is angular but not strikingly decurrent. The pinnules are slightly confluent basally and, like the midvein, are acropically inclined. On the basis of the quality of preservation, the pinnule lamina appears to have been robust. There are no indications of transverse bars on the ultimate rachis. The overall morphology of this fragment also is reminiscent

of the mariopterids, including *Mariopteris* Zeiller, although a mariopterid assignment appears to be excluded as a possibility by the absence of transverse bars on the rachis, and *Fortopteris* Boersma. Mariopterids have been reported from the Stephanian (e.g., Wagner and Álvarez-Ramis, 1967), although not of this particular morphology.

Cf. *LAVEINEOPTERIS* cf. *RARINERVIS* (Bunbury) Cleal, Shute, and Zodrow. Several small fragments of pteridosperm foliage possibly attributable to *Laveineopteris* Cleal, Shute, and Zodrow were found in the Beeman Formation in the largest of the Alamogordo collections. These specimens all consist of small neuropteroid pinnules, strongly adherent to a segment of rachis (Figure 82B). The pinnules are rectangular in shape, with slightly undulate lateral edges and a blunt apex; the base is cordate, but the pinnules are attached to the upper arc of the rachis, making it appear, superficially, as if the pinnules

FIGURE 81. *Mixoneura* (*Odontopterus*) *subcrenulata* (A–E) and cf. *O. subcrenulata* (F–H). (A) USNM specimen 713240p, USNM locality 43871. (B) USNM specimen 713221, USNM locality 43839. (C) USNM specimen 713141, USNM locality 43836C. (D) USNM specimen 713244, USNM locality 43430. (E) USNM specimen 699114, USNM locality 43868A. (F) Pinnule of uncertain affinity, possibly *O. subcrenulata* or perhaps a neuropterid. USNM specimen 699101, USNM locality 43868A. (G) Cf. *O. subcrenulata*, but perhaps with venation too sparse; note particularly partially free lateral pinnule on right. USNM specimen 699097, USNM locality 43868B. (H) Cf. *O. subcrenulata*, but perhaps with venation too sparse. USNM specimen 713110, USNM locality 43868A. Scale bars = 1 cm.

FIGURE 82. Miscellaneous pteridosperms. (A) Cf. *Dicksonites pluckenettii*. USNM specimen 699100, USNM locality 43868A. (B) Cf. *Laveineopteris rarinervis*. USNM specimen 713080, USNM locality 43891. (C) Cf. *Reticulopteris* sp. USNM specimen 713169, USNM locality, 43836F. Scale bars = 1 cm.

adhere to the rachis broadly. The midvein is moderately strong and extends about 2/3 of the distance from base to apex. Lateral veins are strongly marked and notably sparse. Their path from midvein to margin is steep, with little or no curvature. In these characteristics, the specimens are similar to *L. rarinervis*, although this occurrence would be above the top of the range of that species, as reported in Europe (Opluštil et al., 2016).

Cf. *RETICULOPTERIS* Gothan. A single pinnule attributable to *Reticulopteris* sp. was found in dark shales from the lower Beeman Formation (Figure 82C). The pinnule is approximately 1 cm in length and 7 mm in width, thus fairly large. A midvein is only poorly developed, extending about 25%–30% of the distance between the base and the apex, and is not notably different in thickness or prominence from the rest of the venation. Lateral veins follow a relatively steep and straight course to the margin, which they meet at an approximately 45° angle. The lateral veins fork 3–5 times and weakly and irregularly anastomose, forming elongate polygonal meshes that become smaller toward the margin. These are characteristics of the genus *Reticulopteris* (see discussions in Zodrow and Cleal, 1993, and Laveine and DiMichele, 2001).

Pinnules of this same type were reported recently from two other Late Pennsylvanian areas in Socorro County, New Mexico (DiMichele et al., 2017a, 2017b), and recently, another isolated pinnule was identified from an early Permian, Wolfcampian (Asselian) age deposit in north central Texas (DiMichele et al., 2018a). In these instances, just as in the present occurrence, the pinnules are isolated, which makes a definitive taxonomic assignment dubious. Thus, we determine these plant remains only to the generic level and even so with considerable tentativeness.

CONIFEROPHYTES

Cordaitaleans. Cordaitalean foliage is the most widespread and abundant type of taxonomically resolvable plant remains among the Alamogordo collecting localities. Because of the lack of a preserved cuticle in these specimens, it is not possible to identify this foliage taxonomically to the resolution recently created by Šimůnek and colleagues (e.g., Šimůnek, 2000, 2015, 2018; Šimůnek and Florjan, 2013). Most of the foliage is represented by multiveined leaves a few centimeters in width (Figures 83A–D, 84A). Occasionally more narrow, around 1 cm wide, leaves are present (Figure 84B–D) and may characterize particular collecting localities, suggesting diversity among the cordaitalean elements of the flora. Tapering leaf segments that terminate in typically rounded leaf tips (Figure 84A), with arcuate, slightly thickened bases indicative of leaf abscission (Figures 83A, 84C), are found on occasion. But the great majority of the remains are simply fragments that are characterized by parallel venation and fine interveinal striae.

This kind of foliage, particularly if highly fragmentary (Figure 83C,D), can be difficult to distinguish from rachial fragments of pteridosperms or marattialean tree ferns. The latter also are of similar size and may have dense longitudinal striations. Those striations, however, are much less regular in thickness and length than are those of cordaitaleans, and they often anastomose or terminate blindly, a pattern that is not found in the veins of cordaitalean leaves. Of course, rachial segments often show a degree of three-dimensionality that permits them to be distinguished. Despite the differences, we suspect, as noted by Bashforth et al. (2016), that cordaitaleans

FIGURE 83. Cordaitalean wide leaves. (A) Leaf base. USNM specimen 706331, USNM locality 43891. (B) Note attached microconchids and interveinal striae. USNM specimen 713168, USNM locality 43836F. (C) From Holder Formation. USNM specimen 713245, USNM locality 43430. (D) USNM specimen 713229, USNM locality 43839. Scale bars = 1 cm.

FIGURE 84. Cordaitalean leaves. (A) Wide leaf. Field photograph, USNM locality 43836, specimen uncollected from sandstone at base of outcrop. (B) Narrow leaf near base. USNM specimen 699136, USNM locality 43868A. (C) Narrow leaf base. USNM specimen 713232, USNM locality 43839. (D) Narrow leaf base. USNM specimen 706329, USNM locality 43891. Scale bars = 1 cm, except where noted.

are underrepresented in the counts performed when quantifying paleofloristic collections because they are mistaken for striate axes.

As with other broad laminar types of foliage, such as *Taeniopteris*, cordaitalean foliage may have floated in the water column prior to its ultimate incorporation into the sediment. In such instances, microconchids (also known as spirorbids) are

frequently attached to the leaf laminae (e.g., Figure 83B). Microconchids may indicate some degree of brackishness, although that is subject to debate (Gierlowski-Kordesch and Cassle, 2015; Zatoń et al., 2016).

New Conifer-Like Cordaitalean. Foliage of a conifer-like plant with distinctive long, relatively straight leaves (Figure 85) was common among the specimens collected

FIGURE 85. New coniferophyte. (A) Vegetative remains. USNM specimen 617966, USNM locality 43891. (B) Vegetative remains. USNM specimen 698903, USNM locality 43891. (C) Fertile segment and vegetative branches attached to a common axis. USNM specimen 617950, USNM locality 43868B. (D) Fertile segment below vegetative leaves on a common axis. USNM specimen 713192, USNM locality 43838. Scale bars = 1 cm, unless otherwise noted.

from the four excavations in unit 206 of the Space History B section, Beeman Formation. We initially identified this plant as some sort of conifer, likely with plagiotropic branches, on the basis of the relatively large branch fragments recovered (e.g., Figure 85A). However, additional collecting, and good fortune, revealed several specimens that combined this distinctive foliage with reproductive structures similar to *Cordaitanthus* Feistmantel, but of a form unknown to us (Figure 85C,D). Furthermore, on the basis of the relative positions of fertile and vegetative remains, on different specimens, it is possible (the evidence is ambiguous and subject to interpretation) that the reproductive organs may have been disposed as fertile regions on the penultimate axes, not in distinct strobili, in which case they would be positioned as ultimate, appendicular organs.

A separate description of this plant is in progress (Looy et al., 2017).

Branch systems in the new plant appear to be orthotropic. Although it appears upon casual observation that the ultimate axes are positioned oppositely to suboppositely, a closer examination suggests that they are helically arranged. Note, for example, in Figure 85A that the distances between ultimate branches are variable; some have no clear opposite branch and, in fact, may originate on the underside of the specimen. Thus, the plants may have borne their branches in a plagiotropic manner, even if not originating as is typical for such a growth habit (see, for example, the walchian conifer *Hanskerpia*; Rothwell et al., 2005). Ultimate branches are axillary. Leaves on all branch orders are straight to gently concave,

are widest at the base, and taper gradually throughout their length to an acute tip (Figure 85C), similar to comparable ultimate leaves of several large-leaved walchian conifer taxa (e.g., *Hanskerpia* and *Barthelia*; Rothwell and Mapes, 2001; Rothwell et al., 2005). The leaf attachment is confined to a small oval area; the bases are not clasping but may contract slightly at the point of attachment. A single vein runs the length of the leaf. Leaves are up to 30 mm long on the penultimate axes and up to ~1.5 mm wide at the base. Penultimate leaves vary in angle of attachment but generally appear to be quite adpressed against the stem. Leaves on ultimate branches are up to 19 mm long and about 1 mm wide at the base and become shorter toward branch apices, between 9 and 14 mm long on the specimens measured. Leaves on ultimate axes are mostly inserted at angles of about 45°, but that angle is somewhat larger at the base of the axis.

Reproductive organs are part of the penultimate axis branching system. As illustrated by Figure 85D, there is an abrupt transition from the vegetative to fertile region. In the fertile region, axillary short shoots are subtended by bracts that have a shape similar to the penultimate leaves; these bracts are up to 25 mm long in the one specimen where they can be observed with certainty. Although mostly obscured in the study specimen, several of the short shoots appear to bear a recurved axis with two swollen tips, which we interpret as a megasporophyll with an ovule attachment scar. On the basis of their positioning, the fertile short radial shoots appear to be arranged decussately in four orthostichies, separated from one another by about 5 mm.

The vegetative portions of this new plant are similar to what Rothwell et al. (2005) have described as *Hanskerpia hamiltonensis* Rothwell, Mapes, and Hernandez-Castillo, which has relatively long, needlelike leaves on the ultimate branches, held at an approximately 45° angle to the supporting axis. These leaves are only slightly curved, but they do trend toward being sinuoidal with moderately elongate bases. *Hanskerpia* has forked leaves on the penultimate branches, and the bracts are forked as well. *Barthelia furcata* Rothwell and Mapes (2001) is similar to the Alamogordo conifer in having concave, long, nonsinuoidal ultimate-shoot leaves. However, like *Hanskerpia*, *Barthelia* has forked bracts, which thickly invest its penultimate axes. Both of these plants are of Late Pennsylvanian age and are known as whole plants, including anatomy and reproductive organs; thus, our basis for comparison is limited.

Two species illustrated in the multipart monograph of Florin (1938, 1939a, 1939b, 1939c, 1940, 1944a, 1944b, 1945) have long, very loosely arranged leaves, described as *Walchia* (?*Lebachia*) *gallica* Florin and *Walchia* (?*Lebachia*) *schlotheimii* Florin. Both species have been transferred to the genus *Hermitia* Visscher, Kerp, and Clement-Westerhof by Visscher et al. (1986). The leaves of *Hermitia gallica* (Florin) Visscher, Kerp, and Clement-Westerhof are 5–12 mm long; those of *W. schlotheimii* are 10–22 mm. The leaves of *H. gallica* are, however, S shaped, whereas those of the new coniferophyte are straighter and have a nondecurent insertion. It is worth noting that leaves on

penultimate branches of the so-called walchian conifers, conifers with plagiotropic branching patterns, generally are not longer than 25 mm (*W. schlotheimii* is a maximum of 22 mm, *Walchia piniformis* Schlotheim ex Sternberg is up to 25 mm). We measured the leaves in the new coniferophyte to be more than 30 mm.

This coniferophyte is relatively common in the lower member of the Beeman Formation, where it occurs at all but one stratigraphic level sampled. Its average frequency of occurrence is 6.83% (Table 4). However, this average occurrence is skewed by a nearly complete absence in the excavations from the lower Beeman black shale localities, where it was found in only two of the seven excavations, at frequencies of 1.72% and 2.38%. When considering only localities at which occurrences are recorded, the plant is present in 11.79% of sampling quadrats. It occurs at >20% in collections from three of the excavations, with a maximum frequency of 24%.

Walchian Conifers. Identifying fragmentary vegetative remains of conifer shoots is among the most challenging tasks faced in late Paleozoic floristic studies. Not only are specimens difficult to identify with confidence, particularly without epidermal characteristics from cuticles or mineralized leaf surfaces, but nomenclatural practice itself is subject to considerable disagreement (e.g., Visscher et al., 1986; Rothwell and Mapes, 2001; Hernandez-Castillo et al., 2014). Here, as best as possible given the state of the Alamogordo material, we follow the classification of Visscher et al. (1986) and subsequent studies of that group, which incorporates some flexibility in the naming of fragmentary remains. We also considered the morphological similarity of the Alamogordo specimens to other conifer remains of Late Pennsylvanian age described from the United States—west central and western Pangea—encompassed by an entirely different scheme of classification. These new taxa are presented as whole-plant genera and species from strata of Late Pennsylvanian (Missourian and Virgilian) through early Permian age in the central and western United States and Mexico (see Hernandez-Castillo et al., 2014, for a summary and comparison with the Visscher et al., 1986, scheme, as subsequently modified).

When emended and recast as a whole-plant genus by Clement-Westerhof (1984), discussed further by Visscher et al. (1986), *Walchia*, then one of the most commonly used names for adpressed conifer foliage, was greatly reduced to only four species. Most of the other species were placed in several new genera reserved for specimens that were only partially known. This diversity was further reduced by the description of *Otovicia* (Kerp et al., 1990), also cast as a whole-plant genus. The emendation of *Culmitzschia* Florin by Clement-Westerhof (1984) for walchian specimens with a known cuticle but lacking association with reproductive organs resulted in 11 species falling within the new circumscription of that genus (Visscher et al., 1986).

As can be seen from descriptions and from fig. 1 of Looy and Duijnste (2013), in table 2 of Hernandez-Castillo et al. (2014), or in table 1 of Winston (1984), the differences in the gross morphology of walchian foliated shoots, excluding the cuticle, are heavily dependent on quantitative features. These include such

attributes as leaf length and width, leaf shape, attitude of the leaf tip, the angle between the leaf and the supporting stem, and the presence or absence of heterophylly, generally expressed as forked leaves on penultimate shoots and unforked leaves on ultimate shoots. Our experience suggests that such quantitative differences are not particularly useful in detail, given the degree of morphological variation, likely intraspecific, within many fossil populations of conifers from single, narrowly constrained excavations. Thus, we assign names to the Alamogordo fossils with reticence and with a warning to the reader that due caution is required in evaluating the assignments.

During quantification of the Alamogordo floras, conifer specimens with small, sinusoidal leaves borne at a narrow angle to the stem were identified simply as walchian conifers and not subdivided. As a whole, these types of conifers have small leaves that curve apically, with somewhat reflexed tips. The leaves appear to be narrow in the longitudinal-section views generally presented in adpression preservation but actually may be seen, where face views are preserved, to be more triangularly wedge shaped in some specimens. This type of foliage is found in 14 of 19 excavations and at all but three stratigraphic horizons, absent in both the uppermost Gobbler Formation and the Holder Formation. In these latter two units, no conifer remains of any type were recovered; the small Holder flora contains leaves of cordaitaleans, a coniferophyte group.

At least three walchian-conifer morphotypes (*sensu* Looy and Duijnste, 2013) occur in the Alamogordo flora, probably more. The most widely occurring of these (Figure 86A–D) has linear to triangular, 2–3 mm long, and 1 mm wide ultimate leaves that are only slightly sinusoidal in shape and recurved at the tips, which are subobtusely tapered. The leaves are strongly imbricate and make less than a 45° angle with the stem. The leaves on the penultimate axes are up to 4 mm longer than those on the ultimate axes but otherwise are similar in shape, aspect, and imbrication to those of the ultimate axes. The gross morphology of these specimens is comparable to those of *Otovicia hypnoides* (Brongniart) Kerp, Poort, Swinkels, and Verwer. *Otovicia hypnoides* has very small leaves that are rarely longer than 2 mm. Another typical feature is that leaves, especially in smaller twigs and near twig apices, are imbricate. *Otovicia* specimens with three orders of branching have been described. Also similar to the Alamogordo specimens is *Thucydia mahoningensis* Hernandez-Castillo, Rothwell, and Mapes (2001), described from the earliest Late Pennsylvanian of the Central Appalachian Basin in the eastern United States. The vegetative shoots of this plant bear ultimate leaves that are short, distinctly triangular in shape, broad relative to their length, imbricated, and addressed to the stem. The lateral branches have three orders of branching, and the female reproductive organs are borne in fertile areas rather than in distinct cones. No conifer remains (apart from the needle-leaved cordaitalean with *Cordaitanthus*-type reproductive organs) demonstrating fertile zones were found among the Alamogordo collections, and neither were lateral branches with three orders of branching. Thus, an identity with this species

cannot be established on either gross morphology or association with reproductive organs, despite similarities in ultimate branch morphological features.

Less commonly encountered is a walchian conifer (Figure 86E,F) with leaves that on the ultimate shoots make an angle with the stem of about 45° and, in combination with relatively wide spacing, give the immediate impression of being nonimbricate. On the basis of a few ultimate leaves preserved in face view, they are lanceolate with an obtuse apex, only slightly sinusoidal in shape but with a recurved somewhat acuminate tip, up to 2 mm long and 0.3 mm wide. The penultimate leaves are lanceolate with an unforked apex and up to 6 mm long and 1 mm wide. *Walchia garnettensis* (Florin) Clement-Westerhof (Winston, 1984) is morphologically similar and is known from similar-aged strata (Missourian) in the United States. Both ultimate and penultimate leaves are simple, but the ultimate foliage of *W. garnettensis* may be too dense for a certain identity in the absence of cuticular evidence. Another described species that resembles this form is *Walchia piniformis* (= *Utrechtia floriniformis* Rothwell and Mapes). However, the subdued sinusoidality of the leaves distinguishes the Alamogordo specimens from what is typically considered *W. piniformis*; furthermore, *W. piniformis* was reported to have forked penultimate leaves, unlike the Alamogordo specimens. A third point of comparison can be made with *Emporia lockardii* Mapes and Rothwell (Hernandez-Castillo et al., 2009), described from slightly younger (Virgilian, Late Pennsylvanian) strata from the Hamilton Quarry of Kansas. There is a general similarity between this species and the foliage reported from Alamogordo. However, *E. lockardii* is documented to have forked leaves on the penultimate branches; reproductive structures and anatomy also are known for this species. Thus, as with other more fully known conifers, we are at the disadvantage of having only gross vegetative morphology, also lacking the cuticle, on which to base comparisons. In this instance, the differences and information deficit require caution in suggesting a strong link with *Emporia*.

The third walchian morphotype, which was represented by only two specimens from a single locality, are ultimate shoots that were part of a plagiotropic branch system similar to *Culmitzschia speciosa* (Florin) Clement-Westerhof (Figure 86G). In these specimens, the leaves make a variable angle with the stem, in many exceeding 45°. The leaves are lanceolate with an obtuse apex, varying from 7–8 mm in length at the base of the shoot to 4 mm near the apex, and 1–1.5 mm in width. The leaves are distinctly sinusoidal in side view with reflexed tips. The bases of the leaves are strongly decurrent and closely ensheath the stem for, perhaps, as much as 1/3 of their length.

Also found were a few specimens of uncertain affinity and not well preserved, possibly comparable in morphology with *O. hypnoides* or *Walchia piniformis* Schlotheim ex Sternberg, but with morphology similar in some respects to “Morphotype A” of DiMichele et al. (2017b), which is similar to *Ernestiodendron filiciforme* (Sternberg) Florin. The cuticle was not present on any of these specimens. Consequently, their attribution to

FIGURE 86. Walchian conifers. (A) Cf. *Otovicia hypnoides*; note adjacent marattialean foliage. USNM specimen 713174, USNM locality 43881. (B) Cf. *Otovicia hypnoides*, including principal branch axis. USNM specimen 706317, USNM locality 43891. (C) Cf. *Otovicia hypnoides*. USNM specimen 713191, USNM locality 43838. (D) Cf. *Otovicia hypnoides*. USNM specimen 713189, USNM locality 43838. (E) *Walchia* cf. *piniformis*. USNM specimen 706313, USNM locality 43891. (F) *Walchia* cf. *piniformis*. USNM specimen 706310, USNM locality 43891. (G) *Culmitzschia speciosa*. USNM specimen 706335, USNM locality 43891. Scale bars = 1 cm.

the natural genera *Otovicia* Kerp, Poort, Swinkels, and Verwer, *Walchia* Sternberg, or *Culmitzschia* Ullrich, as envisioned by Visscher et al. (1986) or Kerp et al. (1990), cannot be established with certainty.

Conifer Reproductive Organs. Three partially intact ovuliferous cones of conifers come from USNM locality 43868A, the second-largest excavation of the Alamogordo collection suite, from the upper part of the lower member of the Beeman Formation. Interestingly, walchian conifers are not abundant in any of the four excavations made at this site (Table 4); the most abundant coniferophyte is *Cordaitea* sp., followed by the new conifer-like cordaitalean. (Pen)ultimate branches of walchian conifers occur at a frequency of no more than 5.5% in any of the four collections and are entirely absent from one. Two of these collections are relatively large, which increases the chances of finding rarer elements.

Specimen 706318 (Figures 87A, 88) is a likely immature ovulate structure of a conifer. The structure is preserved without apex and thus cannot be determined to be a fertile zone or a terminal cone. There is, however, what appears to be an ultimate, vegetative branch attached at the bottom of the specimen. That branch bears walchian-type recurved, sinusoidal leaves that are relatively closely imbricated and is about 1.5 mm in length. Its leaves and organization are similar to those of *Otovicia hypnoides* (Figure 88). The presence of this branch indicates that the base of the structure is present. The structure's main axis, 5 mm in diameter, is a continuation, without evident morphological or anatomical change, from the vegetative axis below. The total width of the cone or fertile zone cannot be meaningfully measured because of its loose organization and the laxity of the bracts. The bracts are relatively widely spaced and support ovuliferous dwarf shoots, some of which may bear an ovule/seed. The bracts are sinusoidal in side view, narrow triangular in face view, about 5 mm in length, and slightly over 1 mm in width, with clasping, decurrent bases and sides that converge on an acute to subobtusate apex. The bracts are lax, and most are disposed horizontally to the axis throughout the majority of their length. In the axils of several bracts, there appear to be flattened, triangular structures, tapered at the base and expanding to a maximum width of about 1.7 mm through a length of about 6.7 mm. These are most likely dwarf shoots. One, possibly mostly complete, dwarf shoot is visible in

Figure 87A, immediately below the dwarf shoot marked by the white arrow. This dwarf shoot has a fanlike, multilobed form. There are indications of seeds on several of the dwarf shoots. One seed in particular is relatively well preserved (Figure 87A at arrow). It is oval in shape and has an offset nucule, which is located in the distal half, thus suggesting that the seed has a recurved orientation, its micropyle directed toward the axis. The pith of the fertile axis is broad and irregularly septate. This architecture compares well with such anatomically preserved coniferophyte stems as *Giblingodendron aridus* Falcon-Lang, Kurzawe, and Lucas, which has been described from strata of nearly identical age in central New Mexico (Falcon-Lang et al., 2016) and of early Permian (Wolfcampian) age in southern New Mexico (Falcon-Lang et al., 2014).

Specimens 699118 and 699119 (Figure 87B–E) are morphologically, and likely taxonomically, identical ovuliferous cones but of uncertain further affinity. These cones are borne at the end of presumed penultimate vegetative branches, which, as in the specimen illustrated in Figure 87B, have vegetative leaves closely adpressed against the axis. This morphology is similar to, but more exaggerated than, the leaf disposition seen in penultimate vegetative branches of walchian conifer specimens from Alamogordo and suggests that the cones were subtended by a peduncle and had an upright posture on the parent plant. The strobili are narrow, 1.25–1.5 cm in diameter and at least 10 cm in length, with a central axis approximately 2.5 mm in width. They are of compound structure, composed of linear bracts about 1.5 cm (maximum of 2.8 cm) in length and 0.75–1.3 mm in width, with ovuliferous dwarf shoots in their axils. The dwarf shoots are not fully preserved, so their morphology cannot be fully described. They appear to be flattened and scale-like, consisting of multiple lobes (10 or more) that are laterally fused through most of their length but appear to be free at their apices (best seen at the base of Figure 87C). Both bracts and dwarf shoots are borne at narrow angles to the cone axis, giving the strobilus a very compact appearance; the bracts are exserted and ensheath the strobilus. No evidence of attached ovules was noted, which, in combination with the dense covering of bracts and size, suggests that these are immature structures.

Some conifers have leaves that are similar in size and shape on both ultimate and penultimate branches, whereas others

FIGURE 87. (*Opposite*) Conifer strobili and bracts. (A) Walchian strobilus, possible seed at arrow. USNM specimen 706318, USNM locality 43891. (B, C) Walchian strobilus. USNM specimen 699118, USNM locality 43868A. (D, E) Walchian strobilus. USNM specimen 699119, USNM locality 43868A. (F) Possible walchian cone scale. USNM specimen 713228, USNM locality 43839. (G–J) Vase-shaped, small bracts or penultimate axis leaves. USNM specimen 713203, USNM locality 43870; USNM specimen 699116, USNM locality 43868A; USNM specimen 706322, USNM locality 43891; USNM specimen 706321, USNM locality 43891, respectively. (K–M) Typically leaflike bracts or penultimate axis leaves. USNM specimen 699117, USNM locality 43868A; USNM specimen 706324, USNM locality 43891; USNM specimen 706323, USNM locality 43891, respectively. (N) *Gomphostrobus*, conifer bract or penultimate axis leaf. USNM specimen 713238, USNM locality 43891. Scale bars = 1 cm.

FIGURE 88. Conifer strobilus with vegetative axis attached at base. This is an enlargement of the specimen illustrated in Figure 87A. USNM specimen 706318, USNM locality 43891. Scale bar = 1 cm.

differ in size (respectively small and large) and/or shape (respectively simple and forked, *Gomphostrobus*-type leaves). Therefore, it may not be possible to distinguish isolated penultimate scale vegetative leaves from strobilar bracts. The similarities are not surprising because the ovuliferous cones are homologous to a branching system in which the axis of penultimate order is the cone axis, the bracts are in the position of penultimate vegetative leaves, and the dwarf shoots, strongly reduced, shortened, and flattened, are homologous with twigs of ultimate order. These homologies are best illustrated in the early diverging conifers where the individual sterile scales and sporophylls of the dwarf shoots are still free and, although flattened, are more or less helically arranged on a short axis.

We offer this preamble as the introduction to the remaining examples of conifer scales and bracts, most of which cannot be said, for certain, to be from reproductive structures. USNM specimen 713228 (Figure 87F) is the only example of an isolated walchian dwarf shoot; it was recovered from the Beeman Formation, in the third largest of our excavations. It is >6 mm in length and 5 mm in width and consists of multiple scales that are fused at the base. The remaining bracts/large leaves fall into three categories. The first (Figure 87G–J) consists of vase-shaped laminae with an oval basal 1/3–1/2, topped by a tapering triangular distal lamina and a subobtusate apex. A distinct vertically elongate oval scar occurs near the base and a single medial vein runs from the top of the scar to the apex of the bract. A small, sharply pointed keel-like extension occurs below the basal oval region. The bracts are about 1.3 cm in length and 3–4 mm in maximum width. The second group (Figure 87K–M) encompasses more typically leaflike forms that are much longer than wide and gradually taper from a slightly swollen base, with a horizontally elongate leaf scar, to a subobtusate apex. A medial vein runs the length of the bract, beginning above the basal scar. The largest dimensions observed (Figure 87K) were 4 mm basal width and >2 cm length. The final group, which was the rarest, has *Gomphostrobus* morphology (Figure 87N). The illustrated specimen is 2.4 cm in length and 5 mm wide at the base, which is distinctly shaped like an inverted triangle. The apex is forked, the division of the lamina taking place about 2/3 of the distance from the base. This type of bract is typical of the penultimate axes, cones, and fertile zones of most walchian conifer taxa.

DICRANOPHYLLUM Grand'Eury. Among the more enigmatic plants found in the Alamogordo flora are fragmentary specimens, generally of forked leaves, that we assign to *Dicranophyllum*. The affinities of the dicranophylls are uncertain, other than that they are gymnosperms, likely in the coniferophyte lineage, and perhaps even within the conifers themselves (Cleal, 1993; Rothwell et al., 2005; see review of their taxonomy in Van Der Pas et al., 2017). They have been reconstructed as bottlebrush-like plants, with small, unbranched or rarely branched, slender stems bearing long, stiff leaves that fork one to several times (Barthel et al., 1998; Barthel and Noll, 1999; Van Der Pas et al.,

2017). The leaves have a single central vein and, on the abaxial side, a pair of well-developed stomatal furrows. There are a number of described species that together give the lineage a range from the Late Mississippian into the early Permian; dicranophylls are particularly common in the Late Pennsylvanian and Permian portions of that range (Wagner, 2005; Barthel and Rößler 2012). Their rarity and common occurrence with other taxa generally not considered part of Pennsylvanian wetland assemblages led Wagner (2005) to suggest a preference for nonwetland habitats. Van Der Pas et al. (2017) suggest them to have been opportunistic strategists, capable of colonizing a wide range of environmental settings; even if that is so, the ecological centroid of the plants appears to have been in areas experiencing more seasonal environmental conditions. Such a preference is consistent with their more common occurrence in floras in western Pangea, contemporaneous with those from the central Pangean coal basins, in which dicranophylls are nearly unknown.

The Alamogordo specimens (Figure 89A,B) are fragmentary and do not have all the characteristics one would wish to see in order to make a firm identification. In most, the venation is not clearly visible, although a midvein can be seen in the right-hand fork of the specimen illustrated in Figure 89B. The absence of the distinctive paired stomatal grooves suggests that these specimens expose adaxial surfaces. Just how long these leaves might have been also is uncertain, but they probably were in the range of 10–20 cm in length, perhaps more, given the size of the fragments.

Wagner (2005) notes that the large number of described species may be artificial, given that most are based on fragmentary leaf material alone, so we confine our comparison to three species. On the basis of their gross morphology and their Late Pennsylvanian age, these are *Dicranophyllum hallei* Remy and Remy, *D. gallicum* Grand'Eury, and *D. readii* Mamay. Of these three, the leaves of *D. gallicum* are the shortest, commonly much less than 10 cm in length, and only a few millimeters in width, too small for a likely match with the Alamogordo specimens (cf. Barthel, 1977; Van Der Pas et al., 2017). In the opposite direction, *D. readii* leaves are significantly longer than the dimensions we project for the Alamogordo specimens; Mamay (1981) reported leaves 75 cm or more in length. *Dicranophyllum readii* is described, mostly from the Kinney Quarry of central New Mexico, in strata of late Missourian age (Lucas et al., 2011), thus closely similar in space and time to the Alamogordo occurrences. Finally, *Dicranophyllum hallei*, at least on the basis of size, is roughly comparable to the specimens illustrated here (Barthel et al., 1998) and may be the best guess for a species affinity.

Note the presence of microconchids on the surface of the leaf illustrated in Figure 89B. These almost certainly indicate that the specimen remained suspended in the water column for some period of time. The salinity tolerance range of microconchids has been debated (Gierlowski-Kordesch and Cassle, 2015; Zatoń et al., 2016), but evidence appears to indicate broad tolerance from near marine to nonmarine salinities.

FIGURE 89. *Dicranophyllum* sp. (A) USNM specimen 699122, USNM locality 43868A. (B) USNM specimen 706339, USNM locality 43891. Note the numerous microconchids/spirorbids on the leaf surface. (C) USNM specimen 706319, USNM locality 43891; this specimen is also similar to *Dichophyllum flabellifera*. Scale bars = 1 cm; note that C is at 1/3 lower magnification than A and B.

Dicranophyll-Like Plants. Several additional specimens merit comparison with the dicranophylls. The first, illustrated in Figure 89C, is representative of a small number of specimens, all found at USNM locality 43891, the largest single excavation carried out in the area. These forked, linear structures most likely represent the tips of multiforked *Dicranophyllum* leaves. Close inspection of the laminae on the left of Figure 89C reveals the presence of two parallel lines running the length of the laminae, which may be paired stomatal grooves, typical of *Dicranophyllum*. Alternatively, the several specimens found of this type may be fragmentary remains of *Dichophyllum* Elias ex Andrews, possibly *D. moorei* Elias, which Elias (in Moore et al., 1936) noted has pinnae of palmate form (which most likely would produce fragmentary pinna tips as seen in the Alamogordo remains). Considering the overall evidence, however, this seems a less likely identification than *Dicranophyllum* cf. *hallei*. A final possibility is that these specimens are leaf tips of a ginkgophyte-like plant, such as *Psymgophyllum* Schimper, one suspected specimen of which was found in the Alamogordo collections (Figure 90) from the same locality as the fragments represented by Figure 89C. This is the least likely of the three possibilities considered here.

A single specimen with *Psymgophyllum*-like morphology (see Naugolnykh, 2007) was found at excavation USNM 43891. The specimen is a single leaf, narrow at the base and expanding outward with gradually arcing lateral margins. The base is fractured, and thus, the exact architecture is not known. The apical region is also missing because of fragmentation of the specimen. Between these points, however, the leaf can be seen to be of ginkgo-like shape, with a deep central cleft, extending through 2/3 of the leaf; each of the two resulting laminar segments is again divided by a median cleft, about 1/3 of the distance from

the preserved top of the specimen. The lamina itself is plicate, with each plication appearing to be longitudinally traversed by a single raised area, possibly a leaf vein. Portions of the laminae are vitrified (note the high reflectivity of the organic matter along the lower side of the specimen, particularly visible in Figure 90B), suggesting a relatively thick, organic-rich architecture. Although this specimen could not be confused with *Dicranophyllum*, it is similar in having elongate segments with single median veins, the segments likely separating and being individualized apically, thus the possible similarity to the specimens illustrated in Figure 89C. We include this plant within the coniferophyte section of this paper because in some seed-plant phylogenies, the ginkgophytes are included among the broader coniferophyte group (e.g., Doyle, 2013).

A third specimen (Figure 91), the only example of this type, was found in the same beds of the Beeman Formation as the other specimens described above in this section and from the second-largest specimen collection, USNM locality 43868A. It consists of an axis bearing lateral appendages, which we presume to be leaves. The axis is approximately 1 cm in diameter and is longitudinally striated and, where its substance is preserved, thickly carbonized. Furthermore, the central axis is covered with relatively closely spaced subhorizontal raised lines, which can be seen on the remains of the organic matter (they appear as indentations on the mold surface; note that we have not further prepared and exposed this stem because of the delicacy of its carbonized structure, thus preserving it for future analyses). We interpret these latter features as resulting from compression, dewatering, and shrinking of the stem. Three lateral appendages are present, all on the same side; the unequal spacing of these appendages suggests that many are not preserved and that they were likely helically arranged. These

FIGURE 90. Cf. *Psymgophyllum* ginkgophytes or dicranophylls. Both images of USNM specimen 713124, USNM locality 43891. (A) Specimen at natural size. (B) Specimen enlarged to show base and central cleft. Scale bars = 1 cm.

FIGURE 91. *Cf. Dicranophyllum* sp. All images of specimen 713124, USNM locality 43836A. (A) Entire specimen at natural size. (B) Main axis showing attachments of lateral appendages. Upper arrow points to a nonattached leaf with paired lines running its length. Lower arrow points to a microconchid/spirorbid attached to the leaf lamina. (C) Base of main axis showing woody texture. Scale bars = 1 cm.

lateral appendages are unforked and reach a width of approximately 1.5 mm and a maximum length of approximately 3.5 cm, with unexposed tips. The appendages also must have been rather thick, given their flaking carbon surfaces. The carbon of the stem expands outward at the attachment point of the appendages, suggesting that, originally, there may have been further surface features on the stem forming raised areas to which the leaves were attached. Given this overall morphology, it seems possible that this is a specimen of *Dicranophyllum*, in the size range of *D. halei* and, to a lesser extent, *D. gallicum*. The appendages would, in this interpretation, represent the basal portions of *Dicranophyllum* leaves, which are generally some centimeters in length prior to the first dichotomy. Unfortunately, no details, such as a midvein or stomatal grooves, are visible on any of the three presumed leaves. Near the top of the image, however, marked by an arrow in Figure 91B, there is an unattached narrow structure with the shape of the attached leaves. Close examination of this presumed leaf reveals the presence of two parallel lines or ridges, which may be the stomatal grooves present on the abaxial side of *Dicranophyllum* leaves. Note also on this specimen the presence of a microconchid/spirorbid near the preserved outer edge of the lowermost leaf (Figure 91B at arrow) and the numerous ostracods present in the matrix (small, white, bean-shaped objects).

INCERTAE SEDIS

SPHENOPTERIS GERMANICA Weiss. We place *Sphenopteris germanica* among incertae sedis. Although its affinities are unclear, there are indications in its morphology that it may belong to the pteridosperms. Originally thought to be characteristic of the Permian (e.g., Remy and Remy, 1959; Remy, 1975; Pfefferkorn and Resnik, 1980; Barthel, 2006), this species has been found, particularly in western North America, well down into the early Late Pennsylvanian. Perhaps the best-documented pre-Permian occurrence of this taxon is what Mamay (1992) described as *Sphenopteridium manzanitanum*, from the Missourian-age Kinney Quarry in New Mexico. The plant appears to have lived in seasonally dry environments and probably formed large stands on the basis of mass occurrences in some taphofloras. Furthermore, the habit of the plant may have been quite different from what has been envisioned for most late Paleozoic plants. The plant described as *Arnoldia kuesii* Ash and Tidwell (1986) is likely a juvenile of *S. germanica* on the basis of the discovery of much larger specimens with the same growth architecture from the Kinney Quarry: a bulbous stem from which a crown of large leaves emerged and isolated leaves over a meter in length (undescribed specimens in the Smithsonian collections).

Identification of *Sphenopteris germanica* in the Alamogordo collections was facilitated by the occurrence of several larger leaf fragments (Figure 92A,B), which added confidence to the identification of smaller fragments that otherwise can be difficult to differentiate from species of small ferns (e.g., Figure 92D). Details of the foliage are illustrated in Figure 92C,D. Differentiation of pinnules and pinnae is difficult and depends on the specific location

on the leaf from where the specimen was drawn. Pinnules generally are strongly decurrent and fused with one another to some degree. This fusion increases toward the apex of any given pinna and toward the apex of the leaf, in general. When clearly individuated, pinnules are elongate and spatulate, with curving lateral margins and a broadly rounded apex. They may be lobed 3 to 7 times, with the central lobe being the largest. In some instances, these lobes may be sufficiently distinct to be considered pinnules. Multiple veins enter the pinnule lamina. There is no clear midvein. The veins are steeply disposed in the pinnule lamina and proceed in slightly diverging paths to the pinnule apex; although it may appear that some veins contact the margin below the apex, this does not appear to be the case, but rather is a happenstance of the broadly rounded shape of the terminal margin. Pinnae are inserted on the penultimate rachis in a subopposite and alternate manner.

Sphenopteris germanica occurs at 7 of the 19 Alamogordo collecting sites. In two instances, both from the lower Beeman Formation organic shale beds, it reaches frequencies of >20% and is, as such, the most abundantly encountered plant in those collections. It is typically found as scraps (Figure 92C,D), but not uncommonly, it is found as relatively larger leaf fragments. This plant has now been found in most of the Pennsylvanian-age deposits we have examined in New Mexico and Texas (e.g., Tabor et al., 2013; DiMichele et al., 2013, 2017a, 2017b), suggesting that it was widespread throughout the western parts of Pangea, perhaps before spreading eastward into lowland basins of present-day Europe during the Permian, accompanying the general west-to-east drying trend of that time.

TAENIOPTERIS Brongniart. *Taeniopteris* is an enigmatic genus of foliage that almost certainly is polyphyletic (not a natural taxon), although likely containing monophyletic groups of morphotypes within it. The foliage is generally elongate-tongue shaped overall, the leaf lamina tapering toward the base, marked by an arcuate abscission zone, and tapering less gradually near the apex, which varies in shape among species but often is somewhat bluntly rounded. Many species of the genus have been described, varying in aspects of base or apex shape, the angle of vein branching, vein density, and other features of gross morphology (Remy and Remy, 1975). The earliest forms we are aware of were described by Bashforth et al. (2016) from the Middle Pennsylvanian of the Illinois Basin, in association with deposits of seasonally dry landscapes, occurring between periods of peat/coal formation. The types of *Taeniopteris* found in the Alamogordo collections most likely belong to those forms related to cycadophytic seed plants (see Mamay, 1973, 1976; Yan et al., 2017).

Taeniopterids occur at two of the seven stratigraphic levels from which plants were collected in the Missourian Beeman Formation and in the single Virgilian Holder Formation collection. They are relatively uncommon in all of these collections. In all instances, they were variously fragmented (Figure 93A–E). Some, such as the specimen illustrated in Figure 93B, had microconchids attached to the lamina surface, similar to cordaitalean foliage; laminae of *Taeniopteris* leaves could be several centimeters in width, contributing to the likelihood that they floated prior to

FIGURE 92. *Sphenopteris germanica*. (A) USNM specimen 713144, USNM locality 43836C. (B) USNM specimen 713145, USNM locality 43836C. (C) USNM specimen 713146, USNM locality 43836C. (D) USNM specimen 706297, USNM locality 43836A. Scale bars = 1 cm.

burial in sediment. Laminae frequently are torn between the veins, giving specimens an almost pinnate appearance (Figure 93A). In general, the specimens fall into two groups based on leaf-lamina width, wider and narrower. However, vein density varies, as can be observed in the specimens illustrated in Figure 93. Veins depart the midrib steeply but over a very short distance curve to a nearly horizontal plane. They run in a nearly straight course across the lamina and curve upward sharply just before meeting the margin. If veins fork, branching tends to be confined to the area close to the midrib. Near the apical regions of a leaf, the vein angles tend to increase and may account for the relatively narrow lamina width and higher (but still low) angle veins found in the specimens illustrated in Figure 93D,E. The fragmentary and somewhat battered nature of the specimens suggests a robust construction, giving the leaves the capacity to survive rigors of transport and perhaps long exposure times in the oxygenated water column.

Seeds. The classification of late Paleozoic dispersed, adpressed ovules and seeds, which we will refer to further simply as “seeds,” is, compared to foliage, relatively little explored for several reasons (but see Yehnjong et al., 2017). Seeds are most often found dispersed from the parental plant, and their affinities

with foliage taxa generally cannot be established with certainty. This uncertainty is exacerbated by the often great differences between seeds found in sediments and the spectrum of seeds produced by the local vegetation (e.g., Sims and Cassara, 2009), by the tendency for seeds in aquatic sediments to reflect primarily water transport and thus to encompass a species pool larger than the local community (e.g., Scheihing and Pfefferkorn, 1984; Van der Burgh, 1994), and by overrepresentation of certain dispersal syndromes (e.g., Vassio and Martinetto, 2012). Additionally, preservational happenstance can lead to many different morphological expressions of what likely are seeds from the same parental species (e.g., Scheihing and Pfefferkorn, 1984). Finally, as with pollen and spores, developmental and physiological circumstances can create intraspecific morphological variability in seeds from the same parental plant, but the nature and degree of such variability have not been studied broadly. Consequently, we have classified the seeds described here first as morphotypes. There are few comprehensive compilations of late Paleozoic seeds, the most notable being that of Crookall (1976).

Here, we present representatives of every seed morphotype encountered in the Alamogordo collections. All are presented at

FIGURE 93. *Taeniopteris* sp. (A) USNM specimen 698905, USNM locality 43891. (B) USNM specimen 699102, USNM locality 43868A. (C) USNM specimen 698928, USNM locality 43836C. (D) USNM specimen 699103, USNM locality 43868A. (E) USNM specimen 713247, USNM locality 43430 (Holder Formation). All images at same magnification; scale bar in A = 1 cm.

the same, relatively high magnification to facilitate comparison and to emphasize the large size range encountered. The stratigraphic ranges encompassed by these seed morphotypes cannot be treated as a reliable indicator of their temporal distributions. The great variation in sample size likely plays a major role in the number of seed specimens encountered, and there is a clear positive relationship between those numbers (Figure 94). Furthermore, because seed production can be strongly seasonal or periodic in its temporal

distribution, the sedimentary origin of the fossiliferous strata can have a major filtering effect; for example, the seed population in rapidly accumulated sediments, such as sands and silts deposited in active channels, is likely to be a much narrower representation of seeds produced by parental plant communities than that of a more slowly accumulated deposit, such as a lake or lagoon.

We have identified approximately 28 seed morphotypes (Table 5). The qualifier “approximately” reflects the fact that

FIGURE 94. Plot of the relationship between sample size from a given excavation and the number of distinct seed morphotypes found in that excavation.

some of these morphotypes are similar in form but differ in size or in subtle aspects of morphology. In the absence of anatomy and given the vagaries of preservation, there could be more or fewer morphotypes than we estimate. Fourteen of the morphotypes appear to be platyspermic and in almost every instance have a wing, symmetrically developed relative to a line bisecting the seed from apex to base. Some of these seeds, but not necessarily all, are likely derived from coniferophytic plant groups (cordaitaleans and conifers). Nine morphotypes appear to be radiospermic on the basis of the shape of their core region. None of these are winged, but several kinds appear to have had a relatively thick sarcotestal layer that has been preserved as a coaly zone, often developed asymmetrically, possibly because of compression. Furthermore, some of these seeds are ribbed, as is often typical of pteridosperm radiospermic seeds. In five instances, the symmetry of the morphotype could not be determined with certainty, although, on the basis of the majority of the characteristics, we suspect that these seeds were radiospermic.

Seed morphotypes are illustrated in Figures 95–100. Those from the two most extensively sampled locations are from the Beeman Formation organic shale (Figure 96) and from the Beeman Formation brackish lake/lagoonal deposit (Figures 97–99). The morphotype group to which the seed is assigned is indicated by the number on each image. A question mark indicates uncertainty in the assignment.

Of the nine stratigraphic levels sampled, seeds are present in seven and absent from the smallest of the collections made in the

upper Gobbler Formation (USNM 43879) and from one Beeman Formation collection (USNM 43871) in which plant material was allochthonous and highly fragmentary or comminuted. At the level of individual excavations, seeds were recovered from 15 of the 19. Of the 28 morphotypes, only 7 were found at more than one stratigraphic level, leaving 21 of the morphotypes confined to a single level. Across the seven stratigraphic horizons from which seeds were collected, the proportion of such unique occurrences ranges from 45% to 100%. Furthermore, when compared against recently published papers in which we have illustrated seed spectra (DiMichele et al., 2017a, 2017b, 2018a, 2018b), about 60% of the Alamogordo morphotypes can be matched to similar forms (Table 6). Comparable seeds are found most frequently, as might be expected, among Pennsylvanian floras of identical age and from the same region of western tropical Pangaea (DiMichele et al, 2017a, 2017b). These patterns, more than any other, indicate the power of seeds to capture species richness while, at the same time, being strongly affected by a host of taphonomic factors.

Platyspermic and radiospermic seeds occur, on a presence/absence basis, in about equal frequency; there are 22 platyspermic occurrences, and if radiospermic and uncertain occurrences are combined, 18 of these. The ratio of platyspermic to radiospermic morphotypes at each stratigraphic level shows no systematic bias in either direction; in the largest collection (862 quadrat surfaces), however, there are >2:1 platyspermic morphotypes.

FIGURE 95. Seeds from the Beeman Formation. (A–D) USNM specimens 713234, 713235, 713233, 713237, respectively, USNM locality 43839. (E, F) USNM specimens 713209, 713208, respectively, USNM locality 43870. The objects in images D and F may be polleniferous organs rather than seeds. All specimens oriented with presumed micropylar region up. All images at same magnification. Scale bar (in A) = 1 cm.

Included here are some observations that have been made on the population of seeds examined: (1) Many of the seeds, especially those of platyspermic form, have a narrow V-shaped notch at the micropylar end. This notch may indicate that these are husks of seeds that had already germinated. The V-shaped region then would mark the point of emergence of the seedling. This characteristic can be seen in some seed plants today. (2) A few of the seeds (examples illustrated in Figures 97A,O and 98A,C and, to some extent, Figure 98E) are surrounded by narrow zones in which the sediment color has been altered. These zones are referred to as “reduction halos” and likely indicate the reduction of iron in the sediment through bacterial action, related to decay of the outer tissues of the seed. They are present only in a small number of seeds, which is surprising and may indicate that most specimens were buried in sediment relatively rapidly, below the aerobic zone. (3) Seed surfaces characterized by dense, longitudinally disposed wrinkles (e.g., Figures 98F, 99A) may represent the effects of dehydration of a once fleshy outer covering. (4) Morphotypes 7 and 12 are good examples of high morphological convergence but exhibit considerable variation in size. Morphotype 7, in fact, is similar to the seed illustrated by DiMichele et al. (2017a) in their fig. 68.1, which is even somewhat smaller than those illustrated from Alamogordo. (5) In the majority of winged seeds, the wing is bilaterally symmetrical and surrounds most of the central nucule, likely contributing to a gliding descent pattern. In some, a continuous wing is most

strongly developed at the chalazal end (Figures 97J, 98A–C); this construction may have focused the micropylar end, from which germination was to occur, into the ground upon landing. Some of these seeds with strongly developed chalazal wings show tendencies toward the development of double wings, especially morphotype 18 (Figure 97G–I), which may have contributed to autorotating flight, thus enhancing dispersal distance from the parent plant (see discussion in Looy and Stevenson, 2014; Stevenson et al., 2015). (6) The seed profile from the several Alamogordo fossiliferous horizons is notable in its paucity of small, unadorned, radiospermic seeds in the size range of <5 mm length and <3 mm width. This profile may reflect the fact that these seeds do not float well and were not readily dispersed into large bodies of water. Although most Alamogordo seed morphotypes are <1 cm in greatest dimension, they, nonetheless, do not include many examples of the smallest seed sizes known from this time period (Sims, 2012). (7) Many of the seeds have attracted the fine roots of modern plants, which have spread networks across the seed surface, scavenging the original organic matter. A good example of this is provided by Figure 98C.

Specimens of Uncertain Affinity. Numerous generally small and often fragmentary specimens of uncertain affinity were found among the Alamogordo plant fossil collections. Several of these are illustrated in Figure 101. These structures are discussed briefly here.

The specimen illustrated in Figure 101A may be a conifer cone scale, given its several lobes, the seeming fusion of those lobes in the lower 25% of the structure, and its apparent flattening into a single plane. No seed attachment scars are visible, leading to the conclusion that the view represents the abaxial surface.

Figure 101B may be one or several independent structures. These could be lobes, perhaps, of a polleniferous organ. Or they could be immature seeds comparable with *Samaropsis crampii* Dawson, with which they share overall morphological attributes of general shape, a small central nucule surrounded by an elongate, relatively broad wing, and a median keel running the length of the wing. They are, however, significantly smaller than is characteristic of this species, 1/3 the characteristic length. They may, in that case, be a new species closely allied with *S. crampii* evolutionarily. A similar example in this same size range was illustrated by DiMichele et al. (2017a: fig. 68.7) from strata of the same age in Socorro County, New Mexico.

The specimens illustrated in Figure 101C,D are suspected to be reproductive organs. Their affinities, however, are not known to us, nor can we offer any suspected candidates. The small size of the specimen in Figure 101C, <1 cm in maximum dimension, and its organization suggest that it might have originally been part of a larger aggregation of such organs. The oval ring of small, radially elongate objects illustrated in Figure 101D brings to mind the reproductive organs of peltasperms and suggests a possible microsporangiate organ. Again, however, the specimen is very small, well below the typical size of *Peltaspermum* Harris or *Autunia* Krasser emend. Kerp.

FIGURE 96. Seeds from several excavations in the Beeman Formation organic shale, USNM 43834 and 43836A-F. USNM specimen number of illustrated specimen, followed by the last two digits and letter, where appropriate, of the source excavation: (A) 713115 (36A), (B) 713183 (34), (C) 713128 (36C), (D) 713184 (34), (E) 713150 (36E), (F) 713182 (34), (G) 713130 (36C), (H) 713129 (36C), (I) 713154 (36C), (J) 713131 (36E), (K) 713151 (36C), (L) 713136 (36C), (M) 713133 (36E), (N) 713152 (36E). All specimens oriented with presumed micropylar region up. All images at same magnification. Scale bar (in A) = 1 cm.

TABLE 5. Distribution of seed morphotypes by stratigraphic level in the Alamogordo collections. USNM loc = USNM locality numbers; Tot = total morphotypes in all collections from that stratigraphic level; Uq = number of morphotypes unique to that stratigraphic level; %Uq = proportion of the seeds at that stratigraphic level that occur only there; Platy:radio = the ratio of platyspermic to radiospermic seeds at that stratigraphic level; Smpl size = sample size computed as described in the text. An X indicates a morphotype is present; a dash (—) indicates it is absent.

USNM loc	Morphotype																												Platy: Smpl				
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	Tot	Uq	%Uq	radio	size
43430	—	—	—	—	—	—	—	—	—	—	—	X	—	—	—	—	—	—	—	—	—	—	—	—	X	—	—	—	2	1	50	0:2	20
43838	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	X	X	2	2	100	2:0	56
43868/69/91	—	X	X	—	—	—	X	—	—	—	—	X	—	—	—	X	X	X	X	X	X	X	X	X	—	—	—	—	13	10	77	9:4	862
43871	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	0	0	0	0:0	34
43881	—	—	—	—	—	—	—	43881	—	—	X	X	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	2	0	0	1:1	28
43880	—	—	X	—	—	—	—	43880	—	—	X	—	X	—	—	—	—	—	—	—	—	—	—	—	—	—	X	—	4	2	50	2:2	34
43839	X	X	X	X	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	4	2	50	2:2	103
43834/36	—	—	X	—	—	—	X	X	X	X	X	X	X	X	X	X	—	—	—	—	—	—	—	—	—	—	—	—	11	5	45	5:6	288
43870	—	—	—	—	X	X	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	2	2	100	0:2	28
43879	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	0	0	0	0:0	8

The specimen illustrated in Figure 101E appears to be an axis bearing leaflike lateral appendages that either are short, only a few millimeters in length, or have been broken. The arrangement of these leaflike lateral organs cannot be determined. Just above, arguably in the axil of, the lower appendage is an attached seedlike structure, similar in its size, shape, and dimpled surface ornamentation to seed morphotype 14 (Figure 96K). If this is so, it implies a type of loose strobiliar organization, the affinities of which are unclear.

The final unidentifiable structure (Figure 101F) appears to be a stem fragment, one that preserves the surface morphology with leaf scars. The scars are nearly round and have a raised rim, with an acuminate apical extension. The image is oriented with what we believe to be “up” at the top. Each scar has a laterally elongated, oval structure at about 1/3 of the distance from the base, which is likely a leaf trace. Within some of the scars, most notably the one marked by a white arrow, two small prints occur below the presumed leaf trace, on either side, in the same position as parichnos in lycopsid leaf scars. For these reasons, we interpret this specimen as a lycopsid but of uncertain affinities.

PALEOBOTANY: DISCUSSION

The floral collections from the Alamogordo area are among the best preserved and most diverse from the western parts of the former Pangean supercontinent, of primarily Missourian age, in American regional stratigraphic terminology. This is early Late Pennsylvanian, or Kasimovian in international terms. The flora is of decidedly “mixed” character, meaning that it includes elements typical of different soil moisture regimes. Some elements, including marattialean ferns, calamitaleans, and some pteridosperms, are generally considered diagnostic of hygromorphic microhabitats, those with high water tables, even a degree of surface flooding, and few or no periods of significant moisture deficit. Others, certain pteridosperm groups, in particular the odontopterids and *Neurodopteris*, are considered to be tolerant of seasonal moisture regimes but are still tied to soils that have water available throughout most of the year, strongly suggested by medullosan pteridosperm anatomy (Wilson et al., 2008, 2015).

Both of these groups tend to include elements that are among the most hygro- to mesomorphic of Pennsylvanian-age plants (but see commentary such as that in White, 1931, and Raymond et al., 2014, noting the skewing of the entire plant morphological spectrum to more xeromorphic morphologies than is characteristic of modern floras from humid environments). Intermixed with these plants are common to abundant kinds of seed plants that typically are associated with periodically moisture-stressed habitats, such as conifers, some cordaitaleans, taeniopterids, and *Sphenopteris germanica*. Floras of this mixed composition are characteristic of western Pangea throughout the Pennsylvanian and even into the early Permian, and their environmental interpretation and taphonomic origin have been the subjects of long-standing discussion (e.g., Elias in Moore et al., 1936; Chaloner, 1958; Cridland and Morris, 1963; Pfefferkorn, 1980; Winston, 1983; Lyons and Darrah,

FIGURE 97. Winged, platyspermic seeds from excavations USNM 43868A and 43891 in the Beeman Formation, Beeman Canyon. USNM specimen number of illustrated specimen, followed by the last two digits and letter, where appropriate, of the source excavation: (A) 713065 (91), (B) 713076 (91), (C) 713070 (91), (D) 713072 (91), (E) 699133 (68A), (F) 699129 (68A), (G) 713071 (91), (H) 713073 (91), (I) 699135 (68A), (J) 713067 (91), (K) 699135 (68A), (L) 699130 (68A), (M) 713069 (91), (N) 699132 (68A), (O) 713074 (91). All specimens oriented with presumed micropylar region up. All images at same magnification. Scale bar (in A) = 1 cm.

1989; Falcon-Lang and Bashforth, 2004; Cazzulo-Klepzig et al., 2007; Gastaldo and Degges, 2007; Bashforth et al., 2014, 2016; Thomas and Cleal, 2017).

Stratigraphic Implications of the Flora

The stratigraphic distribution of fossil plant collecting localities is shown in Figure 47. The stratigraphic distribution of

species is illustrated in Table 4, which also shows the counts of specimens from each collecting locality.

Overall, the Alamogordo Missourian-age species pool contains elements typical of the Late Pennsylvanian, but its complex taxonomic character, mixing xeromorphic, mesomorphic, and hygromorphic taxa, and the fragmentary nature of many of the potentially important, stratigraphically diagnostic forms make it a problematic basis for stratigraphic determination. Plants

FIGURE 98. Winged, platyspermic seeds, unwinged radiospermic seeds, and seeds of uncertain morphology from excavations USNM 43868A and 43891 in the Beeman Formation, Beeman Canyon. USNM specimen number of illustrated specimen, followed by the last two digits and letter, where appropriate, of the source excavation: (A) 699131 (68A), (B) 699128 (68A), (C) 713068 (91), (D) 699125 (68A), (E) 713075 (91), (F) 699126 (68A). All specimens oriented with presumed micropylar region up. All images at same magnification. Scale bar (in A) = 1 cm.

FIGURE 99. Large radiospermic unwinged and platyspermic winged seeds from excavation USNM 43868A, Beeman Formation, Beeman Canyon. (A) USNM specimen 699127. (B) USNM specimen number 699137. Both specimens oriented with presumed micropylar region up. Images at same magnification. Scale bar (in A) = 1 cm.

and many biological attributes of the ecosystems they comprise are notoriously sensitive to environmental conditions, including those accompanying glacial–interglacial cycles, as amply revealed by any number of published works on Pleistocene and Holocene floras, too numerous to cite extensively (for some examples, see Tzedakis, 1994; Pennington et al., 2004; Gosling et al., 2008; Hanselman et al., 2011; Velásquez-R and Hooghiemstra, 2013; Boisvert-Marsh et al., 2014; Fei et al., 2018). There is no reason not to assume similar sensitivity in the deep past (Elias in Moore et al., 1936; Arnold, 1941; Falcon-Lang, 2004; DiMichele, 2014). Thus, it can be difficult to base highly refined biostratigraphic determinations on fossil floras from biomes different from those on which late Paleozoic Euramerican biostratigraphic schemes have been constructed.

The Alamogordo flora of the Beeman and upper Holder Formations consist of mixtures of plants that range from Middle

Pennsylvanian to Permian in their temporal distributions. Overall, the flora seems to be of Late Pennsylvanian age, but exactly where it may fit floristically in the Late Pennsylvanian is uncertain on the basis of comparison with floras composed largely or entirely of wetland elements (e.g., Cleal, 2007; Wagner and Álvarez-Vázquez, 2010; Opluštil et al., 2016, 2017). Regionally, it compares to some extent with floras from central New Mexico, including that of the Tinajas Member of the Atrasado Formation at the Kinney Brick Quarry (DiMichele et al., 2013), the Burrego Member of the Atrasado Formation at Abo Pass (DiMichele et al., 2017b), and the Atrasado Formation of Socorro County (DiMichele et al., 2017a). All of these also are strongly mixed floras, with abundant conifers, cordaitaleans, *Dicranophyllum* and *Sphenopteris germanica*, intimately intermixed with elements such as *Neuropteris ovata*, *Macroneuropteris scheuchzeri*, calamitaleans, and various kinds of marattialean fern foliage. The

FIGURE 100. Seeds from the Holder and Beeman Formations. (A, B) USNM specimens 713242p, 713243, respectively, USNM locality 43430, Holder Formation. (C, D) USNM specimens 713170, 713171, respectively, USNM locality 43881, Beeman Formation. (E–H) USNM specimens 713181, 713180, 713178, 713179, respectively, USNM locality 43880, Beeman Formation. (I, J) USNM specimens 713199, 713200, respectively, USNM locality 43838, Beeman Formation. All specimens oriented with presumed micropylar region up. All images at same magnification. Scale bar (in A) = 1 cm.

TABLE 6. Comparison of Alamogordo seed morphotypes to published literature. Abo Pass = DiMichele et al. (2017a), Doña Anas = DiMichele et al. (2018b), Kola = DiMichele et al. (2019), Sanzenbacher = DiMichele et al. (2018a), Socorro = DiMichele et al. (2017b). A question mark (?) = uncertainty of identity.

Morphotype No.	Equivalents in the literature
1	No equivalents
2	Sanzenbacher pl. X.11
3	Socorro fig. 68.2; Sanzenbacher pl. X.12, X.13
4	Socorro fig. 69.1, compare with <i>Carpolithus pseudosulcatus</i> Crookall (1976: 948, pl. CLXIX, figs. 12–15)
5	Abo Pass, unit 8, fig. 10h
6	No equivalents
7	Socorro fig. 68.1? (size variation)
8	No equivalents
9	Kola figs. 8a, 13a,b
10	Socorro fig. 68.10?; Sanzenbacher pl. X.14
11	Socorro fig. 68.6; Kola fig. 8b
12	Socorro fig. 69.6 (smaller); Abo Pass, unit 8, fig. 10i (larger)
13	Socorro fig. 69.6; Doña Anas fig. 26D
14	No equivalents
15	No equivalents
16	Socorro fig. 69.7?; Doña Anas fig. 26E
17	Socorro fig. 68.11, 68.12?
18	Socorro fig. 68.4
19	No equivalents
20	Socorro fig. 69.13?
21	No equivalents
22	No equivalents
23	No equivalents
24	No equivalents
25	Socorro fig. 69.3
26	Abo Pass, unit 10, fig. 14e,f?
27	No equivalents
28	No equivalents

authors of the most recent revision of the Kinney Quarry flora (DiMichele et al., 2013) did not agree on an age determination and ultimately opted for Virgilian (Gzhelian), which is at odds with associated marine fossils (Lucas et al., 2011) that indicate a Missourian (Kasimovian) age. Much the same uncertainties

would surround the Alamogordo flora were there no marine fossils to further constrain its age.

The Missourian/Kasimovian appears to have been a period of increased seasonality of rainfall across most of Euramerican Pangea at all phases of glacial–interglacial cycles. Floras rich in conifers, callipterids, cordaitaleans, and other xeromorphic elements appear in well-preserved condition, indicative of only local transport at most. Thus, they almost certainly were growing in the lowland basins under seasonal climates and were not likely carried there by long-distance transport from upland areas. Similar Missourian-age floras, or elements thereof, are known from the central Appalachian Basin (Bassler, 1916; Darrah, 1936; McComas, 1988; Lyons and Darrah, 1989; Martino, 2017), Illinois Basin (Šimůnek, 2018), the U.S. Midcontinent (Elias, 1936; Cridland and Morris, 1963; Winston, 1983; Rothwell and Mapes, 1988), and, as noted above, other areas of the western United States, representing the more westerly portions of Pangea. Red beds also were prominent at this time in the southern and central European portions of Euramerica, the early Stephanian, where they likewise were associated with xeromorphic floras (Wagner and Álvarez-Vázquez, 2010; Opluštil et al., 2017).

In the coal basins of the more central portions of Pangea, ranging from the Midcontinent of North America into central Europe, significant volumes of peat accumulated during the more humid portions of glacial–interglacial cycles, and mixed floras are stratigraphically interdigitated with more typical, strictly wetland assemblages. The wetland assemblages, however, are far more numerous and better preserved than most of the mixed floras. This pattern is a reflection of the greater likelihood of preservation of organic matter deposited under the elevated water tables of humid climatic conditions compared to that deposited during seasonally dry time periods (e.g., Gastaldo and Demko, 2011; Looy et al., 2014). In western Pangea, even the wettest parts of glacial–interglacial cycles appear to have remained seasonally dry compared to conditions in more central regions of Pangea.

Interpreting “Mixed” Floras

Mixed floras, such as those described here, present an important point of departure for understanding the taphonomy of the plants that comprise them. The abundance and diversity of plants typical of variable soil moisture, likely reflecting periodic seasonal moisture stress, and their often excellent preservation as large specimens (e.g., Figures 73A, 82A, 83A) suggest little transport from the site of growth of the source plants. Furthermore, such mesomorphic to xeromorphic plants are as well preserved as those presumed to have grown in microhabitats with more consistently high water tables—the wetland elements, most of which have less robust construction. Add to these observations the additional finding that dominance, on the basis of quantitative sampling, is relatively evenly distributed among the elements drawn from these various soil moisture regimes. From data such as these, we conclude that mixed macrofloras,

FIGURE 101. Specimens of uncertain affinity. (A) Possible conifer cone scale. USNM specimen 699115, USNM locality 43868A. (B) Possibly seeds comparable to *Samaropsis crampii*. USNM specimen 713158, USNM locality 43836D. (C) Unidentified, suspect reproductive organ. USNM specimen 713147, USNM locality 43836B. (D) Unidentified, suspect reproductive organ. USNM specimen 699123, USNM locality 43868A. (E) Axis bearing seedlike structure similar to seed morphotype 14 (see Figure 96K). USNM specimen 706327, USNM locality 43891. (F) Possible lycopsid stem of undetermined affinity; arrow = particularly diagnostic leaf scar (see text). USNM specimen 706333, USNM locality 43891. Scale bar in image F applies to all others. Scale bar = 1 cm.

as distinguished from microfloras, most likely were drawn from spatially heterogeneous, local landscapes, characterized by multiple microhabitats in close proximity.

An important corollary of this interpretation is that the xeromorphic taxa, those interpreted as having lived in periodically moisture-stressed settings, were not transported from some distant (or even nearby), unseen “upland” or “extrabasinal lowland” into a wet bottomland. They were most likely living “on the spot,” proximate to the more mesomorphic to hygromorphic plants. In making this claim, we are challenging a long-standing

and, we would argue, largely erroneous generality that has taken deep root in the literature of late Paleozoic paleobotany: that “elevation” causes enough “drainage” to induce “drought” and thereby to create conditions conducive to a biome entirely distinct from that living in habitats with high soil moisture. Under this model, elevationally induced “drainage” is the key to placing seasonal-drought-tolerant, xeromorphic plants on ancient landscapes. We argue that the elevation = drainage = drought-tolerant biome argument, as a blanket model, is not empirically supported by observations of the modern world or by data from

the fossil record, where such data are available. Rather, it is the interaction of climate and landscape topography that dictates the underlying physical factors controlling site dominance by plants requiring high soil moisture versus those tolerant of, or even requiring, seasonal drought, deep water tables, or freely draining, intermittently dry soil.

Using the modern tropics as a guide, climate and topography will interact to dictate the microhabitat heterogeneity of a local to regional landscape, ignoring, for our purposes here, additional edaphic factors dependent on parent-material chemistry, such as heavy metals (e.g., serpentine-rich soils) or pH (e.g., limestone soil parent material; Scarano, 2002). This is, of course, a scale-dependent phenomenon. Under humid to perhumid climate, using the rainfall classification of Cecil (2003, table 1), which in the tropics means 10 months or more during which rainfall exceeds evapotranspiration, the soil moisture levels will be elevated to the point where variations in landscape topography are greatly muted. At the broader scale, this decrease in variation will lead to the development of a spatially widespread species pool (e.g., Dexter et al., 2017). On a microhabitat scale, there indeed will be compositional variation on these humid-climate landscapes, often correlated with topographic and drainage conditions (e.g., Svenning, 1999).

However, even in the modern tropics, where it can be observed with less ambiguity than in the fossil record, such correlation does not necessarily mean causation, as indicated by inconsistent patterns (e.g., Vormisto et al., 2004), and is not indicative of drastic species pool changes. Examined in detail, such microhabitat differentiation has been suggested to be more reflective of the demography and the history of any given local area than of “drainage” effects (e.g., Kenfack et al., 2014). Most importantly, even considerable topographic variation under such a moist climatic regime does not result in the close proximity of vastly different species pools. Rainforest (e.g., Oliveira-Filho and Fontes, 2000; Bayliss et al., 2014) occurs at elevations from hundreds of meters to over 1,000 m. In the Amazon Basin and in high-elevation Brazilian Atlantic forests, for example, there certainly are variations with space and elevation, but the overall species pool varies enormously from that characteristic of tropical dry forest (Oliveira-Filho and Fontes, 2000; Hughes et al., 2012). Even peat accumulation has been shown to occur at tens of meters of elevation above sea level (e.g., Wüst and Bustin, 2004; Dargie et al., 2017) under humid climatic conditions, in and of itself refuting that elevation = drainage = drought. It is also important to keep in mind the effects of temperature associated with elevation. Lower temperatures, both annual mean and seasonal, and temperature seasonality at higher elevations can be major factors controlling species distributions in the tropics, and elevation also may increase cloudiness and rainfall (thus, under some circumstances, making it “wetter” at higher elevations, not “drier”).

For our argument here, it is necessary to note that modern tropical rainforests are hyperdiverse (e.g., Cannon et al., 1998; Linder, 2001; Hubbell et al., 2008) compared to those of the

Pennsylvanian (e.g., Martín-Closas and Galtier, 2005; Cleal et al., 2012; Moore et al., 2014). Consequently, the probability of habitat specialization might appear to be greater in modern rainforests (more species to fit into and compete for similar resource spaces) than in tropical rainforests of the Pennsylvanian, which, at any one time, consisted of fewer than 150 species of all growth habits. Alternatively, if species-level habitat specialization in modern rainforests is less than (as suggested by neutral models; e.g., Rosindell et al., 2012) or even the same as that of the Pennsylvanian flora (phylogenetic niche conservatism; e.g., DiMichele and Phillips, 1996a; Prinzing et al., 2001), clade-level habitat specialization might be expected to be somewhat more clearly visible in the older, less diverse systems, where it is not masked by high species diversity and finer levels of resource partitioning.

Phylogenetic niche conservatism has been suggested to exist in modern tropical dry forest (e.g., Pennington et al., 2009), which is composed of lineages different from those populating wetlands. A similar pattern of evolutionary lineage-level niche conservatism has been found in tropical rainforest (e.g., Webb, 2000). At the larger scale, modern tropical biomes have distinct evolutionary and ecological histories and are composed of different phylogenetic groups (Hughes et al., 2012). This history takes on added importance when combined with observations of modern tropical ecosystems that suggest the size and diversity of the species pool (biome) have a major effect on such things as local diversity and spatial heterogeneity (e.g., Kraft et al., 2011). Thus, the great similarity across tropical Euramerica of the species that make up Pennsylvanian biomes, be they wetland or seasonally dry, is potentially a reflection of fewer overall species. The small size of Pennsylvanian species pools, whether encompassing plants in need of high water tables and intolerant of drought or consisting of plants adapted to moisture stress, is a major factor in determining the distinctiveness of biomes and the degree to which there is compositional overlap between them. The combination of phylogenetic niche conservatism and variability of landscape heterogeneity as a function of climate has further importance for the interpretation of mixed floras, emphasizing again that by “mixed” we mean fossil assemblages that include taxa from what are generally considered to be distinct biomes.

We assert that under humid to perhumid climate there will be little to no environmentally suitable place for the establishment of drought-adapted plant populations. This will be true especially in the very flat, extensive, coastal plain landscapes that characterized much of the Pennsylvanian equatorial region in Euramerica. As shown by modern vegetation, rainforests can exist on slopes and at high elevation also, as long as rainfall is sufficient, there are only short intervals of drought, and temperature effects do not interpose limitations. This observation alone is *prima facie* evidence that elevation does not, in and of itself, bring on drought.

In contrast, there are often abundant microhabitats with high soil moisture within what are otherwise seasonally dry

landscapes. These may exist along water bodies or water courses. Even swamps may occur under seasonally dry climatic conditions: for example, bottomland swamp forests in eastern North America, which are compositionally distinct from the periodically dry soils of the surrounding terra firma habitats (McKnight et al., 1981).

Here, we direct the reader to a few instructive examples of empirical examinations of the fossil record relevant to the points made here. These particular plant fossil assemblages are from penecontemporaneous intramontane basins dominated by plants considered to be hygromorphic to mesomorphic and thus growing in soils of high moisture content with little periodic moisture stress. If slope and elevation alone cause drought, in such settings, the walls of the adjacent valleys should be expected to have harbored plants adapted to periodic moisture stress, such as coniferales, cordaitales, taeniopterids, and callipterids. The close proximity of a unique drought-tolerant biome of xeromorphic plants and one dominated by mesomorphic to hygromorphic plants typical of high-moisture soils *should have* led to fossil assemblages of mixed character. Elements of the drought-adapted biome would supposedly have been transported downslope into depositional environments of the basin floor (see, e.g., Broutin et al., 1990, for consideration of transport mechanisms). This mixing of biomes also might be expected to inflate local diversity. What is important about these examples is that elements of the seasonally dry biome are *not* part of the assemblages. In these instances, negative data are as important as positive—if elevation causes drought, *then* it does, and here is where drought-adapted taxa regularly should be found in the floras. Examples come especially from the Variscan mountainous areas of central Pangea in Germany (Uhl and Cleal, 2010), the Czech Republic (Libertín et al., 2009; Bashforth et al., 2011), the Iberian Peninsula (Wagner and Álvarez-Vázquez, 2010), and Sardinia (Cleal et al., 2017). These studies reveal a notable paucity of conifers or other plants considered typical of moisture-limited substrates. This scarcity is summed up quite lucidly by Wagner and Álvarez-Vázquez (2010:305), who note “it seems surprising that the presence of upstanding relief with alluvial fan deposits in the Peñarroya Basin (Westphalian) has not led to any conifer finds. There are also very few conifer records in the Stephanian B of NW Spain, despite the evidence for palaeovalleys associated with a rugged landscape in the near vicinity of the basin.” Surprising indeed. Perhaps the model of slope and elevation = drought = seasonally dry biome is incorrect, at least in its simplest form, if lacking consideration of prevailing climate.

Adding climate to the equation greatly clarifies this matter. Had climate been seasonally dry during the time these selected paleofloras lived in the intramontane, narrow, steep-sided valleys, the slopes very well may have been populated by the seasonally dry biome. Much of the floodplain on the valley floor may have been similarly populated. But in places, along the streams and ponds of the floodplain, there may have been elements of the other biome, composed of plants with obligate requirements for high soil moisture, tolerant of only short periods of drought.

That this is not the case suggests that the paleofloras referenced above were formed when the valleys were under humid to perhumid climate, when the valley walls would have been populated by, at most, the more mesomorphic elements of the moisture-dependent flora. Such an example may be provided by Opluštil and Cleal (2007:431). They described a flora from the Lower Radnice Member of the Czech Republic that is high in diversity (140 species), including many unusual taxa. The flora is preserved in deposits of a narrow paleoriver valley that was flanked by walls up to 200 m in height. The unusual elements, mainly sphenopterid ferns and pteridosperms, are interpreted to have grown on the slopes surrounding the depositional environment. These plants were not members of the typical, xeromorphic, seasonally dry biome but rather were uncommon members of the wetland biome that might be considered “mesomorphic,” tolerant of short periods of moisture stress. Expectations such as these would be more in conformance with patterns observed in the modern tropics than are blanket assertions that drought-tolerant taxa grew in uplands or that elevation causes drought.

A different way to look at the problem is represented by the discovery of conifer remains in the late Middle Pennsylvanian of the Illinois Basin (Falcon-Lang et al., 2009). At that time, there were no bone fide uplands within 1,000 km. The remains were found in strata between coal beds, not in the position of typical coal roof shales, and are not composed of wetland assemblages. Rather, they are dominated by cordaitalean remains. A similar paleoflora, although not containing conifers, but with abundant *Taeniopteris*, *Lesleya*, and cordaitalean foliage, was reported from the eastern margin of the Illinois Basin in strata of Middle Pennsylvanian age between, but not associated with, coal beds (Bashforth et al., 2016). Again, this paleoflora formed in an area far distant from any contemporaneous high-elevation areas. The conclusion we draw from findings such as these, which similarly are reflected in the Garnett, Kansas, flora discussed below, and other examples (see Tabor et al., 2013, for intercalated, but distinct, wetland and seasonally dry floras in the Late Pennsylvanian of Texas; floras of the Hamilton Quarry in Kansas [Cunningham et al., 1993] and the Kinney Quarry in New Mexico [DiMichele et al., 2013] present somewhat different environmental situations but also are mixed floras) is that in the more central portions of Pangea, distant from contemporaneous uplands, floras characteristic of high soil moisture and seasonal drought occurred in distinct strata recording different environmental conditions at different times. Each was periodically dominant in tropical lowlands but at different times, most likely reflecting floristic tracking of the climate changes associated with glacial–interglacial cyclicality.

Special notice, in this regard, needs to be made of the highly influential paper of Cridland and Morris (1963). These authors argued that the conifer-dominated flora from Garnett, Kansas, in the Rock Lake Shale (Winston, 1983) was derived from an upland of about 6 m elevation in the midst of an otherwise wetland, peat-forming landscape; they illustrated this conclusion in a much reproduced figure showing the upland in close lateral

proximity to the swamplands. They used landscape variability in present-day Kansas as their point of comparison. Nevertheless, a first-order examination of this scenario renders it unlikely. Modern-day Kansas is not a good model for Pennsylvanian peat-forming landscapes. Its semiarid climate cannot support tropical rainforest, and that climate, in addition, greatly magnifies and accentuates landscape elevational variability in soil moisture. The Garnett site itself, particularly given the taphonomy and associated vertebrate fauna (e.g., Reisz et al., 1982), does not support the Cridland and Morris (1963) environmental reconstruction.

The Rock Lake Shale, in which the Garnett flora and fauna are preserved, fills channels formed on a seasonally dry landscape (based on paleosols laterally equivalent to the channel fill; Joeckel, 1989) that existed between periods of widespread wetland development and peat formation (Feldman et al., 2005). This location leads uncontroversially to the conclusion that the flora of the Garnett locality was not living in close proximity to vast peat-swamp habitats or other wetland and high-soil-moisture settings that occupied the Midcontinent during times of humid climate. Although the drought-tolerant species pool that comprises the Garnett paleoflora dominated the landscape, under a subhumid to semiarid climate, there may have been patches of wetlands within that landscape, just as has been documented for other times and places in the past where late Paleozoic paleofloras of seasonally dry habitats are dominant (e.g., DiMichele et al., 2006). However, it must be kept in mind that the Garnett flora and fauna come from a time in a glacial-interglacial cycle of intense, likely orbitally driven sea-level and climate change (e.g., West et al., 1997), which overprinted the tropical regions and was accompanied by changes in flora and fauna. Patterns like this characterize nearly all of the Pennsylvanian in Euramerica (e.g., Falcon-Lang, 2004; Heckel et al., 2007; Eros et al., 2012; Bashforth et al., 2014, 2016; Van den Belt et al., 2015).

There is strong evidence that climate controlled the composition of the plant communities, interacting with slope and aspect. This is evident in areas where there was undoubted uplift and elevation proximate to coastal, depositional basins. Such areas include portions of New Mexico such as that considered here, arid regions such as the Paradox Basin to the northwest of the study area (Dubiel et al., 1996; Soreghan et al., 2002), and even in the Appalachian Basin. In such settings, there is evidence that the seasonally drought-tolerant flora grew both in the surrounding uplands (e.g., Condon, 1997; Stanesco et al., 2002; Falcon-Lang et al., 2014) and in the basinal lowlands (e.g., Mamay and Breed, 1970; Tidwell, 1988; DiMichele et al., 2014; Martino, 2017), mixed or unmixed with wetland elements in the latter landscape. In the basinal deposits there is no distinct preservational difference between the drought-tolerant and wetland elements of the paleoflora; all are relatively well preserved and with no indication of long-distance transport. In certain extreme cases, characterized by unusual preservational conditions, the drought-tolerant flora has been found rooted in place

in coastal depositional environments (Falcon-Lang et al., 2011, 2016), where its lowland, basinal presence cannot be disputed. All these observations point to the conclusion that ancient landscapes were complex, just as they are today, and that ancient, late Paleozoic plants responded in much the same way as modern plants to climatic and edaphic conditions in the terrestrial environment. The upland = drought model, as an inflexible truism, is not supported either by observations of the modern world or by detailed dissection of the late Paleozoic fossil and geological records.

In conclusion, the fossil floras described herein, from the Gobbler, Beeman, and Holder Formations, almost certainly grew under seasonally dry conditions. The times during which these floras occupied the regional landscape may have been the wettest periods of glacial-interglacial cycles. That there were even drier periods is indicated by the characteristics of limestones and other marine to perimarine deposits, including evidence of dust and faunal composition, discussed elsewhere in this paper. Although the lithostratigraphic expression, and even detectability, of these cycles was strongly affected by contemporaneous tectonism in the Alamogordo region, climate changes associated with such cyclicity would have happened independently of such tectonism and affected the region, regardless of the presence or lack of appropriate accommodation space to preserve evidence of clearly cyclic sedimentary patterns. Thus, although not regular in their occurrence, the floras described here capture the times when wetter, subhumid climate periods happened to overlap with conditions favorable for the preservation of the deposits in which they were entombed.

ARTHROPOD HERBIVORY

The Pennsylvanian was a key time interval in the evolution and expansion of modern forms of arthropod herbivory (Scott and Taylor, 1983; Labandeira and Phillips, 1996a, 1996b; Labandeira et al., 1997). However, few comprehensive studies of arthropod herbivory associated with Pennsylvanian compression floras have been performed (Castro, 1997; Xu et al., 2018), which limits our understanding of the timing of first appearances, extent of interactions, and host preferences of arthropod herbivores during this time. The Alamogordo paleobotanical collections (USNM localities) represent mixed floras, ideal for assessing the preferences of Pennsylvanian arthropod herbivores for plants occupying various microhabitats. The paleobotanical collections are organized into nine stratigraphic levels from oldest to youngest (Table 4, Figure 47). We inspected all plant fossil specimens from these collections for evidence of arthropod and fungal damage. We assigned damage types (DTs; Table 7), distinct categories of arthropod and fungal damage defined by shared morphological characteristics, using the *Guide to Insect (and Other) Damage Types on Compressed Plant Fossils* (Labandeira et al., 2007) and described all observed damage, categorized by host plant. Damage was found only on floras from the Beeman Formation.

TABLE 7. Summary of arthropod damage types on leaves, pinnules, and axes from the Beeman Formation floras.

USNM specimen	USNM locality	Stratigraphic level	Plant species	Damage type	Functional feeding group
698978	43868A	9	<i>cf. Odontopteris subcrenulata</i>	DT12	Margin feeding
698981	43868A	9	<i>Macroneuropteris scheuchzeri</i>	DT12	Margin feeding
698913	43891	9	<i>Sphenophyllum</i> sp.	DT5	Hole feeding
706341	43891	9	<i>Macroneuropteris scheuchzeri</i>	DT75	Surface feeding/fungal damage
713196	43838	9	<i>Macroneuropteris scheuchzeri</i>	DT75	Surface feeding/fungal damage
698912	43891	9	<i>Neurodontopteris auriculata</i>	DT97	Surface feeding/fungal damage
698919	43891	9	Unidentifiable axis	DT101	Oviposition
698927	43836C	7	<i>Taeniopteris</i> sp.	DT12	Margin feeding
713121	43836A	7	<i>Cordaites</i>	DT54	Oviposition

EXTERNAL FOLIAGE FEEDING

Margin feeding is the excision of plant tissue along the margin of a leaf or a pinnule, which is typically produced by mandibulate insects. The damage results in the formation of a dark rim of thickened reaction tissue along the wounded edge of the leaf. We found examples of margin feeding on *cf. Mixoneura* (*Odontopteris*) *subcrenulata* and *Macroneuropteris scheuchzeri* at Beeman Canyon (USNM locality numbers 43891 and 43868A) and on *Taeniopteris* sp. at La Luz Canyon (USNM locality number 43836C). U-shaped margin feeding on *cf. Mixoneura* (*Odontopteris*) *subcrenulata* is 2.9 mm wide by 2.9 mm deep (Figure 102A) with a 0.2 mm wide reaction rim (DT12). On *M. scheuchzeri*, a wide and shallow, partially preserved margin-feeding excision (10.0 mm wide by 1.9 mm deep) is surrounded by a 0.1–0.2 mm wide reaction rim (DT12; Figure 102B). Margin-feeding damage on *Taeniopteris* sp. is represented by a semicircular excision measuring 3.5 mm wide by 1.5 mm deep with a 0.2–0.3 mm wide reaction rim (Figure 102G). Although the most common form of damage, margin feeding is still very rare on the examined floras and is found on only three hand-sample quadrats (0.41% and 0.22% of informative quadrats at USNM localities 43836 and 43891, respectively). At USNM locality 43891, margin feeding is associated with 14.29% of quadrats with *cf. M. subcrenulata* and 1.25% of quadrats with *Macroneuropteris scheuchzeri*. At USNM locality 43836, margin feeding is associated with 16.67% of quadrats with *Taeniopteris* sp.

The second form of external foliage feeding found in this study is hole feeding, which is typically caused by mandibulate insects that chew through the entire thickness of laminar leaf tissue. A single oval hole (DT5; Figure 102C) is associated with a possible *Sphenophyllum* sp. leaf or unidentifiable axis at Beeman Canyon (USNM locality 43891). The hole is 12.5 mm long by 4.2 mm wide, and its long axis is parallel to the length of the leaf. A reaction rim measuring 0.4–0.9 mm wide surrounds the

hole. Hole feeding is very rare, with only one occurrence, found on 0.11% of informative quadrats from USNM locality 43891 (20% of quadrats with *Sphenophyllum* sp.).

Surface feeding, the third form of external foliage feeding damage observed in this study, is typically made by mandibulate insects that preferentially feed on the upper layers of leaves and stems, often damaging epidermal or hypodermal cells. Two *Neurodontopteris auriculata* pinnules at Beeman Canyon (USNM locality 43891) are associated with possible surface feeding. In Figure 102D, possible surface-feeding damage is semicircular and lighter in color than the surrounding unaffected tissue (DT97), suggesting that the tissue was thinner in the damaged area at the time of deposition. The surface-feeding zone measures 4.8 mm long (distance along the margin) by 3.2 mm deep, surrounded by a 0.4 mm wide reaction rim. Surface feeding on a second pinnule (Figure 102E) is more elongate and encompasses the area from the margin of the pinnule to the midvein (DT97). The surface feeding zone is 13.0 mm long by 3.7 mm wide but is not surrounded by an obvious reaction rim.

On two *Macroneuropteris scheuchzeri* pinnules, one from USNM locality 43838 (Figure 74C) and the other from USNM locality 43891 (Figure 74B), V-shaped necrotic areas extend from near the pinnule midveins, widening toward the leaf margins and typically following secondary veins. The zones of damaged tissue are 0.9–1.8 mm wide and 4.4–12.1 mm long and are surrounded by a rim of callus tissue (0.2–0.4 mm wide).

Surface-feeding damage is found on 0.32% of informative quadrats at USNM localities 43868, 43869, and 43891, all at the same stratigraphic level (3.03% of quadrats with *Neurodontopteris auriculata* and 2.45% with *M. scheuchzeri*). Although DT97 and DT75 have been interpreted as surface feeding in some past studies (Labandeira et al., 2007; Xu et al., 2018), pathogenic fungal damage has also been suggested as another explanation for these surface lesions (Labandeira et al., 2007; Stull et al., 2013; Labandeira and Prevec, 2014).

FIGURE 102. Arthropod damage on fossil plants from the Beeman Formation. Beeman Canyon (A–F) and La Luz Canyon (G, H). (A) Excision into the leaf margin of *cf. Odontopteris subcrenulata* (DT12; USNM specimen 698978, USNM locality 43868A). (B) Margin feeding on *Macroneuropteris scheuchzeri* (DT12; USNM specimen 698981, USNM locality 43868A). (C) Oval hole on stem of *Sphenophyllum* sp. leaf (DT5; USNM specimen 698913, USNM locality 43891). (D) U-shaped surface feeding or fungal damage extending from the leaf margin of *Neurodontopteris auriculata* (DT97; USNM specimen 698912, USNM locality 43891). (E) Elongate patch of surface feeding or fungal damage between the leaf margin and midvein of *Neurodontopteris auriculata* (DT97; USNM specimen 698912, USNM locality 43891). (F) Oviposition mark on plant axis (DT101; USNM specimen 698919, USNM locality 43891). (G) Semicircular excision into the leaf margin of *Taeniopteris* sp. (DT12; USNM specimen 698927, USNM locality 43836C). (H) Oviposition mark on cordaite leaf; box expands to I (DT54; USNM specimen 713121, USNM locality 43836A). (I) Close-up of oviposition mark from H. Scale bars = 5 mm in A–E, 1 mm in F, 2 mm in G, I, 1 cm in H.

OVIPOSITION

A female insect inflicts oviposition damage on plants when she uses her elongate ovipositor to deposit an egg on the surface of a plant or within plant tissue. We found oviposition damage on an unidentifiable axis from USNM locality 43891 and on a *Cordaitea* leaf from USNM locality 43836A. An isolated oval oviposition mark on an unidentified stem is oriented parallel to the length of the axis and measures 1.3 mm long by 0.6 mm wide (DT101; Figure 102F). A dark circular depression near the center of the oviposition mark (0.3 mm in diameter) may represent the ovipositor insertion point, suggesting that the surrounding lighter-colored area is callus tissue. A 0.2 mm reaction rim surrounds the oviposition mark. Oviposition damage is found on 0.11% of informative quadrats at USNM localities 43868, 43869, and 43891 (0.67% of quadrats with axes).

At stratigraphic level 7, oviposition is found on 0.40% of informative quadrats (1.85% of quadrats with *Cordaitea* leaves). On the cordaitalean leaf associated with oviposition, 13 elongate ellipsoidal bulges are oriented parallel to the length of the leaf (DT54; Figure 102H,I). The oviposition marks are 2.7–3.5 mm long by 0.5–0.9 mm wide. Most of the oviposition marks are covered in a layer of plant tissue, suggesting that the eggs were deposited endophytically, leading to bulging. Eleven out of 13 oviposition marks are positioned on the right half of the specimen in the view in Figure 102I and are possibly arranged in a zigzag or arcuate pattern. Extant damselflies in the family Coenagrionidae produce arcuate or zigzag oviposition mark arrangements (Sarzett et al., 2009). Similar behavior extends back to the Pennsylvanian on possible leaves of *Cordaitea* from the Gzhelian of Germany and may have been produced by Archizygoptera, an extinct odonatopteran suborder (Laaß and Hoff, 2014). The oviposition marks on the cordaitalean leaf from the Beeman Formation are not as densely positioned as typical coenagrionid-type oviposition damage (Sarzett et al., 2009), so more specimens are necessary to determine if the zigzag or arcuate arrangement is a consistent pattern.

DISCUSSION OF ARTHROPOD HERBIVORY

Arthropod herbivory is associated with Beeman Formation floras from Beeman Canyon (USNM localities 43868A, 43891), La Luz Canyon (USNM localities 43836A, 43836B), and Horse Ridge (USNM locality 43838). Floras from the two stratigraphic levels represented by these collections are the best sampled in the study. Damage was found on only 10 plant fossils total (nine quadrats): nine leaves and pinnules and one axis. The functional feeding groups present, including external foliage feeding (margin feeding, hole feeding, surface feeding) plus oviposition, have been previously recognized in the Pennsylvanian, and the Paleozoic histories of these groups of arthropod damage are discussed below. Four important functional feeding groups are not present in these floras, including leaf mining, which first appeared in association with mid-Triassic floras

(Tillyard, 1922; Rozefelds and Sobbe, 1987; Scott et al., 2004; Labandeira, 2006; Meller et al., 2011); galling, which was present in the Late Pennsylvanian but diversified in the Permian (Labandeira and Phillips, 1996b; Schachat and Labandeira, 2015; Schachat et al., 2015; DiMichele et al., 2018a,b); piercing and sucking, which was also present in the Pennsylvanian (Kevan et al., 1975; Labandeira and Phillips, 1996a; Schachat et al., 2014) but often difficult to recognize in compression floras; and skeletonization, which was rare in the Paleozoic (Feng et al., 2014).

From all of the plant groups composing the Alamogordo floras, pteridosperms (*Macroneuropteris scheuchzeri*, cf. *Mixoneura* [*Odontopteris*] *subcrenulata*, and *Neurodopteris auriculata*) had the highest number of damage occurrences, with six damaged pinnules. The targeting of pteridosperms by arthropods is a consistent feature of the food web structure observed in many studies of Pennsylvanian and Permian compression floras (Scott and Taylor, 1983; Beck and Labandeira, 1998; Labandeira and Allen, 2007; DiMichele et al., 2018a,b; Xu et al., 2018). Pteridosperms were a readily available food source for herbivorous arthropods because of their high abundance and ecological dominance in early Permian ecosystems. In the Permian of Euramerica, herbivory was concentrated on medullosan pteridosperms, peltasperms, gigantopterids, and *Taeniopteris* sp. (Beck and Labandeira, 1998; Labandeira and Allen, 2007; Schachat et al., 2014, 2015; DiMichele et al., 2018a,b; Xu et al., 2018), whereas in Gondwana, herbivory was most dominant on *Glossopteris* species (Prevec et al., 2009; McLoughlin, 2011; de Souza Pinheiro et al., 2012; Gallego et al., 2014; Cariglino, 2018). Some seed plants from the Pennsylvanian and Permian exhibit probable antiherbivore defenses, including chemical defenses (Van Bergen et al., 1995; Zodrow, 2014) and trichomes (Krings and Kerp, 1999; Krings et al., 2002).

Studying Paleozoic compression floras for arthropod herbivory allows us a glimpse of the habitat and food source preferences of early arthropod herbivores. Because of the rarity of damage in this study, clear microhabitat preferences for herbivores are difficult to discern, although plants that presumably occupied different microhabitats were damaged. For example, cf. *Mixoneura* (*Odontopteris*) *subcrenulata* preferred seasonally dry habitats (Wagner and Álvarez-Vázquez, 2010), whereas *Macroneuropteris scheuchzeri* (Stull et al., 2012) and *Sphenopteris* sp. preferred wetland environments. These results suggest that during the Missourian, herbivorous arthropods fed on plants occupying a variety of microhabitats across the Pennsylvanian landscape, which continued into the Permian (Beck and Labandeira, 1998; Labandeira and Allen, 2007; Schachat et al., 2014, 2015; Xu et al., 2018).

Paleozoic History of Arthropod Herbivory

Margin feeding is one of the most common forms of arthropod damage in the Paleozoic and has been documented in numerous studies. The earliest evidence of margin feeding is small, cusped

excisions (DT12, DT13) associated with the earliest known liverwort, *Metzgeriothallus sharonae* from the Middle Devonian of New York State (Labandeira et al., 2014). A gap in the fossil record of margin feeding extends to the Late Mississippian (Serpukhovian) in the Sydney Basin, New South Wales, Australia, where DT12 margin feeding was associated with *Triphyllopteris austrina*, a pteridosperm (Iannuzzi and Labandeira, 2008). In the Pennsylvanian, margin-feeding damage has been documented in association with pteridosperms, such as *Macroneuropteris scheuchzeri* (Scott and Taylor, 1983; Xu et al., 2018), *Pseudomariopteris cordato-ovata* (Xu et al., 2018), *Mixoneura subcrenulata* (DiMichele et al., 2013), and *Neurodontopteris auriculata* (DiMichele et al., 2013). Margin feeding has also been found on the sphenophyte *Annularia carinata* (Xu et al., 2018). In Permian floras, margin feeding (DT12, DT13, DT14, DT15) is associated with a variety of seed plants (Beck and Labandeira, 1998; Glasspool et al., 2003; Looy, 2007; Prevec et al., 2009; Cariglino and Gutiérrez, 2011; Schachat et al., 2014, 2015; Cariglino, 2018; Xu et al., 2018; Maccracken and Labandeira, 2019).

Evidence of hole feeding in the Paleozoic is rare, especially before the Permian. The first documented occurrence of hole feeding is on *Metzgeriothallus sharonae*, a liverwort from the late Middle Devonian of New York State, the same plant associated with the earliest margin-feeding damage (Labandeira et al., 2014). After the Middle Devonian, there is a major gap in the fossil record of hole feeding until the Late Pennsylvanian (Missourian) with the *Sphenophyllum* sp. association discussed in this study. In the late Gzhelian, hole-feeding arthropods targeted *Macroneuropteris scheuchzeri* from the Williamson Drive flora of north central Texas (Xu et al., 2018). A single occurrence of hole feeding on *Pseudomariopteris cordato-ovata* was also documented (Xu et al., 2018). A hole is associated with *Mixoneura wagneri* of Stephanian C age from the La Magdalena Coalfield in León, northwest Spain (Castro, 1997). During the Permian, the morphological and host plant diversity apparently increased (Beck and Labandeira, 1998; Looy, 2007; Prevec et al., 2009; de Souza Pinheiro et al., 2012; Gallego et al., 2014; Cariglino, 2018). However, the near absence of hole feeding from the Pennsylvanian fossil record may in part stem from a lack of investigations into arthropod herbivory on Pennsylvanian compression floras.

The earliest evidence of surface feeding (DT29, DT30) is associated with the Middle Devonian liverwort *Metzgeriothallus sharonae*, the same plant associated with the first hole- and margin-feeding interactions (Labandeira et al., 2014). In the Pennsylvanian, surface feeding is associated with pteridosperms, such as *Macroneuropteris scheuchzeri*, *Alethopteris zeilleri*, *Pseudomariopteris cordato-ovata*, and *Neuropteris ovata* (Scott and Taylor, 1983; Xu et al., 2018). Surface feeding is found on a broader range of seed plants in the Permian, including medullosans, gigantopterids, peltasperms, and conifers (Beck and Labandeira, 1998; Vasilenko, 2007; Prevec et al., 2009; Srivastava and Agnihotri, 2011; de Souza Pinheiro et al., 2012; Gallego et al.,

2014; Labandeira and Prevec, 2014; Schachat et al., 2014, 2015; Xu et al., 2018). As noted above, fungal infection may be another explanation for some of these damage types, particularly DT75 and DT97 (Labandeira et al., 2007; Stull et al., 2013; Labandeira and Prevec, 2014). Detailed histological studies of damaged tissue are necessary to determine the origin of these structures.

The earliest endophytic oviposition marks are from the Middle Pennsylvanian on *Calamites cistii* from the Graissessac Basin in Hérault, France (Béthoux et al., 2004). In the Late Pennsylvanian (Gzhelian), arcuate patterns of endophytic (Laaß and Hoff, 2014) and exophytic (Laaß and Hauschke, 2019) oviposition marks are found on a *Cordaites* leaf from the Siebigerode Formation of the Saale Basin in Germany. In north central Texas, 11 oviposition DTs were recognized on the Gzhelian Williamson Drive flora on sphenophyte stems, medullosan pteridosperm stems and pinnules, and lycopsid leaves (Xu et al., 2018). Evidently, the use of foliar tissue as a substrate for oviposition increased in the Permian, particularly on glossopterid-dominated floras (Prevec et al., 2009; McLoughlin, 2011; de Souza Pinheiro et al., 2012; Gallego et al., 2014; Cariglino, 2018).

MARINE MICROFOSSILS

ALGAE AND FORAMINIFERS

We studied the calcareous microfossils (primarily algae and foraminifers) in thin sections of limestone of the Gobbler, Gray Mesa, Beeman, and Holder Formations. Detailed assemblages are listed below and illustrated in Figures 103–110. The classifications of carbonate microfossils used here are those of Vachard (2016a, 2016b).

Gobbler Formation Assemblage

Algae from the Gobbler Formation limestones are not diverse and are identified as *Archaeolithophyllum missouriense* Johnson and epimastoporacean? dasyclads. The foraminifers are relatively diverse and represent the Fusulinata, Miliolata, and Nodosariata. The Fusulinata are *Eotuberitina reitlingerae* Miklukho-Maklay, *Earlandia* sp., N. gen. 1 cf. *Pseudolituotuba* sp. (Figure 103H), *Endothyra* ex gr. *similis* Rauzer-Chernousova and Reitlinger, *Planoendothyra* sp. (Figure 103K,L), *Bradyina* ex gr. *lepida* Reitlinger (Figure 103O), *Tetrataxis* spp., *Globivalvulina* sp., *Eostaffella* cf. *ampla* Thompson (Figure 103D), *Eostaffella* sp. 1 (Figure 103E), *Seminovella* sp. (Figure 103A), *Millerella extensa* Thompson (Figure 103B), *M. pressa* Thompson (Figure 103C), *Profusulinella* sp. (Figure 103F,G), and *Eoschubertella/Schubertina* sp.

The Miliolata are *Pseudospira* sp., *Calcivertella* spp., N. gen. 2, cf. *Eoglomospiroides* sp. (Figure 103I,J,N,S), *Trepeilopsis* sp., *Orthovertella*? sp. (Figure 103P), N. gen. 3 aff. *Plumerinella* sp. (Figure 103T), *Orthovertellopsis* sp. indet.,

Palaeonubecularia sp., *Cornuspira* sp., and *Hemigordiellina* sp. The Nodosariata are *Syzrania bella* Reitlinger (Figure 103M), *Syzrania* sp. (Figure 103Q,U), and *Syzrania confusa* Reitlinger (Figure 103R).

Gobbler Formation Foraminiferal Biostratigraphy

At the Space History A section (Figure 11), a thin limestone bed (Figure 11, Space History A, unit 3, sample SHA-3), exposed 5 m above the Mississippian Lake Valley Formation, yielded several millerellins: *Seminovella* sp., *Millerella extensa*, and *M. pressa*. This level is probably Morrowan in age according to the criteria of Groves (1983, 1984a). Foraminifer recovery is poor from higher samples from the Space History A section. However, the LO of *Profusulinella* in the Space History B section, sample SH 23 (Figure 12A), indicates the early Atokan, whereas the LO of *Syzrania* in the Space History B section, sample SH 41 (Figure 12A), possibly indicates the upper part of this substage or, already, the late Atokan, according to the data of Wilde (1975, 2006), Krainer et al. (2011, 2015, 2017b, 2017c), Vachard et al. (2012, 2013a, 2013b), and Lucas et al. (2017). These boundaries are relevant to those indicated by the conodonts (see below).

The Gobbler Formation displays coralline red algae, *Archaeolithophyllum missouriense*, and the oldest dasycladalean green algae of the succession, which are generically undetermined epimastoporaceans. It is noteworthy that the genus *Orthovertellopsis*, recently described from the lower Permian of New Mexico by Vachard et al. (2013b), appears regionally as early as the Desmoinesian. Free Miliolata are relatively rare and represented by the classical quartet of *Turrispiroides*, *Cornuspira*, *Hemigordiellina*, and *Hemigordius* (Cushman and Waters, 1928; Reitlinger, 1949, 1950; Groves, 1983, 1984b; Vachard, 2016a; Krainer et al., 2017b, 2017c; Vachard et al., 2018).

Gray Mesa Formation Assemblage (High Rolls)

At the High Rolls section, limestones of the Gray Mesa Formation contain algae and diverse foraminifers. The algae are identified as *Nostocites vesiculosa* Maslov (Figure 105J), *Stacheoides* sp. (Figure 104J), and *Komia eganensis* Wilson, Waines, and Coogan (Figure 104A,F–H).

As was the case with the Gobbler Formation foraminifers, those of the Gray Mesa Formation can be divided into Fusulinata, Miliolata, and Nodosariata. The Fusulinata are *Eotuberitina reitlingerae* Miklukho-Maklay, *Tuberitina bulbacea* Galloway and Harlton, *Earlandia* sp., *Monotaxinoides transitorius* Brazhnikova and Yartseva (Figure 104I), *Endothyra* sp., *Planoendothyra* sp., *Bradyina* spp., *Bradyinelloides* sp. (Figure 105I), *Palaeotextularia* sp., *Climacammina* ex gr. *moelleri* Reitlinger, *Globivalvulina* sp. (Figure 104C), *Tetrataxis acuta* Durkina (Figure 104E), *Tetrataxis* ex gr. *pressula* Malakhova,

Polytaxis sp., *Pseudoacutella* sp. (Figure 104D), *Wedekindellina* sp. (Figure 104K), *Nipperella* sp. (Figure 105D), and *Beedeina schellwieni* (Staff) (Figure 105E).

The Miliolata are much less diverse: *Pseudospira?* sp., *Calciwertella* spp., *Calcitornella* sp., *Trepeilopsis* sp. (Figures 104B, 105H), *Orthovertellopsis* sp. indet. (Figures 104L, 105A), *Cornuspira* sp. (Figure 105C), and *Turrispiroides* sp. (Figure 105F,G). Only one species of Nodosariata was identified: *Syzrania bella* (Figure 105B).

Foraminiferal Biostratigraphy of Gray Mesa Formation (High Rolls)

A few foraminifers and algae were recovered from the Gray Mesa Formation at High Rolls (samples HRO; Figure 20). The lower samples (Figure 20, HRO 1, HRO 2 from the uppermost Gobbler Formation) yielded *Komia eganensis*, which are possibly late or latest Atokan and/or earliest Desmoinesian in age, as well as the millerellin *Pseudoacutella*. As a consequence, no markers of the late Atokan, especially *Fusulinella* itself, were found. The early Desmoinesian is first characterized in unit 27 (sample HRO 10; Figure 20) of the section with *Wedekindellina* and then in unit 43 (sample HRO 13; Figure 20) with *Nipperella* sp. and in unit 53 (sample HRO 17; Figure 20) with *Beedeina schellwieni*. Algae are relatively rare in our material, except for *Komia*, which are everywhere well documented in the New Mexico Pennsylvanian sections (Krainer et al., 2003, 2009a, 2009b, 2011, 2015, 2017b, 2017c; Vachard et al., 2012; Lucas et al., 2015, 2016a, 2016b). In the lower Desmoinesian strata of the Gray Mesa Formation of High Rolls, there are well-preserved cyanobacteria assigned to *Nostocites vesiculosus*.

Unlike the genera *Wedekindellina* and *Beedeina*, *Nipperella* is rarely cited in the fusulinid assemblages of the base of the Desmoinesian. *Nipperella* was created by Solovieva (1984) to designate an advanced form of fusulinellid. Our species has similarities to *Nipperella matura* (Thompson), initially described as *Wedekindellina*, but it differs in its smaller parameters, shape of chomata, and deviated early whorls. *Nipperella matura* is similarly mentioned at the base of the Desmoinesian in Colorado, Arizona, and Utah (Thompson, 1945; Ross and Sabins, 1965; Slade, 1961) and in the Myachkovian-Kasimovian of Russia (Solovieva, 1984).

Gray Mesa Formation Assemblage (Space History B)

A much less diverse assemblage of algae and foraminifers was recovered from limestones of the Gray Mesa Formation at the Space History B section (Figure 13). One algal species can be identified, *Anthracoporellopsis novamexicana* Vachard, Krainer, and Lucas (Figure 106D).

The Fusulinata are *Eotuberitina reitlingerae*, *Tuberitina bulbacea*, *Tuberitina collosa* Reitlinger, *Earlandia* sp., *Bradyina* sp., *Globivalvulina* sp., and *Staffella* sp. The Miliolata are

more diverse: *Pseudospira*? sp., *Calcivertella* sp., *Calcitornella* sp., *Orthovertellopsis* sp. indet. (Figure 106H), *Palaeonubecularia* cf. *rustica* Reitlinger (Figure 106B), *Latitubiphytes* spp. (Figure 106A,C,K), *Turrispiroides* sp. (Figure 106E), *Hemigordiellina* sp., *Hemigordius* cf. *simplex* Reitlinger (Figure 106F,G), *H. cf. discoideus* (Brazhnikova and Potievskaya) (Figure 106I), and *H. ex gr. harltoni* Cushman and Waters (Figure 106J).

Foraminiferal Biostratigraphy of Gray Mesa Formation (Space History)

This part of the Gray Mesa Formation, in the Space History B section (Figure 13), is possibly late Desmoinesian because of *Anthracoporellopsis novamexicana* (LO in unit 153). Its upper part can be earliest Missourian because of the appearance of the oldest regional *Latitubiphytes* (LO in unit 160; see Vachard et al., 2012). With such a biostratigraphy, the diversification of the genus *Hemigordius*, which occurs also in unit 160, may be added as another regional character for characterizing the base of the Missourian (see Groves, 1984b). Similar results are provided by the conodonts (see below), which indicate more precisely that the LO of the lowest Missourian fauna is in unit 155 and that the lower Missourian conodont fauna becomes typical from unit 160.

Beeman Formation Assemblage

No algae were recovered from the Beeman Formation. The foraminifers are not very diverse. The Fusulinata are *Pseudojanischewskina* sp. 1 (Figure 107I,J), *P. sp. 2* (Figure 107K), *Bradyina* sp., *Climacammina* sp., *Globivalvulina* sp. (Figure 107E), and *Tumulotriticitites* sp. (Figure 107A).

The Miliolata are *Pseudospira* sp. (Figure 107L), *Calcivertella* sp., *Calcitornella* sp., *Glomospiroides* sp. (Figure 107G), *Orthovertellopsis*? sp. (Figure 107H), *Palaeonubecularia* sp. 2

(Figure 107B,D), and *Hemigordiellina* sp. The single species of *Nodosariata* we recognized is *Syzrania bella* (Figure 107F).

Beeman Formation Foraminiferal Biostratigraphy

A late? Desmoinesian-Missourian age of the Beeman Formation is suggested by *Tumulotriticitites* (a so-called genus that is probably a junior synonym of evolved *Protriticitites* and/or primitive *Montiparus*) and by the transitional stages between *Palaeonubecularia* and *Latitubiphytes* (see Wilde, 2006; Vachard et al., 2012, 2013a). The other smaller foraminifers are not age indicative. Conodonts (see below) provide more precise data on the base of the Missourian in the Beeman Formation in the Space History C section (Figure 25, top of unit 44) and Space History B section (Figure 13, unit 212).

Holder Formation Assemblage

We recovered a diverse assemblage of algae, cyanobacteria, and foraminifers from limestones of the Holder Formation. The algae and cyanobacteria are identified as *Bacinella* sp. (Figure 109K), *Ivanovia tenuissima* Khvorova (Figure 108A), *Eugonophyllum johnsonii* Wray (Figures 108I?, 110M), *Paraepimastopora* sp., *Epimastopora* sp., *Costacheoides* sp. indet. (Figure 109D), *Komia eganensis* Wilson, Wainess, and Coogan (Figure 106B), and *Efluegelia johnsoni* (Flügel) (Figure 108N).

The Holder foraminifers are diverse. The Fusulinata are *Insolentitheca* sp. (Figure 108H), *Pseudovidalina multihelicis* Pinard and Mamet (Figure 110K), *P. cf. modificata* (Potievskaya) (Figure 110L), *Endothyra* sp., *Endothyranella* sp. (Figure 108G), *Timanella* sp. indet. (Figure 109H), *Pseudojanischewskina*? sp. 3 (Figure 109E,G), *Bradyina* cf. *lucida* Morozova (Figure 108C), *B. sp. 4* (Figure 108L,M), *B. sp. 5* (Figure 110A), *Bradyinelloides* sp. 2 (Figure 110C), *Palaeotextularia* sp., *Climacammina* sp., *Spireitlina* sp., *Tetrataxis paraconica* Reitlinger, *Polytaxis* ex gr.

FIGURE 103. (Opposite) Calcareous microfossils from the Gobbler Formation (Space History A [samples SHA] and B [samples SH]). (A) *Seminovella* sp. Subaxial section. Sample SHA 3-1. Morrowan or early Atokan. (B) *Millerella extensa* Marshall. Axial section. Sample SHA 3-7. Morrowan or early Atokan. (C) *Millerella pressa* Thompson. Axial section. Sample SHA 3-24. Morrowan or early Atokan. (D) *Eostaffella* cf. *ampla* Thompson. Axial section. Sample SHA 14-5. (E) *Eostaffella* sp. 1. Axial section. Sample SH 70-36. Early Atokan. Morrowan or early Atokan. (F, G) *Profusulinella* sp. Two subtransverse sections. (F) SH 75-16. (G) Sample SH 24-14. Early Atokan. (H) N. gen. 1 cf. *Pseudolituotuba* sp. Transverse section. Sample SH 24-19. Morrowan or early Atokan. (I, J, N, S) N. gen. 2, cf. *Eoglomospiroides* sp. Four random sections. (I) Sample SH 24-27. (J) Sample SH 24-22. (N) Sample SH 24-6. Morrowan or early Atokan. (S) Sample SH 78-9. Early Atokan. (K, L) *Planoendothyra* sp. Two axial sections. (K) Sample SH 11.5. Morrowan or early Atokan. (L) Sample SH 38-7. Early Atokan. (M) *Syzrania bella* Reitlinger. Longitudinal section. Sample SH 41-16. Early Atokan. (O) *Bradyina* ex gr. *lepida* Reitlinger. Axial section. Sample SH 70-9. Early Atokan. (P) *Orthovertella*? sp. Subaxial section. Sample SH 41-2. Early Atokan. (Q, U) *Syzrania* sp. Two sublongitudinal sections. (Q) Sample SH 41-7. Early Atokan. (U) Sample SH 94-8. Early or late Atokan. (R) *Syzrania confusa* Reitlinger. Axial section of the proloculus. Sample SH 108-7. Early or late Atokan. (T) N. gen. 3 aff. *Plumerinella* sp. Transverse section. Sample SH 108-7. Early or Late Atokan. Scale bars = 0.5 mm in A, 0.1 mm in B–U.

maxima (Schellwien), *Globivalvulina* sp., *Nankinella*? sp., *Triticites* cf. *borregoensis* Wilde (Figure 110N), *Triticites cellamagnus* Thompson and Bissell (Figure 108I,J), *Triticites* cf. *creekensis* Thompson (Figure 110H), and *Leptotriticites hughesensis* (Thompson) (Figure 109A).

The Miliolata are *Pseudospira*? sp., *Calcivertella* spp., *Calcitornella* sp. (Figure 109C), *Orthovertellopsis* sp., *Palaeonubecularia* sp. (Figure 108I), *Latitubiphytes* sp. indet. (Figure 109B), *L.* cf. *rauzeriae* (Vachard and Moix) (Figures 108I?, 110G,I), *L.* sp. 4 (Figure 109J), *Hedraites*? sp., *Ellesmerella* cf. *permica* (Pia) (Figure 110F), *Turrispiroides* sp. 2 (Figure 108F), *Cornuspira* sp., *Hemigordiellina* sp., and *Hemigordius* sp. 4 (Figure 108D,E). The Nodosariata are less diverse: *Syzyrania* sp. 4 (Figure 108K), *S. bulbosa* Groves (Figure 109E), *S. bella* Reitlinger (Figure 110B), *S.*? sp. (Figure 110E), *Protonodosaria* sp. 1 (Figure 110D), *P.* sp. 2 (Figure 110J), and *P.* sp. 3.

Holder Formation Foraminiferal Biostratigraphy

The Holder Formation is Virgilian because of the LOs of advanced *Triticites*, primitive *Leptotriticites*, *Protonodosaria*, and *Pseudovidalina* spp. Possible elements of further biozonation of these Virgilian strata could be the LOs of the species: *Triticites cellamagnus* (sample HB 22, Beeman Canyon, and SHB 332, Space History B), *Leptotriticites hughesensis* (HB 25, Beeman Canyon), *Triticites creekensis* (SHD 42, Space History D), *Protonodosaria* spp. (SHD 98, Space History D), and *Triticites borregoensis* (SHD 104, Space History D). Detailed references about these taxa may be found in Thompson (1954), Pinard and Mamet (1998), Wilde (2006), Wahlman (2013), Vachard et al. (2013a), and Krainer et al. (2017b).

Wahlman and Barrick (2018) assigned an early Virgilian age to the lower Holder Formation on the basis of fusulinids. Our collections include *T. borregoensis* from near the top of the Holder Formation, and it is a late Virgilian species according to Wilde (2006). The fusulinids thus indicate that the Holder

Formation spans most of the Virgilian, although the earliest Virgilian strata are in the upper Beeman Formation below it, and the youngest Virgilian strata are in the lower Bursum Formation above it.

Summary

Algae, relatively rare except for *Komia*, are well documented in the New Mexico Pennsylvanian section (Krainer et al., 2003, 2009a, 2009b, 2011, 2015, 2017b, 2017c; Vachard et al., 2012; Lucas et al., 2015, 2016a, 2016b). The Gobbler Formation contains coralline red algae, *Archaeolithophyllum missouriense*, and the oldest dasycladalean green algae of the succession. The early Atokan interval of the Gray Mesa Formation contains well-preserved cyanobacteria, *Nostocites vesiculosus*. The Beeman Formation samples did not contain algae. The Holder Formation algal flora is diversified, with various epimastoporacean taxa. Many of the algae identified in the Newell Peak section of the Pennsylvanian–lower Permian Horquilla Formation in the Big Hatchet Mountains of southwestern New Mexico (Krainer et al., 2017b) are, nevertheless, absent from the material of this study.

The foraminifers from the Pennsylvanian section in the northern Sacramento Mountains also are well known in the upper Paleozoic strata of New Mexico, with the genera *Eotuberitina*, *Tuberitina*, *Earlandia*, *Monotaxinoides*, *Pseudovidalina*, *Endothyra*, *Endothyranella*, *Bradyina*, *Palaeotextularia*, *Climacamina*, *Tetrataxis*, and *Polytaxis*. Locally, the bradyinid genera *Bradyina*, *Bradyinelloides*, and *Pseudojanischewskina* seem to be relatively diversified, as well as some uncommon genera of fusulinids (*Annulofusulinella* and *Tumulotriticites*), described by Wilde (2006), which are possibly junior synonyms of *Putrella* and *Protriticites*, respectively. More classical genera of fusulinids are present with *Eostaffella*, *Millerella*, *Pseudoacutella*, *Profusulinella*, *Beedeina*, *Triticites*, and *Leptotriticites* (see Vachard et al., 2013a; Krainer et al., 2017b).

FIGURE 104. (Opposite) Calcareous microfossils from the Gray Mesa Formation at the High Rolls section (samples HRO). (A, F, G, H) *Komia eganensis* Wilson, Wainess, and Coogan. Late/latest Atokan or earliest Desmoinesian. (A) Transverse section with well-preserved skeletal network of pillars and laminae. Sample HRO 1-3. (F) Numerous sections in a bioclastic wackestone. Sample HRO 2-7. (G) Numerous sections in a peloidal and bioclastic grainstone. Sample HRO 2-7. (H) Longitudinal section with the circular emplacement of a cylindrical substrate. Sample HRO 4-2. (B) *Trepeilopsis* sp. Late/latest Atokan or earliest Desmoinesian. Longitudinal section. Sample HRO 1-4. (C) *Globivalvulina* sp. Late/latest Atokan or earliest Desmoinesian. Subtransverse section. Sample HRO 1-5. (D) *Pseudoacutella* sp. Axial section. Sample HRO 1-6. (E) *Tetrataxis acuta* Durkina. Late/latest Atokan or earliest Desmoinesian. Axial section. Sample HRO 2-3. (I) *Monotaxinoides transitorius* Brazhnikova and Yartseva. Late/latest Atokan or earliest Desmoinesian. Axial section. Sample HRO 6-4. (J) *Stacheoides* sp. Late/latest Atokan or earliest Desmoinesian. Longitudinal section. Sample HRO 4-5. (K) *Wedekindellina* sp. Recrystallized axial section. Sample HRO 10-1. Early Desmoinesian. (L) *Orthovertellopsis* sp. indet. Late/latest Atokan or earliest Desmoinesian. Oblique section. Sample HRO 8-1. Scale bars = 1 mm in F, G, H, K, 0.1 mm in all other images.

The Miliolata are diversified as attached taxa of the families Glomospiridae and Calcivertellidae: *Pseudospira*?, *Calcivertella*, *Calcitornella*, *Orthovertella*?, *Glomospiroides*? sp., *Plumerinella*? sp., *Orthovertellopsis* sp., *Palaeonubecularia* sp., *Latitubiphytes*, *Hedraites*? sp., *Ellesmerella*, *Turrispiroides*, *Cornuspira* sp., *Hemigordiellina* sp., and *Hemigordius*. It is noteworthy that the genus *Orthovertellopsis*, recently described from the lower Permian of New Mexico by Vachard et al. (2013b), appears regionally as early as the Desmoinesian. Free Miliolata are relatively rare and represented by the classical *Turrispiroides*, *Cornuspira*, *Hemigordiellina*, and *Hemigordius*. The Nodosariata, the most evolved foraminiferal class, is represented by only *Syzrania* and *Protonodosaria*.

Biostratigraphically, the Pennsylvanian succession in the northern Sacramento Mountains can be summarized as follows:

1. Rare Morrowan? deposits occur in the lowermost part of the Gobbler Formation.
2. Early Atokan (*Profusulinella* zone) is well represented in the Gobbler Formation.
3. Late Atokan is poorly characterized in the Gobbler Formation, with an apparent absence of the genus *Fusulinella*.
4. Latest Atokan–early Desmoinesian is the age of the lower part of the Gray Mesa Formation, with *Beedeina* and *Wedekindellina*.
5. Late Desmoinesian?–Missourian in the Beeman Formation is characterized by *Tumulotriticites* (= *Protriticites*) and the transitional stages between *Palaeonubecularia* and *Latitubiphytes*.
6. The Virgilian Holder Formation is well characterized by evolved *Triticites*, primitive *Leptotriticites*, and the first occurrences of *Pseudovidalina* and *Protonodosaria*.

CONODONTS

We sampled for conodonts at most of the stratigraphic sections we measured. Conodont recovery was good, and age

diagnostic conodonts were recovered for some intervals of the Gobbler, Gray Mesa, and Beeman Formations (Figures 111–118).

Underlying Mississippian Strata

At the Space History A section (Figure 11), the Gobbler Formation rests on strata of the Mississippian Lake Valley Formation, the uppermost bed of which produced a small fauna dominated by *Polygnathus communis communis* Branson and Mehl, 1934, with a few small *Gnathodus* specimens. At Mule Canyon, the lower part of the limestone unit 3–5 m above the cliff formed by the Mississippian Lake Valley Formation (Figure 9, units 6–7) produced several *Rhachistognathus*, a *Cauvugnathus* species, and examples of the *Gnathodus texanus* group. Although species of *Rhachistognathus* are usually indicative of Morrowan strata, the presence of *Cauvugnathus* and the *G. texanus* Roundy, 1926 group clearly indicates a Late Mississippian age.

Lane (1974) reported this assemblage from the Rancheria Formation (his Fauna 9), and Baesemann and Lane (1985) discussed the age of this assemblage, assigning this oldest occurrence of a “*Rhachistognathus*” species a late Meramecian age. Later, Lane and Brenckle (2005) stated that the age of Faunal Unit 9 (Lower *bilineatus* Zone) is early Chesterian in age. Lane (1974: fig. 4) showed that the most northern extent of the Las Cruces–Rancheria–Helms wedge comprises strata with Faunal Unit 9 in Alamo Canyon, a short distance south of our Space History A section (Figure 2). This limestone unit at Mule Canyon appears to be part of the tip of this Late Mississippian wedge.

Gobbler and Gray Mesa Formations

At the Space History A section, a thin limestone that lies 5 m above the top of the Lake Valley Formation (unit 3) yielded several representatives of *Adetognathus lautus* (Gunnell, 1933), *Rhachistognathus muricatus* (Dunn, 1965) (Figure 112B–D), and possibly *Rh. primus* Dunn, 1966 (Figure 112A). Lane

FIGURE 105. (*Opposite*) Additional calcareous microfossils from the Gray Mesa Formation at the High Rolls section (samples HRO). (A) *Orthovertellopsis* sp. Longitudinal section. Sample HRO 21a-2. Early Desmoinesian. (B) *Syzrania bella* Reitlinger. Longitudinal section. Sample HRO 3-1. Late Atokan. (C) *Cornuspira* sp. Transverse section. Sample HRO 2-1. Late Atokan. (D) *Nipperella* cf. *matura* (Thompson). Axial section. Sample HRO 13a-6. Early Desmoinesian. (E) *Beedeina schellwieni* (Staff). Axial section. Sample HRO 17-2. Early Desmoinesian. (F, G) *Turrispiroides multivolutus* (Reitlinger). Two subaxial sections. (F) Sample HRO 19-7. (G) Sample HRO 19-5. Early Desmoinesian. (H) *Trepeilopsis* sp. Longitudinal section. Sample HRO 6-2. Late Atokan. (I) *Bradyinelloides* sp. Axial section. Sample HRO 4-3. Late Atokan. (J) *Nostocites vesiculosa* Maslov. Random section. Sample HRO 12-5. Early Desmoinesian. Scale bars = 1 mm in E, 0.1 mm in all other images.

(1974) reported the presence of Faunal Unit 12 conodonts, the earliest Morrowan fauna in the first carbonate (level not specified) above the base of the Gobbler in Dog Canyon and at the base of the Gobbler at Alamo and Grapevine canyons. Benne (1975) recovered Faunal Unit 12 (identified by Lane) at Deadman Canyon from a limestone 30 feet (10 m) above the base of the Gobbler. Baesemann and Lane (1985) discussed the problems of distinguishing specimens of *Rh. muricatus* from *Rh. primus* and noted that their own criteria had changed since Lane (1974). Nearly all of our *Rhachistognathus* specimens possess a medial trough, which is diagnostic of *Rh. muricatus* (Baesemann and Lane, 1985), but at least one specimen appears to lack the trough, which would make it *Rh. primus*. The distinction is important, for although many forms of *Rhachistognathus* range from the Morrowan into the early Atokan, *Rh. primus* ranges from the base of the Morrowan and up only as high as the early Morrowan *Idiognathoides sinuatus* Zone (Baesemann and Lane, 1985).

The next oldest Pennsylvanian conodont fauna in our sections appears at Mule Canyon, 19 m above the highest Rancheria limestones (Figure 9, unit 18). *Idiognathodus sinuosus* Ellison and Graves, 1941 (Figure 111V,W), *Neognathodus bassleri bassleri* Harris and Hollingsworth, 1932 (Figure 111S,T), *Idiognathoides* sp., and possibly *Idiognathodus parvus* (Dunn, 1966) (Figure 111U) and *Idiognathodus incurvus* Dunn, 1966 occur in a small fauna. The first two species appear no lower than the middle of the Morrowan but may range up into lower Atokan strata. The first occurrence of *I. parvus* lies somewhat below the base of the Atokan (e.g., Manger and Sutherland, 1984; Davis et al., 1994), and *I. incurvus* may appear near the base of the Atokan (Barrick et al., 2013a). The next highest sample at Mule Canyon (Figure 9, unit 26) yielded abundant *Adetognathus* elements, a few *Idiognathodus* elements of the *I. parvus*–*I. klapperi* Lane and Straka, 1974 types, and possibly *I. incurvus* (Figure 111G,H).

A diverse and abundant likely middle Atokan conodont fauna appears higher at Mule Canyon in unit 35 (Figure 9). A variety of *Idiognathoides* morphotypes, most of which can be assigned to

Id. sinuatus (Harris and Hollingsworth, 1932) (Figure 111L–N) or *Id. ouachitaensis* (Harlton, 1933) (Figure 111O), are present, as are a few *Diplognathodus* (Figure 111A,B). Three or more morphotypes of *Idiognathodus* occur. Discrimination and comparisons of Atokan species of *Idiognathodus* are difficult because of a general lack of taxonomic study. In North America, some authors have called Atokan *Idiognathodus* P₁ elements *I. delicatus* Gunnell, 1931 (e.g., Grayson, 1984; Brown et al., 2013), but this is the name of a late Desmoinesian species (see below), and others have used a less specific name, *I. aff. delicatus* (e.g., May et al., 2009). Grayson et al. (1990) assigned Atokan P₁ elements to two species, *I. klapperi* and *I. incurvus*, whereas Barrick et al. (2004, 2013b) tried to avoid the problem, using the term “*I. incurvus* descendents.” More recent work on Atokan sections from other parts of New Mexico suggests that several Atokan species of *Idiognathodus* may be present, but their taxonomy has yet to be resolved (Saelens, 2014; Treat, 2014; Moore, 2017).

One *Idiognathodus* morphotype, *I. incurvus*? (Figure 111E,Q), is gently incurved, has a pointed dorsal tip, and possesses a well-developed rostral lobe and a weakly developed rostral lobe of a few nodes. This morphotype is similar to the holotype of *I. incurvus*, but the dorsal tip is not as sharply incurved as in the holotype. The second, more common morphotype has a more slender outline, is less curved, and lacks caudal and rostral lobes or nodes. The rostral adcarinal ridge is high and merges with the rostral platform margin. Moore (2017) used the name *Idiognathodus* sp. Q to designate similar morphotypes from the Sandia Mountains (Figure 111C,D). The third common morphotype has a short, wide platform with a blunt dorsal tip. The greatest width of the platform lies near the middle of the platform. A small caudal lobe is present, but no rostral lobe or nodes. The rostral adcarinal ridge is high and extends onto the dorsal surface of the platform (Figure 111E,I–K,P,R). This morphology is similar to that described for the late Atokan species *I. gibbus* Lambert, 1992, and it probably comprises part of the same species group. Specimens like these from the Gobbler Formation have been illustrated from the middle part of the Red House Formation

FIGURE 106. (*Opposite*) Calcareous microfossils from the Gray Mesa Formation at the Space History B section (samples SH). (A, C, K) *Latitubiphytes* spp. Latest Desmoinesian/early Missourian. (A) Longitudinal section. Sample SH 160-1. (C) Transverse section. Sample SH 160-6. (K) Subtransverse section. Sample SH 167-8. (B) *Palaeonubecularia* cf. *rustica* Reitlinger. Subaxial section. Sample SH 160-2. Latest Desmoinesian/early Missourian. (D) *Anthracoporellopsis novamexicana* Vachard, Krainer, and Lucas, 2012. Longitudinal section. Sample SH 153-5a. Latest Desmoinesian/early Missourian. (E) *Turrispiroides* sp. Subaxial section. Sample SH 149-1. Latest Desmoinesian/early Missourian. (F, G) *Hemigordius* cf. *simplex* Reitlinger. Two subaxial sections. Latest Desmoinesian/early Missourian. (F) Sample SH 160a-1. (G) Sample SH 163-2. (H) *Orthovertellopsis* sp. Longitudinal section. Sample SH 168-2. Latest Desmoinesian/early Missourian. (I) *Hemigordius* cf. *discoideus* (Brazhnikova and Potievskaya). Subaxial section. Sample SH 160a-3. Latest Desmoinesian/early Missourian. (J) *Hemigordius* ex gr. *harltoni* Cushman and Waters emend. Groves. Subaxial section. Sample SH 167-5. Latest Desmoinesian/early Missourian. Scale bars = 1 mm in A, B, 0.1 mm in all other images.

at Whiskey Canyon (Saelens, 2014; Lucas et al., 2016a). The older form, here assigned questionably to *I. parvus*, may also belong to the *I. gibbus* group.

At the Space History A section (Figure 11), samples from 33 to 43 m above the base of the Gobbler (units 16, 18, 20) lack examples of *Idiognathoides*, *Neognathodus* species are rare, and the faunas are strongly dominated by *Idiognathodus* species similar to those occurring at Mule Canyon. *Idiognathodus incurvus* morphotypes are common but are variable in platform shape and curvature (Figure 112F–H,K–O,Q–S,V,W). Slender P_1 elements like *Idiognathodus* sp. Q are present, but most are a little broader and have a single rostral node (Figure 112I,J). The dominant morphotypes have robust, wide, short platforms that are wider dorsally and constrict in width on the ventral platform margin (Figure 112P,T,U). These tend to have rostral nodes in addition to a well-developed caudal lobe. The platform outline is variable, ranging from longer forms to short forms like the *I. gibbus* group, but the rostral margin is much more heavily nodose than either of those forms. These wide forms also resemble some P_1 elements illustrated as *I. sp. G* by Saelens (2014) from the upper part of the Red House Formation in south central New Mexico.

The only *Neognathodus* fragments recovered suggest species like *N. atokaensis* Grayson, 1984 or *N. uralicus* Nemirovskaya and Alekseev, 1994, both middle Atokan species (Figure 112E). *Idiognathodus* sp. Q occurs with and above the range of *N. atokaensis* in the Sandia Mountains (Moore, 2017) and *I. sp. G* ranges higher into the upper part of the Atokan (Saelens, 2014; Moore, 2017). A similar conodont fauna occurs in the Indian Wells Canyon section, 22 to 25 m above the base of the Gobbler (Figure 13A, units 7, 9). *Idiognathodus incurvus* morphotypes are present (Figure 113Y–AB), as well as a different group of P_1 elements with short, wide, nodose platforms with well-developed lobes that may be related to *Idiognathodus* sp. G (Figure 113V–X). These conodont faunas in our Space History A and Indian Wells Canyon sections appear to be middle to possibly late Atokan in age.

Conodont recovery was very poor from higher samples from Space History A and from the next 70 m at Indian Wells Canyon. At Space History A, a more typical specimen of the

late Atokan species *Idiognathodus gibbus* was found at 48 m (Figure 11, unit 26). One specimen of *Neognathodus colombiensis* (Stibane, 1967) (or *N. bothrops* Merrill, 1972; specimen is broken) was obtained at 93 m (SHA-36). Both of these *Neognathodus* species appear near the Atokan-Desmoinesian boundary. From units 42 and 44, a few examples of *Idiognathodus* sp. H occur, which appears near and above the base of the Desmoinesian elsewhere in New Mexico (Saelens, 2014; Treat, 2014; Moore, 2017). A single *N. colombiensis* was recovered 97 m above the base of the Gobbler Formation at Indian Wells Canyon (Figure 13, unit 51). Small faunas with *I. sp. H* occur from 134 to 139 m at Indian Wells Canyon (Figure 13, units 62, 64).

The beds near the base of the Space History B section (Figure 12A, 1 m above the base, unit 1; 4 m, unit 3; and 25 m, unit 9) are early Desmoinesian in age. Sparse examples of *Neognathodus colombiensis*? (Figure 113H) and *N. bothrops* (Figure 113M) occur, some of which display the slight reduction of platform symmetry found in the early Desmoinesian, early Cherokee, faunas (Figure 113J,L). The most common form of *Idiognathodus* possesses a very slightly curved, elongate platform that tapers gradually to the dorsal tip, features of the early Desmoinesian species *I. amplificus* Lambert, 1992 (Figure 113N–Q). A few broad specimens with inclined fine transverse ridges are *I. obliquus* Kozitskaya and Kossenko, 1978 (in Kozitskaya et al., 1978) (Figure 113R–T). Slightly higher, at 39 m (Figure 12A, unit 18), *N. asymmetricus* (Stibane, 1967) (Figure 113I) appears and occurs with *Idiognathodus* morphotypes like *I. attenuatus* Youngquist and Heezen, 1948 (Figure 113F,G). Shorter, broader forms fall into the *I. rectus* Youngquist and Downs, 1949–*I. iowaensis* Youngquist and Heezen, 1948 group (Figure 113A–E). These are early, but not earliest, Desmoinesian assemblages (lower-middle Cherokee Group equivalents; Barrick et al., 2013a).

Early late Desmoinesian conodonts appear in the Space History B section at about 64 m (Figure 12A, units 41, 43, 46). In this interval, *Idiognathodus* specimens occur that are like those described as *I. acutus* Ellison, 1941 (Figure 114A–D,F–I,L,M,U–Z), from the lowermost Marmaton cyclothem, the Lower Fort

FIGURE 107. (Opposite) Calcareous microfossils from the Beeman Formation (Space History B [samples SHB] and Space History C [samples SHC]). (A) *Tumolotriticitis* (= *Protriticitis*) sp. Subaxial section. Sample SHC 163-4. Middle? Missourian. (B, D) *Palaeonubecularia* sp. 2. (B) Oblique section. Sample SHB 183-17. (D) Longitudinal section. Sample SHC 40-15. Missourian. (C) *Palaeonubecularia* sp. 3. Longitudinal section. Sample SHB 209-3. Missourian. (E) *Globivalvulina* sp. Subtransverse section. Sample SHC 163-13. Middle? Missourian. (F) *Syzrania bella* Reitlinger. Axial section. Sample SHB 209-6. Late? Missourian. (G) *Glomospiroides* sp. Transverse section. Sample SHC 56-1. Missourian? (H) *Orthovertellopsis*? sp. Oblique section. Sample SHB 183-2a. Late? Missourian. (I, J) *Pseudojanischewskina* sp. 1. (I) Transverse section. Sample SHB 183-12. (J) Transverse section. Sample SHB 183-22. Late? Missourian. (K) *Pseudojanischewskina* sp. 2. Subtransverse section. Sample SHB 183-20. Late? Missourian. (L) *Pseudospira* sp. Random section. Sample SHB 232-6a. Late? Missourian. Scale bars = 0.1 mm.

Scott (Excello Shale; Nestell et al., 2016). In unit 43, *Gondolella wardlawi* Nestell and Pope, 2016 (in Nestell et al., 2016) (Figure 114R,S), which first appears in the Excello Shale, is present. *Neognathodus roundyi* (Gunnell, 1931) (Figure 114K,N,P,Q), which first appears in the underlying Verdigris cyclothem in the Midcontinent, also occurs. Some *Neognathodus* specimens retain complete platform margins, unlike *N. roundyi*, and may be early forms of *N. expansus* (Jones, 1941) (Figure 114O), which is better known from younger Desmoinesian strata (Rosscoe, 2005). A similar late Desmoinesian *Idiognathodus*-dominated conodont fauna was recovered from Indian Wells Canyon, 197 m above the base (IWC 93; Figure 116X–Z).

From 85 through 93 m in the Space History B section (Figure 12A, units 75, 82, 84, 86), typical Marmaton (late Desmoinesian) faunas occur that are strongly dominated by *Idiognathodus* morphotypes. Marmaton *Idiognathodus* species are poorly described, and most have been generally assigned to an older species, *I. delicatus* Gunnell, 1931, or a younger species, *I. expansus* Stauffer and Plummer, 1932 (e.g., Ritter et al., 2002; Barrick et al., 2013a,b). Our collections include some morphotypes like *I. delicatus* specimens obtained from topotype collections, but most are closer in features to *I. expansus*. The *I. delicatus* morphotypes have fewer transverse ridges and less well developed lobes (Figure 115L,M,O,P,S–U,W) compared to *I. expansus*, in which a more curved dorsal platform has more lobes and more closely spaced transverse ridges and the lobes are better developed and more nodose (Figure 115H–K,N,V,X). Other *Idiognathodus* morphotypes are also present (Figure 115Q,R). *Neognathodus* species include the typical Marmaton species, *N. roundyi* (Figure 115C–G) and *N. dilatus* Stauffer and Plummer, 1932, both with strongly reduced margins, and *Neognathodus expansus*-like morphotypes (Figure 115A,B). Few conodonts were recovered from 93 to 218 m at the Space History B section. The highest definitive Desmoinesian species in this section (Figure 12A), a specimen of *Neognathodus dilatus*, was obtained from unit 118, 135 m above the base of the section.

Few conodonts were recovered from the Gray Mesa Formation at High Rolls (Figure 20), either from the section of LaPeter (2007) or from our newer collections. The lower

samples (LaPeter, 2007: pl. 2, sample 1-7.9; Figure 20, unit 2) in the sections yielded a few specimens of *Neognathodus* that are transitional from *N. colombiensis* to *N. roundyi*, which are interpreted to be late Cherokee in age (see Moore, 2017, for a discussion of these forms). Typical *N. roundyi* appears 50 m higher in the High Rolls section (Figure 20, unit 38) and 16 m higher in the LaPeter section (sample 1-24). The *Idiognathodus* morphotypes associated with *N. roundyi* are like those found in other Marmaton faunas in the upper Gobbler Formation. The uppermost samples of the Gray Mesa Formation produced mostly tiny undiagnostic *Idiognathodus* and *Adetognathus* elements.

The youngest Desmoinesian conodont fauna was recovered from the Indian Wells section (Figure 13, unit 134), 20 m below the top of the Gobbler Formation. A series of morphotypes occurs that generally fits into the morphological spectrum of the late Desmoinesian species *Swadelina neoshoensis* Lambert, Heckel, and Barrick (2003) (Figure 116T–V,AA,AB) and may include a few *Sw. nodocarinata* (Jones, 1941). In addition to the more characteristic long morphotypes, short, stubby forms bearing just a few transverse ridges are present (Figure 116R,W,AC). Few elements of *Idiognathodus expansus* and rare *Neognathodus roundyi* (Figure 116M,N) occur. This fauna can be correlated approximately with the *Sw. neoshoensis* Zone of the Midcontinent region, which is the next to last Desmoinesian conodont zone (Barrick et al., 2013a).

Early Missourian conodonts appear in carbonate beds in the uppermost 25 m of the Gobbler Formation in the Space History B, Space History C, and Beeman Canyon A sections. The lowest Missourian fauna appears 22 m below the top of the Gray Mesa Formation at Space History B (Figure 12B, unit 155), where *Idiognathodus cancellosus* Gunnell, 1933 (Figure 116A) occurs with *I. turbatus* Rosscoe and Barrick, 2009 (Figure 116E), *I. swadei* Rosscoe and Barrick, 2009, and similar species. Slightly higher samples 10 to 12 m below the top of the Gobbler (Figure 12B, unit 161, Figure 25, unit 1; Figure 27, units 2, 4) yielded a more abundant and diverse, but comparable, fauna (Figure 116B–D,F–L,O–Q). This morphologically diverse fauna that includes *I. cancellosus*

FIGURE 108. (Opposite) Calcareous microfossils from the Holder Formation (Holder bioherm section, samples HB). (A) *Ivanovia tenuissima* Khvorova. Several thalli in longitudinal sections. Sample HB 3-4. Virgilian. (B) *Komia eganensis* Wilson, Waines, and Coogan. Several oblique sections. Sample HB 5-1. Virgilian. (C) *Bradyina* cf. *lucida* Morozova. Subtransverse section. Sample HB 3-2. Virgilian. (D, E) *Hemigordius* sp. 4. Two subaxial sections. (D) Sample HB 9-2. (E) Sample HB 9-12. (F) *Turrispiroides* sp. 2. Axial section. Sample HB 12-1. Virgilian. (G) *Endothyranella* sp. Oblique subtransverse section. Sample HB 12-2. Virgilian. (H) *Insolentitheca* sp. Longitudinal section. Sample HB 5-4. Virgilian. (I, J) *Triticites cellamagnus* Thompson and Bissell. (I) Two subtransverse sections with *Eugonophylum* sp., *Palaeonubecularia* sp., and *Latitubiphytes* sp. Sample HB 22-9. (J) Subaxial section. Sample SHB 332-5. Virgilian. (K) *Syzrania* sp. 4. Axial section. Sample SHD 5-9a. Virgilian. (L, M) *Bradyina* sp. 4. (L) Subaxial section. Sample SHD 32-3. (M) Subtransverse section. Sample SHD 32-6a. Virgilian. (N) *Efluegelia johnsoni* (Flügel). Axial section. Sample SHD 32-7. Virgilian. Scale bars = 1 mm in A, B, I, K, 0.1 mm in all other images.

is characteristic of the lower Missourian Swope cyclothem (Hushpuckney Shale; Rosscoe and Barrick, 2013) and the corresponding *I. cancellosus* Zone (Barrick et al., 2013b) in the Midcontinent region.

Beeman and Holder Formations

Previously, Raatz et al. (1994), Raatz (1996), and Raatz and Simo (1998) had placed the base of the Beeman Formation at a thin dolostone bed within the siliciclastic section that lies above the uppermost Gobbler limestone in the Horse Ridge area. Raatz et al. (1994) reported a small ammonoid fauna from a siltstone in Indian Wells Canyon that he placed in the lower part of the Beeman Formation (his cycle 1S-1a), and later Raatz (1996) reported conodonts. The ammonoid fauna included species of *Eoschistoceras*?, *Glaphyrites*, and *Neoaganides* (see Boardman et al., 1994:44). Raatz et al. (1994) and Raatz (1996:172) proposed that this fauna was latest Desmoinesian in age. However, *Neoaganides* first appears in lower Missourian strata in the Midcontinent region, either near the base of the Missourian (Boardman et al., 1994: fig. 16) or only as low as the Dennis cyclothem (Boardman and Work, 2013: text-fig. 1). The conodonts were collected later by Steve Schutter and Barrick from a shaly bed below a limestone ledge, and the exact stratigraphic position is unknown, but they were likely in the upper part of the Gobbler. The only conodonts identified were a generalized form of *Idiognathodus*, questionably identified as *I. expansus*, a species known from highest Desmoinesian strata. We include the section of siliciclastic units above the uppermost Gobbler Formation limestone in the Horse Ridge area in the lower part of the Beeman Formation (Horse Ridge LA and all of the 1S cycles of Raatz et al., 1994; Raatz, 1996; and Raatz and Simo, 1998). This entire siliciclastic interval, from which all the plant fossils of the Beeman were collected, is early Missourian in age by its position above the uppermost Gobbler limestone and below the overlying middle Missourian Beeman carbonate section. It appears to span only the time represented by two Midcontinent cyclothem (Dennis and Hogshooter).

The lowest conodont faunas in our collections from the Beeman Formation appear with the occurrence of sandy

limestones 65 m above the base of the formation at the Space History C section (Figure 25, top of unit 44) and 87 m above the base at the Space History B section (Figure 12C, unit 212). In Space History C unit 44, the most common forms are examples of *Idiognathodus symmetricus* Gunnell, 1933 (Figure 117N–P,R–U), as characterized by Rosscoe (2008), in association with representatives of the *Streptognathodus excelsus* Stauffer and Plummer, 1932 group of species (Figure 117E,V,W), and *Hindeodus*. A few specimens of *I. magnificus* Stauffer and Plummer, 1932 (Figure 117M) and other *Idiognathodus* morphotypes also occur. A similar *Streptognathodus* fauna was recovered from Space History B unit 212 (Figures 12, 117A,C,D), but the *Idiognathodus* morphotypes, most of which can still be referred to *I. symmetricus* in the broadest sense, possess a slightly depressed dorsal platform. Both samples are no older than the middle Missourian Cherryvale cyclothem and could be as young as the overlying Dewey cyclothem in the Midcontinent region (Barrick et al., 2013a). These limestones should correspond approximately to the base of lowest carbonate cycle, 2C-1a, of Raatz (1996) and Raatz and Simo (1998). Raatz (1996) and Raatz and Simo (1998) reported that *I.* (as *Streptognathodus*) *confragus* occurs in 2C-1a on the basis of identifications by Barrick at that time. Since then, revisions of Missourian conodonts (e.g., Rosscoe, 2008; Rosscoe and Barrick, 2013) have clarified the taxonomy of early Missourian forms and have shown that the specimens from cycle 2C-1a previously called *I. confragus* are likely *I. symmetricus*.

At the Space History B section (Figure 12C), in two beds slightly higher in the Beeman limestone interval, units 229 (105 m) and 235 (108 m), large, well-preserved conodont faunas are dominated by two morphotypes of *Idiognathodus magnificus* (Figure 117B,G–J) that have been described by Hogancamp et al. (2017). The full spectrum of morphotypes of the *Streptognathodus excelsus* group occurs (*S. sulcatus* Gunnell, 1933, *S. gracilis* Stauffer and Plummer, 1932, and *S. elegantulus* Stauffer and Plummer, 1932), of which the lobed morphotypes (*S. excelsus*) are more common (Figure 117E,K,L,Q). A few *Hindeodus* and rare *Ellisonia* elements were recovered. Faunas strongly dominated by typical forms of *I. magnificus* are usually correlated with the Dewey cyclothem (e.g., Barrick et al.,

FIGURE 109. (Opposite) Additional calcareous microfossils from the Holder Formation (Holder bioherm [samples HB] and Space History D [samples SHD]). (A) *Leptotriticites hughesensis* (Thompson). Subaxial section. Sample HB 25-3. Virgilian. (B) *Latitubiphytes* sp. Indet. Transverse section. Sample SHD 21-10. Virgilian. (C). *Calcitornella* sp. Oblique section. Sample SHD 21-4. Virgilian. (D) *Costacheoides* sp. indet. Oblique section. Sample SHD 32-13. Virgilian. (E) *Syzrania bulbosa* Groves. Axial section in the proloculus. Sample SHD 32. Virgilian. (F, G) *Pseudojanischewskina*? sp. 3. (F) Subaxial section. Sample SHD 42-5. Virgilian. (G) Transverse section. Sample SHD 42-6. Virgilian. (H) *Timanella* sp. indet. Axial section. Sample SHD 44-3. Virgilian. (I) *Latitubiphytes* cf. *rauzeriae* (Vachard and Moix). Oblique section. Sample SHD 6-6. Virgilian. (J) *Latitubiphytes* sp. 4. Transverse section. Sample SHD 70-2. Virgilian. (K) *Bacinella* in a gastropod. Sample SHD 72-2. Virgilian. Scale bars = 1 mm in A, I; 0.1 mm in all other images.

2013b). In the overlying limestone beds, Space History B units 245 (122 m), 259 (129 m), and 267 (135 m), *Streptognathodus* species, mostly ones with small (*S. gracilis*) to no lobes (*S. elegantulus*) occur. This association is characteristic of the Missourian post-Dewey Midcontinent cyclothems in the upper part of the *S. gracilis* Zone, where *Streptognathodus* species dominate and *Idiognathodus* species are rare.

In Dry Canyon, along U.S. Highway 62, the *Streptognathodus gracilis* fauna continues to near the top of the carbonate-dominated section in the middle of the Beeman (Raatz, 1996, cycle 2C-4a). About 4 m higher in the overlying section of mixed siliciclastics and carbonates, a 1.5 m interval of dark shale with zones of *Dunbarella* molds (Dry Canyon 1 section of Raatz, 1996) produced an abundant conodont fauna dominated by morphotypes of *S. pawhuskaensis* Harris and Hollingsworth, 1932 (Figure 118Q,Y,Z), many of which possess a relatively long carina, apparently gradational with the older species *S. firmus* Kozitskaya in Kozitskaya et al. 1978 (Figure 118R,S). This fauna was originally reported to correlate with the South Bend cyclothem (Raatz, 1996, cycle 2C-5a), but Keairns (2002) suggested that this fauna could correlate with either the upper Missourian South Bend or the overlying Iatan cyclothem.

Conodont faunas from higher in the Beeman Formation at Dry Canyon and the nearby La Luz B section (Figure 26) are difficult to correlate precisely. A variety of morphotypes of *Streptognathodus pawhuskaensis* dominate the collections (Figure 118E,G,I-K,N-Q,U-Z) but provide little correlation information. Lower samples contain mostly slender unnoded morphotypes, but a few have a couple accessory nodes, and some show the deflection of the carina that Ritter (1994) named *S. pawhuskaensis deflectus*. Rare forms from higher in the section (La Luz B, top of unit 33) possess the squat platform of *S. bitteri* Ritter, 1994 (Figure 118L,M,T). Ritter (1995) indicated that *S. p. deflectus* and *S. bitteri* occur from the Oread cyclothem (early Virgilian; basal Gzhelian) to the Lecompton cyclothem, but their ranges are still poorly constrained. A more definitive faunal association appears near the top of the Beeman (La Luz B,

unit 47), where *Streptognathodus* elements grading from *S. vitali* Chernykh, 2002 to *S. virgolicus* Ritter, 1995 (Figure 118F) occur. This association is no older than the Lecompton cyclothem (Barrick et al., 2013a, 2013b).

Few conodonts were recovered from the Holder Formation. *Streptognathodus pawhuskaensis* (Figure 118B,H) was obtained at the Space History D section (Figure 34; unit 44, 52 m above the base; unit 82, 100 m above the base). At Long Ridge one sample ~57 m above the base of the Holder (Figure 24, unit 149) produced *S. pawhuskaensis* (Figure 118A,C) and *S. vitali* (Figure 118D). Keairns (2002) reported a similar Virgilian (Gzhelian) species association from the lower beds of the Holder Formation in Dry Canyon (cycles J to I2; Cox, 1998).

SOME BEEMAN FORMATION FOSSILS

The plant localities in the Beeman Formation have yielded some fossils of animals, namely, ostracods, conchostracans, limulids, insects, bivalves, and fishes. However, as noted above, we made no effort to collect these animal fossils intensively, so our collection of them is relatively small and lacks diversity. The largest collection is from the most extensively collected plant locality, USNM 43868 in Beeman Canyon (=NMMNH locality 12016). The second most extensively collected plant site yielded the second-largest collection of animal fossils, USNM 43834 on the south side of La Luz Canyon (=NMMNH locality 9109). A less intensively collected locality also yielded some animal fossils that were collected, USNM 43836 on the north flank of La Luz Canyon (=NMMNH locality 7981). It is fair to say, though, that all of the fossil plant localities in the Beeman Formation yielded some animal fossils, at least fossils of ostracods and/or conchostracans.

The conchostracans are the most abundant animal fossils we collected, but most are poorly preserved. The NMMNH collection includes numerous conchostracan fossils cataloged as NMMNH P-559787 (locality 7981) and 55978 (Figure 119B)

FIGURE 110. (Opposite) Additional calcareous microfossils from the Holder Formation (Space History D, samples SHD). (A) *Bradyina* sp. 5. Subaxial section. Sample SHD 69-5. Virgilian. (B) *Syzrania bella* Reitlinger. Axial section. Sample SHD 85-4. Virgilian. (C) *Bradyinelloides* sp. 2. Axial section. Sample SHD 85-13. Virgilian. (D) *Protonodosaria* sp. 1. Subaxial section. Sample SHD 93-2. Virgilian. (E) *Syzrania?* sp. Subaxial section. Sample SHD 93-4. Virgilian. (F) *Ellesmerella* cf. *permica* (Pia). Two oncoids. Sample SHD 98-4. Virgilian. (G, I) *Latitubiphytes* cf. *rauzeriae* (Vachard and Moix). (G) Oblique section. Sample SHD 100-5. (I) Sublongitudinal and oblique sections. Sample SHD 100-4. Virgilian. (H) *Triticites* cf. *creekensis* Thompson. Subaxial section. Sample SHD 42-17. Virgilian. (J) *Protonodosaria* sp. 2. Axial section. Sample SHD 98-5. Virgilian. (K) *Pseudovidalina multihelicis* Pinard and Mamet. Axial section. Sample SHD 104-3. Virgilian. (L) *Pseudovidalina* cf. *modificata* (Potievskaya). Subaxial section. Sample SHD 104-13a. Virgilian. (M) *Eugonophyllum johnsonii* Wray. Longitudinal section. Sample SHD 78-1. Virgilian. (N) *Triticites* cf. *borregoensis*. Wilde. Axial section. Sample SHD 104.8. Virgilian. Scale bars = 1 mm in C, F, G, H, M, N, 0.1 mm in all other images.

and 77189 from locality 9109. The well-preserved specimens display diagnostic features of *Lioestheria* and resemble species of that genus described by Martens and Lucas (2005) from the Missourian Tinajas Member of the Atrasado Formation in Socorro County, New Mexico. However, given the poor preservation of most of the Beeman conchostracans, we identify them only as *Lioestheria* sp.

The ostracods in the NMMNH collection are cataloged as NMMNH P-55958 (Figure 119A) and 55959 from locality 12106. They are unornamented, simple shells readily assigned to the Darwinulidae and likely to the form genus *Darwinula*.

Lucas et al. (2014b) have documented the limulid fossils, which were identified as *Euproops danae*. We have more complete material of this taxon now, including the cephalon (Figure 120), but it adds little to what Lucas et al. (2014b) concluded. All of these fossils are from NMMNH locality 9109 and are cataloged as NMMNH P-55975, 55976 (Figure 119D), 68701 (Figure 119C), and 68702 (Figure 119E). An important point, though, is that late Paleozoic limulids were apparently freshwater animals and thus are not indicative of marine or marginal marine conditions (Lamsdell, 2016).

Two kinds of bivalves are present, anthracosiids and myalinids. The anthracosiids can be assigned to *Palaeonodonta* (cf. Rinehart and Lucas, 2011). The specimens of *Palaeonodonta* in the NMMNH collection are from locality 9109 and are cataloged as NMMNH P-77189, 77330, 77332, 77334, 77339, 77345, and 77346 (Figure 120D). The myalinids (e.g., Figure 119F) have the short hinge line and acute terminal beaks characteristic of *Septamyalina* (cf. Kues, 2004b).

Microconchids have been found attached to plant material at more than one locality. A good example is USNM 706337 from USNM locality 43891 (Figure 119G). Recently, Gierlowski-Kordesch and Cassle (2015) argued that microconchids are indicative of marine habitats, but this is not fully agreed on (Gierlowski-Kordesch et al., 2016; Zatoń et al., 2016).

Insect wings are relatively common at NMMNH locality 12016, and most are of blattoids (cockroaches). The NMMNH collection has these specimens from locality 12016:

NMMNH P-55961, 55962, 55963, 55964 (Figure 120B), 55965 (Figure 120C), 55966 (Figure 120A), 55967, and 55968. From locality 9109, insect wings are cataloged as NMMNH P-77349 and 77350. Joerg Schneider (Bergakademie Freiberg, Germany, personal communication, 2019) has examined the blattoid wings and notes that they are very similar to those described by Schneider et al. (2017) from the Missourian Tinajas Member of the Atrasado Formation in Socorro County, New Mexico, which overlaps in age the Beeman Formation localities. Thus, he identifies NMMNH P-55966 (Figure 120A) as *Phyloblatta obscura* (Sellards), P-55964 (Figure 120B) as *Sysciophlebia* cf. *S. rubida* Schneider, and P-55965 (Figure 120C) as *S. sp.*, possibly *S. praepilata*. Schneider further notes that better-preserved wings, particularly with color patterns intact, are needed to make more definite taxonomic assignments.

The fish fossils are isolated scales, mostly incomplete. Specimens in the NMMNH collection are from locality 9109 and are cataloged as NMMNH P-55979, 77194, 77195, 77346 (Figure 120D), and 77347. They appear to be palaeoniscoid, but little more can be said about their identity.

A handful of spiral coprolites were collected and cataloged into the NMMNH collection. These are NMMNH P-77351 (Figure 120F) from locality 9109 and 53981 from locality 7981. The specimen illustrated appears to be assignable to *Crassocoprurus* because it is heteropolar and macrospiral and has about 10 spirals that make up about 75% of the length of the coprolite (cf. Hunt and Lucas, 2017). This is likely the coprolite of a shark, and it has been found previously in both freshwater and estuarine deposits in the Pennsylvanian of New Mexico (Hunt and Lucas, 2017).

All of these fossils are compatible with freshwater environments of deposition, especially the conchostracans and ostracods. However, the myalinid bivalves, microconchids, and shark coprolites could indicate brackish or shallow marine conditions at some of the plant localities. In general, it seems likely that most of the Beeman Formation plant localities are freshwater deposits, especially those with abundant conchostracans. However,

FIGURE 111. (Opposite) Conodonts from the lower part of the Gobbler Formation at Mule Canyon. Upper views of P_1 elements, except as noted. (A, B) *Diplognathodus* sp. Lateral views of P_1 elements. Mule Canyon, bed 35. NMMNH specimens (A) P-80100, (B) P-80101. (C, D) *Idiognathodus* sp. Q. Mule Canyon, bed 35. NMMNH specimens (C) P-80102, (D) P-80103. (E, I–K, P, R) *Idiognathodus gibbus* group morphotype. Mule Canyon, bed 35. NMMNH specimens (E) P-80104, (I) P-80105, (J) P-80106, (K) P-80107, (P) P-80108, (R) P-80109. (F, Q) *Idiognathodus incurvus*? Mule Canyon, bed 35. NMMNH specimens (F) P-80110, (Q) P-80111. (G, H) *Idiognathodus incurvus*? Mule Canyon, bed 26. NMMNH specimens (G) P-80112, (H) P-80113. (L–N) *Idiognathoides sinuatus*. Mule Canyon, bed 35. NMMNH specimens (L) P-80114, (M) P-80115, (N) P-80116. (O) *Idiognathoides ouachitaensis*. Mule Canyon, bed 35. NMMNH specimen P-80117. (S, T) *Neognathodus bassleri bassleri*. Mule Canyon, bed 18. NMMNH specimens (S) P-80118, (T) P-80119. (U) *Idiognathodus parvus*? Mule Canyon, bed 18. NMMNH specimen P-80120. (V, W) *Idiognathodus sinuosus*. Mule Canyon, bed 18. NMMNH specimens (V) P-80121, (W) P-80122. Scale bar = 0.25 mm.

further study is needed to determine the exact depositional environments of these sites.

CYCLES OF DEPOSITION

DISCUSSION

AGE AND CORRELATION

Ages can be assigned to the Pennsylvanian section in the northern Sacramento Mountains, and it can be correlated to other Pennsylvanian sections in New Mexico (Figure 121) primarily on the basis of the conodont biostratigraphy reported here and supplemented and reinforced by already published fusulinid biostratigraphy and that presented here. We summarize as follows:

1. Conodonts and fusulinids reported here from the Mule Canyon Space History A and B and Indian Wells Canyon sections suggest the lowermost Gobbler Formation may be of Morrowan age, but the rest of the formation is of Atokan–late Desmoinesian age.
2. The base of the Missourian is stratigraphically high in the Gray Mesa Formation on the basis of conodont biostratigraphy. Earlier authors (Pray, 1961; Raatz, 1996; Raatz and Simo, 1998) considered the base of the Missourian to be the Gobbler-Beeman contact, with the Gray Mesa strata included in the upper Gobbler as the Bug Scuffle Member. The Missourian base is thus stratigraphically a little lower than previously thought.
3. The upper Beeman Formation (Horse Ridge) yields conodonts that indicate it is of middle to late Missourian age.
4. The stratigraphically lowest Virgilian conodonts appear in the uppermost beds of the Beeman Formation, as do the stratigraphically lowest Virgilian fusulinids.
5. Fusulinids of the Holder Formation indicate an early late Virgilian age.

Several relatively recent publications identify glacioeustatic cycles in Pennsylvanian strata across New Mexico: in the Lucero uplift of Valencia County (Scott and Elrick, 2004; Elrick and Scott, 2010), in the Sandia uplift of Bernalillo County (Wiberg, 1993; Wiberg and Smith, 1994; Smith, 1999), in the mountain ranges of much of south central New Mexico (Soreghan, 1992, 1994a,b), and in the northern Sacramento Mountains of Otero County (e.g., Wilson, 1967, 1972a; Van Wagoner, 1977a,b; Algeo et al., 1991; Algeo, 1992; Raatz et al., 1994; Raatz, 1996; Raatz and Simo, 1998). We have restudied all of these sections, and the published work on them suffers from either inaccurate and/or insufficiently detailed lithologic description, a prevalence of disorganized facies stacking patterns, and/or a failure to demonstrate lateral continuity of the strata that supposedly represent glacioeustatic cycles (e.g., Lucas et al., 2012, 2016a, 2016b; Lucas and Krainer, 2015). Furthermore, most of these studies lack good age control, so assigning durations to the strata in question, which is done by identifying cycles supposedly driven by long (400 ka) or short (100 ka) eccentricity cycles, is highly conjectural.

The classic template for understanding late Paleozoic glacioeustatically driven sedimentation is in the Midcontinent, where cyclothems accumulated in a quiescent tectonic setting (e.g., Heckel, 2003, 2008) on broad, relatively flat platforms. The cyclothems are inferred to record glacioeustatic cycles that were orbitally forced by 100 and 400 kyr eccentricity cycles (Heckel et al., 2007). Soreghan (1992, 1994a,b) posited that the cycles she identified are glacioeustatic cycles that reflect the 400 ka long eccentricity cycle. In the Midcontinent, there are 7 Missourian and 10 Virgilian cycle groupings inferred by Heckel (2003, 2008) to correspond to long eccentricity cycles.

In the New Mexico sections, strata with good biostratigraphic age control generally contain fewer cycles than do correlative strata in the Midcontinent. Examples include well-studied Pennsylvanian sections in Sierra, Socorro, Valencia,

FIGURE 112. (*Opposite*) Conodonts from the lower part of the Gobbler Formation at the Space History A section. Upper views of P_1 elements. (A) *Rhachistognathus primus*? Space History A, bed 3. NMMNH specimen P-80123. (B–D) *Rhachistognathus muricatus*. Space History A, bed 3. NMMNH specimens (B) P-80124, (C) P-80125, (D) P-80126. (E) *Neognathodus atokaensis*? Space History A, bed 16. NMMNH specimen P-80127. (F–H) *Idiognathodus incurvus*? Space History A, bed 16. NMMNH specimens (F) P-80128, (G) P-80129, (H) P-80130. (I, J) *Idiognathodus* sp. Q? Space History A, bed 20. NMMNH specimens (I) P-80131, (J) P-80132. (K–M, Q, R, V, W) *Idiognathodus incurvus*? Space History A, bed 20. NMMNH specimens (K) P-80133, (L) P-80134, (M) P-80135, (Q) P-80136, (R) P-80137, (V) P-80138, (W) P-80139. (N, O, S) *Idiognathodus incurvus*? Space History A, bed 18. NMMNH specimens (N) P-80140, (O) P-80141, (S) P-80142. (P, T, U) *Idiognathodus* sp., wide morphotype. Space History A, bed 20. NMMNH specimens (P) P-80143, (T) P-80144, (U) P-80145. Scale bar = 0.25 mm.

and Bernalillo Counties, where cyclostratigraphy identifies between 25% and 50% of the required number of cycles (Lucas and Krainer, 2009; Lucas et al., 2013). Instead, the majority of the late Paleozoic section has a disorganized stratigraphic architecture and a lack of completeness (hiatuses) indicative of prolonged, repeated, and/or complex tectonic events. Thus, ARM tectonism must be considered to have exerted significant influence on Pennsylvanian sedimentation patterns and on the degree to which glacial–interglacial cyclicity can be detected and mapped in New Mexico (Lucas and Krainer, 2015).

Most of the analysis of sedimentation of the Pennsylvanian strata in the Sacramento Mountains has focused on identifying cycles supposedly forced by glacioeustasy. That said, it also must be considered that there is a strong relationship between local/regional climate and terrestrial sediment yield (Cecil and Dulong, 2003), documented empirically in studies of the sediment load in tropical rivers under different climatic regimes (Cecil et al., 2003a). This relationship is reflected in the linkages between climate and glacioeustasy (e.g., Cecil et al., 2003b; Horton et al., 2012; Rosenau et al., 2013). So even if glacioeustatic cycles are masked or difficult to detect because of the effects of tectonism, such cycles may, through coupled changes in climate, have influenced sedimentation patterns and thus affected the nature of the stratigraphic record. Opluštil et al. (2013, 2015), for a parallel example, have studied the relative effects of tectonism and climate on stratal successions in the Variscan Mountains of central Pangea (modern central Europe) and found clear signals of cyclic climatic change, even if partially obscured by tectonic events. Thus, although we find detailed, cyclothem-like glacioeustatic rhythms difficult to impossible to ferret out in the numerous late Paleozoic small basins that characterize central New Mexico, the presence of correlated effects needs to be considered when making interpretations of paleoecology.

For example, high elevation in an area will not, alone, lead to movement of high volumes of sediment from upland areas into coastal depocenters. During periods of seasonal, subhumid to semiarid climate regimes, sediment load will be maximized (e.g., Cecil et al., 2003a). This kind of climate appears to have prevailed, at times, in the Alamogordo study area during the Pennsylvanian, an interpretation based on the composition of the fossil plants, such paleosols as have been studied, and sedimentology. Such a subhumid climate was most likely the wettest end of the climate spectrum in the study region during the Pennsylvanian. There also appear to have been periods of significantly more arid climate. Limestone formation, for example, would have been significantly suppressed by the influx of large volumes of siliciclastics and therefore mainly would have taken place under a climate background of semiarid to arid conditions, when, even if there were sediment to move, there would have been insufficient moisture available to mobilize it. It is unrealistic to think tectonically driven uplift followed by erosion to baselevel accompanied periods of siliciclastic deposition in coastal regions and that limestone formation in shallow shelfal settings was restricted to periods of tectonic quiescence when there were no nearby “uplands.” Were uplift the only driver of erosion, it would require so-called yo-yo tectonics that were long ago called on to explain (and rejected as a viable explanation for) the cyclothems in Midwestern coal basins of the United States.

There is little question that the Alamogordo study area was a coastal region flanked inland by mountainous regions (see discussion below). There were uplands as part of the landscape, and they closely bordered the coastal regions. However, the alternation of marine carbonates and terrestrially derived siliciclastic deposits, rather than reflecting periodic pulses of tectonic activity, is more likely a reflection of the interplay of glacially driven climate cycles and tectonic creation of elevated areas (potential

FIGURE 113. (*Opposite*) Conodonts from the lower part of the Gobbler Formation at the Space History B and Indian Wells Canyon sections. Upper views of P_1 elements. (A–E) *Idiognathodus rectus*–*I. iowaensis* group. Space History B, bed 18. NMMNH specimens (A) P-80209, (B) P-80146, (C) P-80147, (D) P-80148, (E) P-80149. (F, G) *Idiognathodus attenuatus*. Space History B, bed 18. NMMNH specimens (F) P-80150, (G) P-80151. (H) *Neognathodus colombiensis*? Space History B, bed 9. NMMNH specimen P-80152. (I) *Neognathodus asymmetricus*. Space History B, bed 18. NMMNH specimen P-80153. (J, L) *Neognathodus bothrops* transitional to *N. asymmetricus*. Space History B, bed 3. NMMNH specimens (J) P-80154, (L) P-80155. (K) *Neognathodus bassleri* transitional to *N. bothrops*. Space History B, bed 3. NMMNH specimen P-80156. (M) *Neognathodus bothrops*. Space History B, bed 9. NMMNH specimen P-80157. (N–Q) *Idiognathodus amplificus* group. Space History B, bed 3. NMMNH specimens (N) P-80158, (O) P-80159, (P) P-80160, (Q) P-80161. (R, S) *Idiognathodus obliquus*. Space History B, bed 3. NMMNH specimens (R) P-80162, (S) P-80163. (T) *Idiognathodus obliquus*? Space History B, bed 9. NMMNH specimen P-80164. (U, AA, AB) *Idiognathodus incurvus*? Indian Wells Canyon, middle of bed 7. NMMNH specimens (U) P-80165, (AA) P-80166, (AB) P-80167. (V–X) *Idiognathodus* sp. G? nodose forms. Indian Wells Canyon, middle of bed 7. NMMNH specimens (V) P-80168, (W) P-80169, (X) P-80170. (Y, Z) *Idiognathodus incurvus*? Indian Wells Canyon, base of bed 9. NMMNH specimens (Y) P-80171, (Z) P-80172. Scale bar = 0.5 mm, except for in H–M, where second scale bar = 0.25 mm.

for erosion) and accommodation space (potential for preservation of sedimentary sequences). The contemporaneity of climate cycles, driven by South Polar glacial–interglacial cycles, leads to a regional system where tectonism masks and disrupts the regularity and areal extent of the cyclical events. But traces of these cycles remain in the overall sedimentary architecture, even if those traces are not found to be rhythmically distributed in time and space.

LOWLANDS AND UPLANDS

Analyses by Kottlowski (1960, 1968) and Meyer (1966) allow us an opportunity to discuss the location and approximate altitude of the Pedernal uplift during the Missourian, the time of most of the deposition of the Beeman Formation. Kottlowski (1968) estimated the average thickness of sediment stripped off of the Pedernal uplift during Missourian, Virgilian, and Wolfcampian time on the basis of a combination of the thickness of detrital sediments (shale and sandstone) in the two basins that adjoined the uplift (Orogrande to the west, Delaware to the east) and of an estimate of geographic size of the uplift (~25,000 km²). On the basis of the regional geology, the uplift appears to have been leveled and inactive during Leonardian time, at least in the area of the Sacramento Mountains. So assuming that the area of the uplift was essentially at (or very close to) sea level during Leonardian time, Kottlowski's (1968) estimates of sediment removal allow an approximation of the average height of the uplift during Pennsylvanian–earliest Permian time. Kottlowski's (1968) estimates are that sediment amounting to 107 m of elevation was stripped during the Missourian, 396 m during the Virgilian, and 213 m during the Wolfcampian, which indicates a maximum elevation of the Pedernal uplift at the onset of erosion of 716 m at the beginning of Missourian time. We emphasize that these are “static” estimates, assuming one pulse during which the entire Pedernal uplift was elevated, followed by erosion. Thus, all change in elevation of the uplift is assumed to be the result of sediment removal, ignoring the possibility of syntectonic movement of the uplift during erosion. Clearly, the

elevation estimate cannot be an exact number. However, these data do not indicate a very high elevation of the Pedernal uplift, say, not substantially higher than the approximately 716 meters estimated during Missourian time.

Furthermore, if we examine Meyer's (1966: fig. 36) isopach of Missourian strata in the Alamogordo area, those strata disappear (their zero isopach line) only about 15–20 km east of our Missourian fossil plant localities in Beeman Canyon (Figure 123). Given that the Holder Formation contact with the Beeman Formation is conformable and the Holder zero isopach line is very close to, or just east of, the Beeman zero isopach, it seems likely that the zero isopach of the Beeman is close to its true depositional pinchout. This relationship means the base of the Pedernal uplift was about 15–20 km east of our Beeman fossil plant sites.

Kottlowski (1968:33) also noted that during Missourian time “the Pedernal uplift took on the aspect of a tilted fault block with the western side apparently higher and providing more detritus directly to the west into the Orogrande Trough.” Finally, the influx of arkose in sandstone of the Beeman Formation indicates exposure of granitic bedrock in the Pedernal uplift source area of the Beeman rivers that deposited these sandstones. It also indicates the contemporaneous creation of sufficient accommodation space to allow relatively rapid burial of the sediment so that the feldspars did not weather to clays.

Climate and Vegetation

These elevational estimates, in combination with Beeman Formation lithostratigraphic patterns and siliciclastic sediment composition and depositional styles, provide an unusually rich source of data on which to base inferences about dynamics of the regional climate and its influence on the composition of paleofloras. As discussed above and as indicated by the studies of Cecil and colleagues in tropical Southeast Asia (Cecil, 2003; Cecil and Dulong, 2003; Cecil et al., 2003a), the transport of siliciclastic sediment is greatest when seasonal rainfall (semi-arid to subhumid climates) characterizes the prevailing climate.

FIGURE 114. (*Opposite*) Conodonts from the middle part of the Gobbler Formation at the Space History B section. Upper views of P₁ elements, except as noted. (A, B, D, F–I, L) *Idiognathodus acutus*. Space History B, bed 43A. NMMNH specimens (A) P-80173, (B) P-80174, (D) P-80175, (F) P-80176, (G) P-80177, (H) P-80178, (I) P-80179, (L) P-80180. (C, M) *Idiognathodus acutus*. Space History B, bed 46. NMMNH specimens (C) P-80181, (M) P-80182. (E) *Neognathodus expansus*. Space History B, top of bed 50. NMMNH specimen P-80183. (J) *Neognathodus dilatatus*. Space History B, bed 43A. NMMNH specimen P-80184. (K, N, P, Q) *Neognathodus roundyi*. Space History B, bed 43A. NMMNH specimens (K) P-80185, (N) P-80186, (P) P-80187, (Q) P-80188. (O) *Neognathodus expansus?* Space History B, bed 43A. NMMNH specimen P-80189. (R, S) *Gondolella wardlawi*. Space History B, bed 43A. NMMNH specimens (R) P-80190, (S) P-80191. (T) *Gondolella* sp., side view of P₂ element. Space History B, bed 41. NMMNH specimen P-80192. (U–Z) *Idiognathodus acutus*. Space History B, bed 41. NMMNH specimens (U) P-80193, (V) P-80194, (W) P-80195, (X) P-80196, (Y) P-80197, (Z) P-80198. Scale bar = 0.5 mm.

During such intervals, plant stabilization of soils through rooting is diminished, relative to periods of humid to perhumid climate (when rainfall exceeds evapotranspiration for 10 months or more of the year), permitting significantly more sediment discharge than during the more continuously wet intervals. Furthermore, during arid periods, waterborne sediment discharge also is reduced greatly, simply through the lack of adequate water for transport.

The rarity of organic-rich deposits in the Beeman Formation section, particularly coals and coaly shales, suggests that rainfall patterns in the region were distinctly seasonal, even during the wettest extremes. Thus, sediment transport may have been abundant during these periods, and the vegetation throughout the region may have been of a type tolerant of seasonal drought. The air circulation in the region was onshore from Panthalassa and may have been monsoonal (Tabor and Poulsen, 2008). This monsoonal flow would have brought about seasonal climates but also likely significant moisture during parts of the year, in the wetter months.

The proximity of the Pedernal uplift during Beeman Formation deposition provides an effectively constant source area for supply of siliciclastic sediment. Yet the Beeman Formation also contains limestones, including some with bedded chert. Limestones and high volumes of siliciclastic sediment transport are incompatible. Plus, the presence of bedded chert in the limestones suggests the possibility of dust plumes (Cecil, 2015; Cecil et al., 2018). One may conclude, therefore, that at times, the regional climate was arid. Again, this stratigraphic architecture is suggestive of fluctuation between periods of arid and subhumid climates, with limestones forming primarily during the more arid intervals and sediment transport concentrated in the more subhumid times.

A final consideration, important in interpreting the vegetation, is that in the study area, the elevational gradient during the Missourian extended from sea level to a maximum of about

750 m. This gradient in the study area was concentrated in a small area about 15–20 km long. Three conclusions follow. The area brings true upland vegetation into proximity with coastal vegetation, during the wetter climatic intervals in particular, a conclusion that seems inescapable. The geographic extent of the area under consideration probably is too small to have had much variation in overarching regional climate, although there may have been elevational effects on microclimate. Finally, the effects of the Pedernal uplift, if it caused microclimate variation, cannot be determined for certain; it is possible that under monsoonal circulation seasonal rainfall volumes were higher in the more elevated areas than at lower elevations because of temperature effects on the moisture-holding capacity of the atmosphere, although 750 m is not a particularly high elevation in a tropical region. If rainfall were greater at higher elevations, “wetland” vegetation could have been equally common in elevated and coastal areas but at different times as climatic conditions changed.

Putting this picture together with the Beeman floras, we propose that the dominant vegetation of the region was xeromorphic, probably coniferous, with local concentrations of species obligately tied to high soil moisture, such as some of the pteridosperms, the rare sigillarian lycopsids, and most of the marattialean ferns. Within this heterogeneous landscape, there most likely were a large number of spatially shifting microhabitats, given the variation in the inferred soil moisture tolerances of the regional species pool. This species pool is captured relatively well by the allochthonous nature of the plant deposits. This allochthony, however, also reduces our ability to infer the proportional composition of the standing vegetation directly from quantitative assessments of fossil-flora composition. Thus, our inferences are based on a combination of plant composition considered in the context of sedimentology, stratigraphic architecture, and inferences from those about regional climatic patterns, as outlined above.

FIGURE 115. (*Opposite*) Conodonts from the middle and upper parts of the Gobbler Formation at the Space History B section. Upper views of P_1 elements. (A) *Neognathodus expansus*? Space History B, base of bed 75. NMMNH specimen P-80199. (B) *Neognathodus expansus*? Space History B, base of bed 82. NMMNH specimen P-80200. (C, D) *Neognathodus roundyi*. Space History B, base of bed 75. NMMNH specimens (C) P-80201, (D) P-80202. (E) *Neognathodus roundyi*. Space History B, base of bed 82. NMMNH specimen P-80203. (F) *Neognathodus roundyi*. Space History B, bed 86. NMMNH specimen P-80204. (G) *Neognathodus roundyi*. Space History B, bed 84. NMMNH specimen P-80205. (H–J) *Idiognathodus expansus*. Space History B, bed 86. NMMNH specimens (H) P-80206, (I) P-80207, (J) P-80208. (O, P, S) *Idiognathodus delicatus*? Space History B, bed 86. NMMNH specimens (O) P-80210, (P) P-80211, (S) P-80212. (K) *Idiognathodus expansus*. Space History B, bed 84. NMMNH specimen P-80213. (L, T, U) *Idiognathodus delicatus*? Space History B, bed 75. NMMNH specimens (L) P-80214, (T) P-80215, (U) P-80216. (M, W) *Idiognathodus delicatus*? Space History B, base of bed 82. NMMNH specimens (M) P-80217, (W) P-80218. (N) *Idiognathodus expansus*? Space History B, bed 84. NMMNH specimen P-80219. (Q) *Idiognathodus* sp. Space History B, bed 86. NMMNH specimen P-80220. (R) *Idiognathodus* sp. Space History B, bed 84. NMMNH specimen P-80221. (V, X) *Idiognathodus expansus*? Space History B, base of bed 75. NMMNH specimens (V) P-57876, (X) P-80302. Scale bar = 0.5 mm.

GEOLOGICAL HISTORY

Pray (1959b) coined the term Orogrande Basin to refer to the late Paleozoic depocenter that formed west of the Pedernal uplift in southern New Mexico (Figure 122). Lawton and Giles (2003) explained the formation of the Orogrande Basin by strike-slip faulting resulting in polycyclic subsidence history, polymodal source areas, and intrabasinal uplift. The eastern edge of this depositional basin was the narrow Sacramento shelf, upon which the Pennsylvanian strata discussed here were deposited (Figure 123). This was an area of active tectonism during the Pennsylvanian–early Permian (e.g., Pray, 1961; Wilson, 1972b; Giles and Lawton, 1996; Lawton and Giles, 2003) that well reflects regional tectonic events of the ARM orogeny (Figure 123).

The first tectonic pulse was the onset of the ARM in New Mexico, which took place during the late Morrowan–Atokan, driving the onset and at least part or all of Gobbler Formation deposition (Figure 122). Its most obvious expression is the basal Pennsylvanian unconformity at which strata of Morrowan or Atokan age rest on rocks that range in age from late Proterozoic to Late Mississippian (Chesterian). In the Sacramento Mountains, this hiatus is the unconformity between the Morrowan base of the Gobbler Formation and underlying Mississippian strata (Figure 122). The Gobbler Formation represents the dominantly clastic fill of what has to have been a tectonically created feature of the onset of ARM tectonism, the Alamo clastic trough, a northwest–southeast-trending sedimentary depression in the Sacramento shelf. This trough essentially paralleled the front of the Pedernal uplift (Figure 122) and was either a synclinal fold or a fault-bounded depression or some combination of such structures. The Gobbler Formation fill, as noted above, shows little evidence of depositional cyclicity, or if present, it is difficult to decipher. This lack of cyclicity suggests

that the signal of climatic cyclicity, which characterized the glacial–interglacial world of the Lower and Middle Pennsylvanian, was swamped or disrupted, spatially and temporally, by local/regional tectonism. This tectonism does not mean that climate, linked to glacial cyclicity, played a small role in Gobbler Formation deposition. Consider empirical documentation of the close interconnection between climate and sediment load in rivers in the modern world (Cecil et al., 2003a; Latrubesse et al., 2005); it is under seasonally dry climates that the greatest volumes of sediment transport are most likely to occur (Cecil and Dulong, 2003). Thus, it is fully in keeping with the data from lithostratigraphy that the Gobbler Formation experienced climate cycles that varied between semiarid and subhumid, thus nearly always under climates that favored sediment transport, amplifying the effects of tectonism and promoting complex autocyclic depositional patterns. Indeed, the Middle Pennsylvanian in the Sacramento Mountains was a time of tectonic deformation with high-angle block faulting along the likely western shelf margin, resulting in a series of north–south-trending faults (Pray, 1961; Giles and Lawton, 1996; Lawton and Giles, 2003).

By late Desmoinesian time, deposition on the Sacramento shelf shifted to the platform-carbonate-dominated deposits of the Gray Mesa Formation. The Alamo clastic trough continued to receive sediment, but it likely was nearly filled to the southeast and east, where the Gray Mesa Formation limestone-dominated sections are relatively thick. In the area of our Space History sections, the trough was still accumulating clastic sediments until very late in Gray Mesa deposition, essentially the interval crossing the Desmoinesian–Missourian boundary. This boundary marks the onset of a major global environmental event during the Kasimovian/Missourian interval, possibly a large-scale deglaciation in Gondwana (Fielding et al., 2008; Eros et al., 2012;

FIGURE 116. (*Opposite*) Conodonts from the upper part of the Gobbler Formation at the Space History B, Space History D, and Indian Wells Canyon sections. Upper views of P_1 elements. (A) *Idiognathodus cancellosus*. Space History B, bed 155. NMMNH specimen P-80223. (B) *Idiognathodus turbatus*. Space History B, bed 160. NMMNH specimen P-80224. (C) *Idiognathodus heckeli*. Space History D, bed SGL-8. NMMNH specimen P-80225. (D, K) *Idiognathodus eccentricus*. Space History B, bed 160. NMMNH specimens (D) P-80226, (K) P-80227. (E) *Idiognathodus turbatus*. Space History B, bed 155. NMMNH specimen P-80228. (F, I) *Idiognathodus harkeyi*. Space History D, bed SGL-8. NMMNH specimens (F) P-80229, (I) P-80230. (G, H) *Idiognathodus sulciferus*. Space History B, bed 160. NMMNH specimens (G) P-80231, (H) P-80232. (J) *Idiognathodus sulciferus*. Space History D, bed SGL-8. NMMNH specimen P-80233. (L, O) *Idiognathodus swadei*. Space History B, bed 160. NMMNH specimens (L) P-80234, (O) P-80235. (M, N) *Neognathodus roundyi*. Indian Wells Canyon, base of bed 134. NMMNH specimens (M) P-80236, (N) P-80237. (P) *Idiognathodus turbatus*. Space History C, bed 1. NMMNH specimen P-80238. (Q) *Idiognathodus eccentricus*. Space History D, bed SGL-8. NMMNH specimen P-80239. (R, W, AC) *Swadelina neoshoensis*, short morphotype. Indian Wells Canyon, base of bed 134. NMMNH specimens (R) P-80247, (W) P-80248, (AC) P-80249. (S) *Swadelina?* sp. Indian Wells Canyon, base of bed 134. NMMNH specimen P-80240. (T–V, AA, AB) *Swadelina neoshoensis*. Indian Wells Canyon, base of bed 134. NMMNH specimens (T) P-80241, (U) P-80242, (V) P-80243, (AA) P-80244, (AB) P-80245. (X, Y, Z) *Idiognathodus* sp. Indian Wells Canyon, base of bed 92. NMMNH specimens (X) P-80250, (Y) P-80251, (Z) 80302. Scale bar = 0.5 mm.

Isbell et al., 2012), a huge event that would have affected sea level (Rygel et al., 2008; Heckel, 2008) and climate (Phillips and Peppers, 1984; Cecil, 1990) in the Pangean tropics, where the Sacramento shelf was located. In New Mexico, we refer to this as the “Amado event.”

In brief (a more extensive discussion will be published elsewhere), detailed correlations and facies analysis along an ~250 km transect from the Sacramento Mountains through the northern Sandia Mountains in southern Sandoval County identify the sedimentary signal of a substantial eustatic event across the Middle to Late Pennsylvanian (Desmoinesian-Missourian) boundary. This event is seen in the Sacramento Mountains in the Alamo clastic trough, a structure created by ARM tectonism, as deposition shifted from clastic-dominated to platform carbonates that cross the Desmoinesian-Missourian boundary. To the north, this event is correlated to the carbonate-dominated Desmoinesian/Missourian Amado Member (the Bug Scuffle Member of former authors, used here in association with the description of the conodonts) within the clastic-dominated lower part of the Atrasado Formation at numerous localities in Socorro, Valencia, Bernalillo, and southern Sandoval Counties. We thus term this the “Amado event,” a substantial glacioeustatic event essentially equivalent to the Swope cyclothem of the Midcontinent. Strata above and below the Amado event interval are mixed siliciclastic and carbonate sediments that generally lack organized facies stacking and show great lateral variations in thickness and facies. Their stratigraphic architecture suggests that an allocyclic driver such as eustasy was not the primary driver of deposition. The laterally extensive limestones of the Amado Member are composed of muddy microfacies (mostly wackestone to floatstone) and have a diverse fauna of brachiopods, crinoids, bryozoans, and other invertebrates. These limestones were deposited in a normal-marine, low-energy shelf environment. There are almost no high-energy sediments (grainstone, rudstone), which means that the Amado interval was deposited during a period of relative tectonic inactivity on a broad marine shelf. The Amado event marks the onset of a major global environmental

event during the Kasimovian/Missourian interval, possibly a large-scale deglaciation in Gondwana, a huge event that would have affected global sea level and climate in the Pangean tropics. This event correlates with major environmental and biotic changes, documented in the Midcontinent (Peppers, 1997), Illinois Basin (Phillips et al., 1974; Phillips and Peppers, 1984; DiMichele and Phillips, 1996b), and Appalachian Basin (Kosanke and Cecil, 1996). These biotic changes suggest a climatic shift toward more aridity at all phases of glacial-interglacial cycles (Phillips and Peppers, 1984; Cecil et al., 1985; Cecil, 1990), accompanied by a diminishment in the number of coal beds and their thickness, during the Kasimovian/Missourian interval.

Beeman deposition began in the early Missourian with what has long been acknowledged to be a tectonic pulse of the Pedernal uplift (Pray, 1961; Raatz, 1996; Raatz and Simo, 1998). The dominantly clastic interval of the lower Beeman Formation (Indian Wells Canyon Member) consists of nonmarine fluvial, deltaic, and marginal marine (estuarine) deposits that likely reflect relatively wet, subhumid climates in combination with differential tectonic uplift/subsidence along the western flank of part of the Pedernal uplift. This pulse is seen elsewhere across central New Mexico in Missourian clastic strata of the Atrasado Formation (Lucas et al., 2009, 2016b).

Subsequent deposition of the upper part of the Beeman Formation (Horse Ridge Member) was not obviously driven by glacioeustasy. These strata were deposited across the Missourian-Virgilian boundary as dominantly shallow marine shelfal deposits in which there is some evidence of parasequence stacking but no significant transgressive-regressive cycles. We suspect further research may document a dominantly local control of sedimentation, particularly by climate, some of which may have been ultimately driven by glacioeustasy.

Virgilian deposition of the Holder Formation began with the development of algal bioherms on a platform developed to the west of the tectonically active La Luz anticline (Wilson, 1972b). Subsequent Holder deposition (Space History Member) is apparently cyclic but, as argued above, a likely mixture of allocyclic

FIGURE 117. (*Opposite*) Conodonts from the middle part of the Beeman Formation at the Space History B and Space History C sections. Upper views of P_1 elements. (A, C, D) *Streptognathodus excelsus*. Space History B, bed 212. NMMNH specimens (A) P-80252, (C) P-80253, (D) P-80254. (B, H–J) *Idiognathodus magnificus*. Space History B, bed 229. NMMNH specimens (B) P-80255, (H) P-80256, (I) P-80257, (J) P-80258. (E, L) *Streptognathodus excelsus*. Space History B, bed 229. NMMNH specimens (E) P-80259, (L) P-80260. (F, V, W) *Streptognathodus excelsus*. Space History C, top of bed 44. NMMNH specimens (F) P-80261, (V) P-80262, (W) P-80263. (G) *Idiognathodus magnificus*. Space History B, bed 235. NMMNH specimen P-80264. (K, Q) *Streptognathodus excelsus*. Space History B, bed 235. NMMNH specimens (K) P-80265, (Q) P-80266. (M) *Idiognathodus magnificus*. Space History C, top of bed 44. NMMNH specimen P-80267. (N–P, R–U) *Idiognathodus symmetricus*. Space History C, top of bed 44. NMMNH specimens (N) P-80301, (O) P-80268, (P) P-80269, (R) P-80270, (S) P-80271, (T) P-80272, (U) P-80273. Scale bar = 0.5 mm.

(glacioeustatic and climatic) and autocyclic (local microclimate and tectonism) driven deposition.

A third pulse of the ARM orogeny is marked by the Bursum Formation and took place across the Virgilian-Wolfcampian boundary. This pulse is the Marathon orogeny of Ross (1986) and was the local/regional peak of ARM tectonic deformation (Dickinson and Lawton, 2003). This deformation during the late Virgilian to early Wolfcampian activated or reactivated major structural elements such as the Fresno fault, La Luz anticline, Dry Valley syncline, and related syndepositional folds, local erosional domes, and low-angle thrust faults along the Pedernal frontal escarpment. The Alamo trough also became an asymmetric intrashelf graben probably bounded by high-angle faults with both vertical and strike-slip movements. The Fresno fault was mainly active after deposition of the Holder Formation and before deposition of the middle Wolfcampian-Leonardian Abo Formation (Otté, 1959a), that is, during deposition of the Laborcita Member of the Bursum Formation. Some of the faults were

probably characterized by strike-slip movements (e.g., Alamo Fault, Bug Scuffle Fault).

Subsequent Wolfcampian-Leonardian deposition of the Abo Formation buried these structures (e.g., Pray, 1961). By Leonardian time, the Pedernal uplift was largely eroded away or buried, and deposition of the arid coastal plain and shallow marine Yeso Group commenced.

ACKNOWLEDGMENTS

We are grateful to the U.S. Bureau of Land Management, the U.S. Forest Service, and the U.S. Army (Fort Bliss) for access to land. Joe Cancellare, Amanda Cantrell, Larry Rinehart, and Tom Suazo provided valuable field assistance. Joerg Schneider generously lent his expertise on the insect wings we collected from the Beeman Formation. We thank the reviewers, whose comments improved the content and clarity of the manuscript.

FIGURE 118. (*Opposite*) Conodonts from the upper part of the Beeman Formation at the Dry Canyon and La Luz B sections and the lower part of the Holder Formation at the Long Ridge and Space History D sections. Upper views of P_1 elements. (A, C) *Streptognathodus pawhuskaensis*. Long Ridge, base of bed 221. NMMNH specimens (A) P-80274, (C) P-80275. (B) *Streptognathodus pawhuskaensis*. Space History D, top of bed 82. NMMNH specimen P-80276. (D) *Streptognathodus vitali*. Long Ridge, base of bed 221. NMMNH specimen P-80277. (E, G, N, O) *Streptognathodus pawhuskaensis*. La Luz B, base of bed 47. NMMNH specimens (E) P-80278, (G) P-80279, (N) P-80280, (O) P-80281. (F) *Streptognathodus virgolicus*. La Luz B, base of bed 47. NMMNH specimen P-80282. (H) *Streptognathodus pawhuskaensis*. Space History D, top of bed 44. NMMNH specimen P-80283. (I–K) *Streptognathodus pawhuskaensis*. La Luz B, top of bed 33. NMMNH specimens (I) P-80284, (J) P-80285, (K) P-80286. (L, M, T) *Streptognathodus bitteri?* La Luz B, top of bed 33. NMMNH specimens (L) P-80287, (M) P-80288, (T) P-80289. (P, U–X) *Streptognathodus pawhuskaensis*. La Luz B, top of bed 24. NMMNH specimens (P) P-80290, (U) P-80291, (V) P-80292, (W) P-80293, (X) P-80294. (Q, Y, Z) *Streptognathodus pawhuskaensis*. Dry Canyon 1. NMMNH specimens (Q) P-80295, (Y) P-80297, (Z) P-80298. (R, S) *Streptognathodus firmus*, late form. Dry Canyon 1. NMMNH specimens (R) P-80299, (S) P-80296.

FIGURE 119. Selected macrofossils of animals from the Beeman Formation. (A) Darwinulid ostracods, NMMNH specimen P-55958 from locality 12016. (B) Conchostracans, NMMNH specimen P-55978 from locality 9109. (C–E) Limulids, *Euproops danae*, NMMNH specimens (C) P-68701, (D) P-55976, and (E) P-68702, all from locality 9109. (F) *Septamyalina*, NMMNH specimen P-55960 from locality 12016. (G) Microconchids, USNM specimen 706337 from USNM locality 43891. Note that NMMNH locality 9109 = USNM locality 43834, and NMMNH locality 12016 = USNM locality 43838. Scale bars = 1 cm in A–F, 1 mm in G.

FIGURE 120. Selected macrofossils and coprolite of animals from the Beeman Formation. (A–C) Cockroach wings. (A) NMMNH specimen P-55966 from locality 12016. (B) NMMNH specimen P-55964 from locality 12016. (C) NMMNH specimen P-55965 from locality 12016. (D) *Palaeonodonta*, NMMNH specimen P-77346 from locality 9109. (E) Palaeoniscoid fish scales, NMMNH specimen P-77195 from locality 9109. (F) Shark coprolite, *Crassocoprus*, NMMNH specimen P-77351 from locality 9109. Note that NMMNH locality 9109 = USNM locality 43834, and NMMNH locality 12016 = USNM locality 43838. Scale bars = 1 cm.

age	Sandia Mountains	Los Pinos Mountains	Cerros de Amado	Oscura Mountains	Little San Pascual Mtns	Fra Cristobal-Caballo Mtns	San Andres Mountains	Sacramento Mountains
Wc	Bursum Fm	Bursum Fm	Bursum Fm	Bursum Fm	Bursum Fm	Bursum Fm	Panther Seep Formation	Bursum Fm
Virgilian	Atrasado Formation	Atrasado Formation	Atrasado Formation	Atrasado Formation	Bar B Formation	Bar B Formation		Holder Formation
Missourian								
Desmoinesian	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	
Atokan	Sandia Formation	Sandia Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation	Gray Mesa Formation
		Sandia Formation	Gray Mesa Formation		Gray Mesa Formation	Gray Mesa Formation		
Mo					fault			Gobbler Formation

FIGURE 121. Correlation of Pennsylvanian strata from north central New Mexico (Sandia Mountains) to southern New Mexico (Sacramento Mountains). Fm = formation; Mo = Morrowan; Wc = Wolfcampian.

FIGURE 122. Tectonic map of the Orogrande basin (after Algeo, 1992). The black dots are the principal control points of Algeo et al.'s (1991) study.

age		lithostratigraphic units		tectonic pulses	ice ages
Permian	Wolfcampian	Bursum Formation			
Late Pennsylvanian	Virgilian	Holder Formation	Mill Ridge Member	← Marathon orogeny	
			Little Dry Canyon Member	ARM folding	
		Beeman Formation	Indian Wells Canyon Member		
	Missourian		Horse Ridge Member		
	Space History Member	Gray Mesa Formation	north-south block faulting		
Middle Pennsylvanian	Desmoinesian			Gobbler Formation	
	Atokan				
Early Pennsylvanian	Morrowan			← begin ARM orogeny	

FIGURE 123. The Pennsylvanian section in the northern Sacramento Mountains, age, lithostratigraphic units, inferred ARM tectonism, and correlation to Gondwana ice ages (ice ages from the compilation of Rygel et al., 2008).

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